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GEOCHEMICAL AND MINERALOGICAL STUDIES OF THE GRANITIC GNEISS AND ASSOCIATED ROCKS OF WESTERN ARDGOUR, ARGYLL.

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UNIV. EDINBURGH 1819

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I. INTRODUCTION

1. LOCATION, TOPOGRAPHY AND OBJECTS OF STUDY

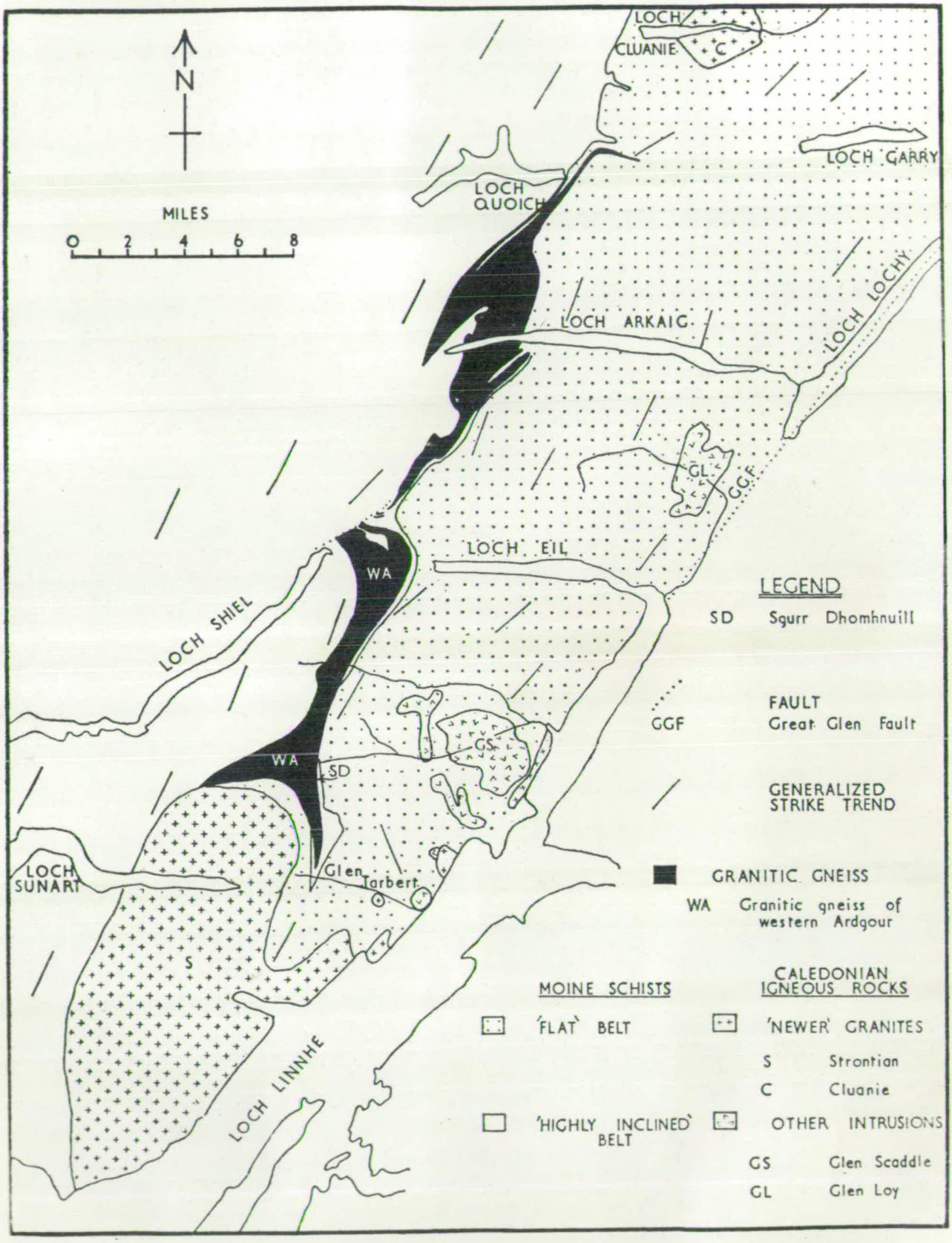
The rocks which are the subject of this thesis come from an area situated in the north-west Highlands of Scotland, and represented on the Ordnance Survey 1-inch sheets 35 and 46 (7th series) and the Geological Survey (G.S.) 1-inch sheets 52, 53 and 62. This area comprises the westernmost part of the parish of Ardgour, Argyll, together with a small portion of the parish of Lochaber, Inverness-shire, and is a narrow belt of country 16 miles (26 km.) long and 1 to 3 miles (1.5 to 5 km.) wide running from near the watershed in Glen Tarbert to the southern slopes of Beinn an Tuim (see Map 1).

The topography is mountainous and shows evidence of extensive glaciation. The highest point in the area is the summit of Sgurr Dhomhnuill (2915 ft., 888 m.) and most of the hills exceed 2000 ft. (650 m.) in height. The valleys and ridges tend to run east-west and the valley floors are usually less than 750 ft. (230 m.) above sea level. Apart from a few plantations of young trees, the vegetation consists everywhere of rough grass or heather. With the exception of the valley floors and some of the slopes between Meall nan Damh and the Callop River, the area is very well exposed.

The geological setting of the area is shown in Fig. 1 (taken/

FIGURE 1.

The regional setting of the Ardgour granitic gneiss, after Leedal (1952), with modifications by Dalziel (1963 b).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WA
SD
GS
LOCH SUNART
Glen Tarbert
S
LOCH LINNHE
LEGEND
SD Sgurr Dhomhnuill
FAULT
GGF Great Glen Fault
GENERALIZED STRIKE TREND
GRANITIC GNEISS
WA Granitic gneiss of western Ardgour
MOINE SCHISTS
'FLAT' BELT
'HIGHLY INCLINED' BELT
CALEDONIAN IGNEOUS ROCKS
'NEWER' GRANITES
S Strontian
C Cluanie
OTHER INTRUSIONS
GS Glen Scaddle
GL Glen Loy

FIG. 1.

(taken from Leedal 1952). The most important rock type is granitic gneiss, a body of which has an outcrop extending from Glen Tarbert in the south to Glenfinnan in the north and 1 to 3 km. wide. This body, whose origin is problematical, is surrounded, apparently conformably, by a sequence of psammitic and pelitic schists and gneisses belonging to the Moine Series and representing a sedimentary sequence believed to be of pre-Cambrian age. These schists and gneisses have reached the sillimanite grade of metamorphism and together with the granitic gneiss contain many veins, lenses and bands of pegmatite and aplite. In addition to these rocks, which are the objects of the present study, intrusive rocks which post-date the folding and metamorphism of the pelitic and psammitic rocks and the formation of the granitic gneiss outcrop in the area. These are the Strontian "granite" and a number of felsite, lamprophyre, and trondhjemite-porphyry dykes, but they were not studied.

Previous workers in the area were concerned mainly with the petrography and structural relations of the rocks, though their geochemistry was touched upon by Mercy (1963). Various theories have been advanced for the origin of the granitic gneiss, but none of these is wholly convincing. The second part of this chapter is devoted to a discussion of previous work, and Chapter Two consists of a brief summary of the petrography of the rocks.

In the present study, rocks and minerals from the area were analysed/

analysed chemically in order to find the most likely mechanism by which the granitic gneiss could have formed. The sampling and analytical techniques used are described in Chapter Three.

The problem of the origin of the granitic gneiss was studied from three different angles and the results of these investigations are presented in Chapters Four, Five and Six.

Firstly, numerous samples of the granitic gneiss were collected from fairly evenly spaced localities all over its outcrop and chemically analysed for all major and some trace elements. The pattern of variation in composition of the granitic gneiss from place to place was compared with the variation pattern in granitic bodies generally accepted to be of magmatic origin; whence it might prove possible definitely to accept or reject a magmatic origin for the granitic gneiss. It was hoped that this method might help to decide whether the granitic gneiss could have formed by the partial melting of the surrounding country rocks. The composition of the granitic gneiss and its variation from place to place are discussed in Chapter Four.

Secondly, samples of the surrounding psammitic and pelitic schists were analysed to see whether it was likely that the granitic gneiss could have formed from them. The composition of the metasediments around the granitic gneiss were compared with previous analyses of Moine metasediments to see whether it was likely that the metasediments around the granitic gneiss had/

had suffered any metasomatism. This material is presented in Chapter Five.

Thirdly, the felspars and micas were separated from some of the analysed rocks and the separated minerals were analysed completely or in part, and Chapter Six is devoted to consideration of these analyses and their implications. The compositions of the co-existing felspar pairs were used in an attempt to estimate the temperatures at which the rocks crystallised or recrystallised and hence to assess the likelihood of partial melting being involved in the origin of the granitic gneiss. The compositions of the biotites were used to clarify the rôle of Al in the rocks of the area and to see if there was any systematic difference in their compositions between the granitic gneiss and the metasediments or within the granitic gneiss which might have a bearing on the origin of the latter.

In Chapter Seven the origin of the granitic gneiss is discussed further in the light of the material presented in Chapters Four, Five and Six and compared with the theories put forward by previous workers for the origin of other migmatitic rocks.

2. HISTORY OF PREVIOUS WORK.

The earliest work in the area was done by Wilson of the Geological Survey (Wilson 1901, pp 120-2, Peach and Wilson 1903 p. 68, and Bailey and Maufe 1916 pp. 84-5 and 88), who mapped the/

the parts of Western Ardgour which lie in the G.S. one-inch sheets 53 and 62, though his map of sheet 62 was never published. On the one-inch G.S. sheet 53, the augen-gneiss facies of the granitic gneiss is labelled "augen gneiss (Sgurr Dhomhnuill)" and marked as a metamorphosed intrusion, while the rest of the granitic gneiss is included with the oligoclase-biotite-quartz gneiss (hereafter referred to as "oligoclase gneiss") as "pelitic gneiss". The augen gneiss was likened (Peach and Wilson 1903 p. 68) to that of Inchbae, but evidently great difficulty was found in drawing its boundaries, as a considerably larger amount of augen gneiss was mapped by Wilson than by Harry (1953). The summit of Sgurr Dhomhnuill was mapped by Wilson as augen gneiss but found by Harry to be oligoclase gneiss with thin psammitic bands. The work of the Geological Survey on the part of the area lying in sheet 53 has been summarised and compared with that of Harry by Bailey (1960, pp. 116-8).

The parts of the area lying in one-inch G.S. sheets 52 and 62 were mapped by the Geological Survey in 1931 (Richey 1931 and MacGregor and Kennedy 1931 p. 106) and 1951-7 (MacGregor 1951, 1952, 1953 and 1957) respectively. The psammites of Glen Garvan were found (MacGregor 1951) to young away from the granitic gneiss. No maps or memoirs for these sheets have yet been published, but seven analyses of rocks from the area have been published by the Geological Survey (Dixon 1934, no.

975/

975; Harvey and Wilson 1954, nos. 1650, 1651 and 1652; and Harvey and Wilson 1955 nos. 1664, 1678 and 1679). The rocks analysed were an augen gneiss and two other granitic gneisses, a pelitic gneiss, a pegmatite vein cutting "striped locally - injected pelitic Moine schists", a trondhjemite-porphyry and a lamproschist, the last two belonging to later (Caledonian) intrusions. The most notable feature of these analyses is the richness of the pegmatite vein 1650 in Na. In this it resembles the aplitic body in the granitic gneiss analysed by Dalziel (1963a, Appendix II, No. 7).

The next important work in Western Ardgour was done by Harry (1953), who mapped the part of the area lying south of Cona Glen, except for the extreme western tongue west of National Grid line 1/84. The results of this mapping are included in Map 1. His paper contains no detailed structural work, being chiefly concerned with the petrology and petrogenesis of the rocks, but the granitic gneiss was found to outcrop in the core of a large antiform. The granitic gneiss has no sharp contacts with the surrounding rocks, but whereas the granitic gneiss-psammite contacts are gradational over a few inches only, the granitic gneiss-oligoclase gneiss contacts are gradational over distances varying from two feet to several yards, while on the south-west slopes of Sgurr a'Chaoruinn there is a zone 300 yards wide consisting of alternating ill-defined layers of granitic gneiss and oligoclase gneiss. The relation-

ship of the granitic gneiss to the various types of aplite and pegmatite bodies found in it were described on hand-specimen and microscopic scales.

The petrogenetic scheme put forward by Harry was based largely on six new analyses, three each of granitic gneiss and oligoclase gneiss, and was strongly influenced by the ideas of Reynolds (1946). By comparison of his analyses of oligoclase gneiss with three analyses of pelitic Moine schists from Moidart and Ross-shire using a von Wolff diagram, Harry concluded that the oligoclase gneisses formed by Ca, Na, Mg metasomatism of Moine pelitic schists. Metasomatism took place by movement of intergranular fluids but the quartz-oligoclase folia which give the rock its "injected" appearance arose by metamorphic differentiation in situ and not by forcible injection of quartzofelspathic material. The granitic gneiss formed from the oligoclase gneiss by Si, K metasomatism, the microcline having developed at the expense of the oligoclase and the quartz being of late replacement origin. The augen gneiss was similar in origin to the rest of the granitic gneiss, except that potash felspar porphyroblasts were able to grow undisturbed for a long period because stress conditions were such as to inhibit crystallisation of quartz till later. All the pegmatite and aplite veins in the granitic gneiss formed by metasomatic concentration of quartz and felspar at the expense of biotite, followed, in a few cases/

cases, by slight mobilisation. The metasomatism was considered in terms of advancing fronts, the front of granitisation responsible for the granitic gneiss pushing ahead of it, the front of Na-Ca-Mg metasomatism responsible for the formation of the oligoclase gneiss. The metasomatism took place during the folding of the area.

More recent workers in the area (Mercy 1963, Dalziel 1963a) strongly criticised Harry's ideas on the petrogenesis of the rocks, but his descriptive work still holds, though his interpretation of the perthitic and other textures in the granitic gneiss was believed by Dalziel (1963a, pp. 163-4 and 199-200) to be wrong.

Four further analyses of the granitic gneiss and ten of pelitic and psammitic gneisses from Ardgour were provided by Mercy (1963), and the geochemistry of the Ardgour rocks was compared briefly with that of Caledonian granites and Dalradian metasediments. The granitic gneiss of Western Ardgour differed from the Caledonian granites in having a $Mg/(Fe^{2+} + Fe^{3+})$ ratio of 0.3 instead of 1.0 and in having a Na/K ratio of about 1 instead of 2 or 3. The pelitic and psammitic gneisses were found to have a $Mg/(Fe^{2+} + Fe^{3+})$ ratio of 0.5 and a Na/K ratio varying from 0.75 to 2, and were thus the poorest in Mg and richest in Na of all the metasediments considered. The potassic nature of the Ardgour granitic gneiss was explained by assuming that it was formed by partial melting early in the metamorphic history of the area and at relatively low temperatures of a series of metasediments which need not have been rich in K. Three

granitic gneiss analyses which formed a traverse near the western margin of the granitic gneiss outcrop on Druim Garbh showed a gradual decrease in Fe^{2+}, Fe^{3+}, and Mg, decrease in Na/K and increase in Fe^{3+}/Fe^{2+} on going towards the centre of the granitic gn eiss outcrop, while the $Mg/(Fe^{2+} + Fe^{3+})$ ratio remained constant. Crystallisation differentiation was thought to have taken place, and the migmatitic nature of the granitic gneiss was questioned. The relatively sodic nature of the metasediments was considered consistent with their being the unmetasomatised derivatives of arkoses.

In his paper, Mercy took care to state that his ideas on the origin of the granitic gneiss were highly speculative. The structural work of Dalziel (1963a) together with his work on zircon shape and size distributions (Dalziel 1963b) are difficult to reconcile with a magmatic origin for the Ardgour granitic gneiss, and it is quite probable that the variations noted by Mercy in his Druim Garbh traverse were random and without significance, especially as they could be accounted for by variation in the amounts of the various minerals in the rocks without any variation in mineral composition across the traverse.

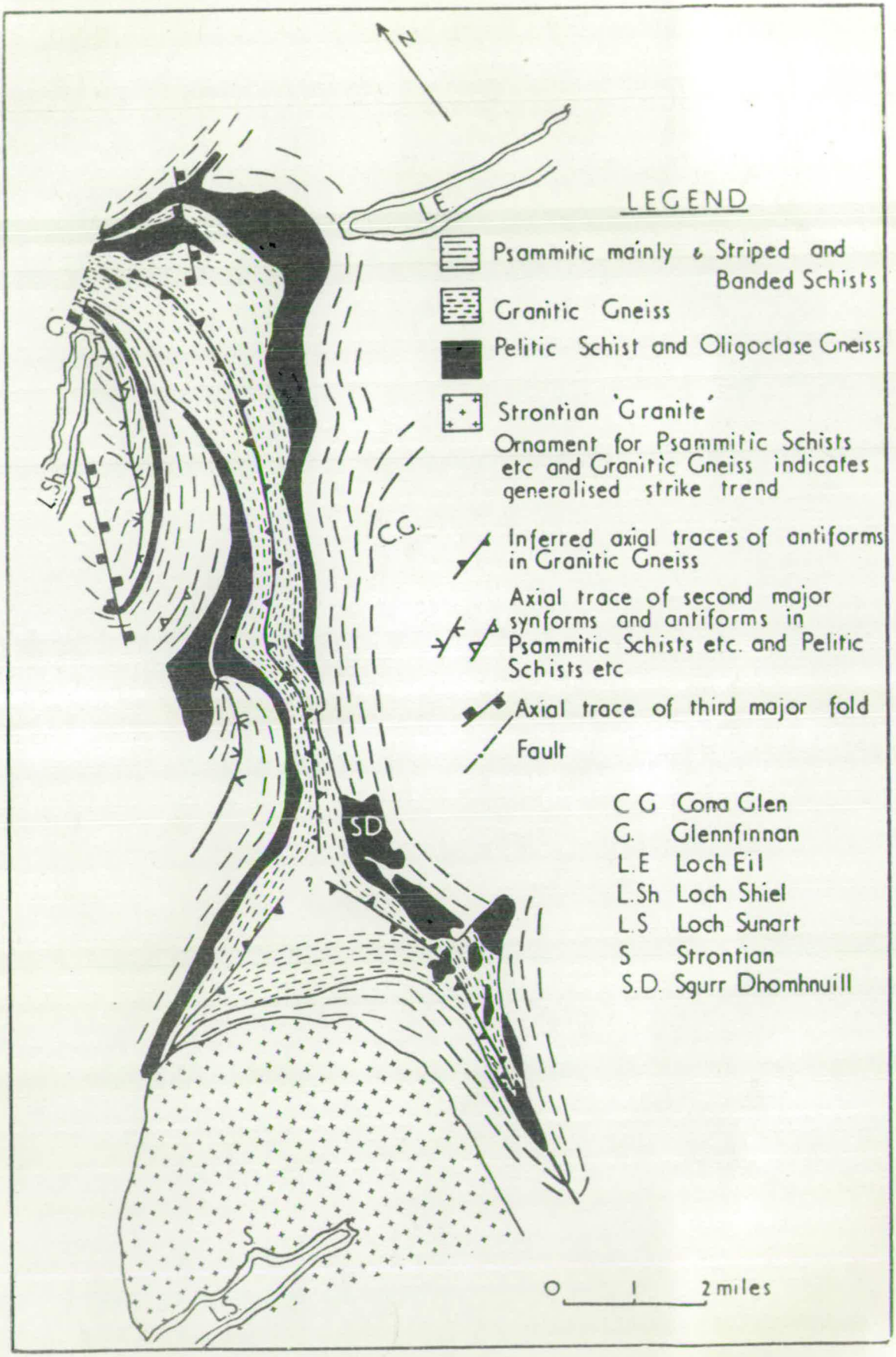
Extensive work on the part of the area from Meall Mor northwards was carried out by Dalziel (1963a). This consisted chiefly of detailed structural mapping, the results of which are included in Map 1, but some petrological work and a brief discussion/

discussion of the geochemistry of the granitic gneiss were also included. The rocks of the area have been affected by four fold episodes, but the major folds of the area belong to the second and third episodes only. The Beinn an Tuim striped group and the granitic gneiss outcrop as if they were stratigraphically equivalent; both lie immediately below the Druim na Saille pelitic group (oligoclase gneiss of Harry). The Beinn an Tuim striped group and the Druim na Saille pelitic group are together stratigraphically equivalent to the Striped and Pelitic group of Moidart and Morar (Richey and Kennedy 1939 and Howkins 1961), while the Glen Garvan psammitic group is equivalent to the Upper Psammitic Group of Moidart and Morar. The major fold running NE-SW through the granitic gneiss and striped group outcrop belongs to the second fold episode and is an almost vertically plunging south-closing fold which is antiformal in some places and synformal in others. The axis of the fold generally runs NE-SW but near Meall Mor and Beinn an Tuim has been deflected by major folds of the third period of movement. The structural mapping of Dalziel and similar work by Johnson (who mapped the area around Glen Hurich and Druim Glas and parts of whose unpublished structural map were used in drawing Map 1) showed the general structural pattern for the area to be that given in Figure 2 (taken from Dalziel and Johnson 1963, Fig. 1.)

In addition to detailed structural mapping, Dalziel also used/

FIGURE 2.

The major structural features of the Ardgour district, from Dalziel and Johnson (1963, p. 245).


N
LE
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C.G.
SD
?
S
L.S
LEGEND
Psammitic mainly & Striped and Banded Schists
Granitic Gneiss
Pelitic Schist and Oligoclase Gneiss
Strontian 'Granite'
Ornament for Psammitic Schists etc and Granitic Gneiss indicates generalised strike trend
Inferred axial traces of antiforms in Granitic Gneiss
Axial trace of second major synforms and antiforms in Psammitic Schists etc. and Pelitic Schists etc
Axial trace of third major fold
Fault
C G Coma Glen
G. Glennfinnan
L.E Loch Eil
L.Sh Loch Shiel
L.S Loch Sunart
S Strontian
S.D. Sgurr Dhomhnuill
0 1 2 miles

FIG. 2.

used the textures of the rocks in an attempt to correlate the structural and metamorphic histories of the area. Similar work on the south-western part of the area was done by Johnson and the results as far as they affect the granitic gneiss are given in Dalziel and Johnson (1963). The granitic gneiss "formed after the F1 folds on the limbs of one of the nappes which was subsequently folded into major F2 folds". The concordant pegmatites formed before or early in the F2 episode and are deformed by F2 structures. The discordant pegmatites are probably of F3 or later age (Dalziel 1963a, Fig. 24). Both the regional migmatisation which formed the oligoclase gneisses and the formation of the granitic gneiss occurred before or early in the F2 episode, but no structural evidence was found to indicate the relationship between the two events.

Detailed petrographic descriptions of the rocks in the area were given by Dalziel (1963a, Appendix I), who concluded (1963a, pp. 163-4 and 199-200) that the evidence adduced by Harry for the formation of the granitic gneiss by metasomatism of the oligoclase gneiss was wrongly interpreted, and that the perthites in the granitic gneiss are mainly of exsolution origin, the examples found of microcline replacing oligoclase and myrmekitic oligoclase replacing microcline representing local recrystallisation and/or melting conditions.

Nine new analyses from the area were presented (Dalziel 1963a Appendix II). The granitic gneiss was much closer in composition/

composition to the psammitic than to the pelitic gneisses, and formed migmatitically from rocks of the Beinn an Tuim striped group. There was no indication that the oligoclase gneisses are chemically different from non-migmatised pelitic schists.

The zircons from the granitic gneiss and the surrounding metasediments were also examined (Dalziel 1963b) and compared with those from the magmatic Strontian "granite" and the pre-metamorphic Inchbae augen-gneiss. The zircons of the granitic gneiss and metasediments have length-frequency distributions significantly different from those of the Strontian "granite" and the Inchbae augen gneiss. Though the zircons of the granitic gneiss are distinguished from those of the metasediments by the presence of outgrowths and intergrowths, they are significantly different in shape from the euhedral and sharply terminated zircons characteristic of the Strontian "granite" and the Inchbae augen gneiss. These findings made it more likely that the granitic gneiss never passed through a magmatic phase.

The hypothesis advanced by Dalziel concerning the origin of the granitic gneiss is the most convincing so far put forward, but some problems remain unanswered. In particular, it was not proved whether the formation of the granitic gneiss from the rocks of the striped group was isochemical or not, though K was thought to have been added during the formation of/

of the granitic gneiss, possibly through the uprise of K-rich volatiles produced by a partial melt below the present level of exposure. No attempt was made to explain the supposed crystallisation differentiation suggested by Mercy on the basis of his Druim Garbh samples. Lastly, and probably most important, difficulty was found in explaining why the granitic gneiss should have formed only from the eastern part of the striped group north of Meall Mor when the western part remained unaffected.

In a recently published paper, Butler, B.(1965a) compared the chemical compositions of previously analysed Moine "pelitic" rocks with new analyses of "uninjected" Moine mica-schists from Ardnamurchan. Certain Na-rich garnetiferous mica-schists from Ardnamurchan which come from parts of the major "pelitic" groups of the Moine series were considered to be the metamorphic equivalents of sediments of greywacké type, and it was thought likely that some of the Moine "pelites" described as injected might be rocks of greywacké composition which had acquired an "injected" appearance as the result of small-scale differential mobilisation, rather than originally pelitic rocks into which new material had been introduced during the regional metamorphism. In so far as they affect the area of Western Ardgour, Butler's conclusions, though slightly at variance with the ideas of workers in the area as to the nature of the sediments from which the oligoclase gneisses were derived, render more plausible the assertions of Mercy (1963) and Dalziel (1963a, p. 194) that the/

the oligoclase gneisses are not the products of metasomatism but reflect the original chemical composition of the sediments from which they are derived.

II. PETROGRAPHY

The petrography of the rocks of Western Ardgour was described in detail by Harry (1953) and Dalziel (1963a, p.22 and Appendix I). For this reason no detailed petrographic work was done, but a thin section of each analysed rock was cut in order to check its mineralogy. In addition an attempt was made to compare the chemical analyses of 49 of the granitic gneisses with modes obtained by point counting. However, the discrepancies between modes obtained from different thin sections cut from the same hand specimen were so great that no purpose would have been served by point counting thin sections of the other analysed rocks.

A brief summary of the findings of Harry and Dalziel is given below, followed by a short statement of the mineralogy of the new analysed samples and an assessment of the results of the modal analysis done in the present study.

1. DESCRIPTIVE PETROGRAPHIC SUMMARY

a. Granitic gneiss (Harry pp. 292-300, Dalziel pp xx-xxx)

This was divided into a normal facies and an augen-gneiss facies on the absence or presence of potash felspar augen up to 2 cm. in size in the rock. Otherwise the two facies are identical.

The granitic gneiss is usually medium to coarse-grained with an average grain size of 1.5 - 2 mm. The mica plates are always in nearly parallel alignment and the rock is usually

banded. In the most typical granitic gneisses the biotite is concentrated into folia 0.5 - 2 cm. apart, but there are all gradations from almost complete lack of compositional banding to a rock consisting entirely of concordant pegmatite veins 1 to 8 cm. thick separated by biotitic selvedges 0.5 - 1.5 cm. thick. Compositional banding seems to be least well marked in the finest (grain size 1 mm.) and coarsest (grain size 4 mm.) granitic gneisses.

The minerals of which the granitic gneiss is comprised are described below.

(i) Quartz: About 30% of the rock consists of quartz which occurs as irregular shaped crystals whose cuspate boundaries ("caries texture") are generally taken to indicate that the quartz has replaced the felspars to some extent. Quartz also occurs as a component of myrmekitic intergrowths with plagioclase.

(ii) Potash felspar: Potash felspar usually forms about 20-25% of the rock, though in some samples the amount may drop to 5% or even less. It is microcline, but the typical microcline "grating" is usually only poorly and patchily developed. The microcline is usually perthitic but the scale and amount of exsolution vary quite considerably from sample to sample.

(iii) Plagioclase: The plagioclase of the granitic gneiss is an oligoclase An 22-28, and the exsolved plagioclase in the perthite has the same composition. It usually forms about 30% of the rock, but when the rock is poor in potash felspar/

felspar it is proportionately rich in plagioclase. The plagioclase is usually antiperthitic and where it abuts against microcline it becomes slightly more sodic and often develops myrmekitic intergrowths with quartz.

(iv) Biotite: The granitic gneiss may contain from 2 to 20% biotite, though 12% is the most usual amount. It is strongly pleochroic and the darkest colour is usually a deep umber brown, though in some granitic gneisses a more reddish brown tint occurs.

(v) Muscovite: Though it only occurs sporadically in the granitic gneisses and usually forms about 2% of the rock when present, muscovite occasionally forms 10% of the rock.

(vi) Garnet: A few of the granitic gneisses contain rounded sieve-like pale coloured garnets, but in amount they always form less than 1% of the rock.

(vii) Accessories: Apatite, zircon and ilmenite are almost ubiquitous. One or two samples contain an occasional crystal of sphene or hornblende.

The potash felspar in the augen gneisses is concentrated into augen whose greatest dimension is 1 - 2 cm., but the amount of potash felspar in the rock remains about the same. Some granitic gneisses of the normal facies have suffered intense rodding and at first glance simulate the texture of the augen gneisses.

The/

The granitic gneiss in parts of the northern half of its outcrop has suffered late-stage hydrothermal alteration which has chloritised the biotite and reddened the felspars of these rocks.

b. Pegmatites and Aplites (Harry pp. 294-5, Dalziel pp. xxx-xxxiii).

The pegmatites found in the area were formed at various stages in its structural history. Those which pre-date the second fold movement are commonest in the granitic gneiss and seem to be connected with its origin. They are concordant with respect to the early foliation but a few have been tectonically isolated by the second fold movement. Those in the granitic gneiss vary in thickness from that of the banding in the groundmass of the granitic gneiss (1 cm.) up to 15 or even 20 cm. Their composition is that of the granitic gneiss except that they are much poorer in dark minerals. The quartz and felspars show the same micrscopic relations to each other, and perthitic and myrmekitic textures are common. The pegmatites have biotite-rich selvedges which often contain sphene, epidote and allanite, minerals which are exceedingly rare in the ordinary granitic gneiss. Pegmatites of the same age which lack potash felspar are rare, and are commonest in, though not confined to, pelitic schists and oligoclase gneisses.

Several later generations of pegmatite occur of which some are earlier than, but most later than, the third fold episode. They/

They are discordant with respect to the foliation of the surrounding rocks, but, in general, they have not disrupted foliation or cleavage planes in the surrounding rocks. They are much coarser grained than the early pegmatites, and the minerals in them are zoned. They are usually of granitic composition.

Aplites are commonest in the granitic gneiss, and form either cross-cutting veins up to 1 m. thick or unfoliated roughly equidimensional bodies 50 cm. long. In thin section, some show alignment of mineral grains, and they consist of plagioclase (An 22-28), biotite and garnet phenocrysts set in a matrix of quartz, potash felspar, plagioclase and biotite. The groundmass grain size is 0.1 - 0.3 mm., and the phenocrysts may reach 3 mm. in size.

c. Psammitic schists and gneisses (Harry pp. 287-8, Dalziel pp. xii-xv).

These rocks form by far the greatest part of the Glen Garvan (Upper) Psammitic Group. The remainder of this group consists of thin interbedded layers of semipelitic and calc-silicate schists.

The psammitic schists and gneisses are flaggy. Parallel alignment of mica flakes gives them a good schistosity which becomes less marked south of Glen Gour where the psammites are coarser grained and more contorted. The psammites usually have a grain size of 0.5 - 0.8 mm. and contain the following minerals/

minerals:

(i) Quartz: This is the most important minerals in the rock and occurs in amounts from 30 to 90%. The margins of the crystals are usually smooth, but myremekitic intergrowths with plagioclase sometimes occur.

(ii) Felspars: Most psammites contain both plagioclase and potash felspar, but the relative proportions vary considerably, and either felspar may be the more abundant. The plagioclase of the psammites may vary from An 22 to 38.

(iii) Micas: Biotite occurs in all the psammites in amounts varying from 1 to 10%. Muscovite is sometimes absent but may form up to 10% of the rock. The micas tend to be concentrated in flaggy partings and give the rock its schistosity. The darkest colour exhibited by the biotite is usually a reddish brown.

(iv) Accessories: Apatite, zircon and ore minerals occur, but are sometimes very sparsely distributed in the rocks.

d. Striped and semipelitic schists (Dalziel pp i-iii, xii, and xv-xviii).

Striped schists form the major part of the Beinn an Tuim striped group and occur to a small extent in the Druim na Saille pelitic group. Semipelitic schists form a minor part of the Glen Garvan psammitic group. The striped schists consist of alternating psammitic and pelitic bands in varying proportions and range from flaggy psammites with micaceous partings/

partings to pelitic schists with psammitic ribs. The semi-pelitic schists are more massive and the micas are more evenly distributed in the quartzofelspathic groundmass. The striped and semipelitic schists have an average grain size of 0.4 - 0.7 mm., and the minerals they contain are briefly described below:

(i) Quartz: The rocks may contain from 20 to 60% quartz, which forms equidimensional crystals whose boundaries are irregular in the striped schists but more rounded in the semipelitic schists.

(ii) Felspars: The plagioclase has a composition An 22-32 and is always present. Potash felspar usually occurs in the semipelitic schists but is often absent from the striped schists. Only where both felspars occur are myremekite, perthite and antiperthite developed.

(iii) Micas: Biotite and muscovite are usually both present, but muscovite is absent from some striped schists. The darkest colour exhibited by the biotite is usually a reddish brown.

(iv) Garnet: A few crystals of garnet frequently occur and are often surrounded by mica spindles.

(v) Sillimanite: Certain semipelitic schists contain pods up to 1 cm. long consisting of quartz, biotite, muscovite, and sillimanite.

(vi) Accessories: Apatite, zircon, and ore minerals occur/

occur in small amounts.

e. Pelitic schists and oligoclase gneisses (Harry pp. 288-290, Dalziel pp. iii-iv and vii-xii).

These rocks form the bulk of the Druim na Saille pelitic group and also occur near the top of the Beinn an Tuim striped group. They are medium to coarse grained. The pelitic schists are medium grained and fairly closely foliated with the quartz and felspar evenly distributed throughout the rock. They grade into coarse oligoclase gneisses whose texture consists typically of bands 0.5 - 1 cm. thick, having a preponderance alternately of quartzofelspathic and micaceous material. The minerals forming these rocks are listed below:

(i) Quartz: The proportion of quartz in these rocks is usually 10-25%, but may vary considerably. The grain size varies with that of the rock from 0.6 to 3 mm., and the crystals are usually very irregular in outline.

(ii) Felspars: Potash felspar is absent from nearly all of the pelitic schists and oligoclase gneisses, and when it is absent myrmekite is also absent. The plagioclase is oligoclase or andesine, An 25-32, is often untwinned, and forms 25-50% of the rock.

(iii) Micas: The pelitic schists and oligoclase gneisses are very micaceous, containing up to 30% of any one mica or 45% of total mica. The amount of biotite always exceeds 15%, but muscovite may be absent, or present in accessory/

accessory quantities. The two micas are usually intergrown and individual flakes may reach 5 mm. in length in the oligoclase gneisses.

(iv) Garnet: Most of these rocks contain some garnet, but the amount rarely exceeds 5% and is usually nearer 2% of the rock. The garnets are usually rounded and often contain small inclusions of quartz.

(v) Sillimanite and staurolite: Only accessory amounts of sillimanite and staurolite are found in the pelitic schists and oligoclase gneisses and then only at one or two localities. The sillimanite forms fine needles which appear to grow on biotite, while the staurolite forms separate crystals up to 2 mm. long.

(vi) Accessories: Apatite, zircon, and ore minerals are commonly present, and together form about 1% of the rock.

f. Calc-silicate schists (Harry, p. 288, Dalziel, pp. iv-viii, xii and xviii-xx).

Thin ribs up to 30 cm. thick of calc-silicate schist occur at many stratigraphic levels, though they are relatively infrequent in the Druim na Saille pelitic group. A number of calc-silicate bands often occur at the boundary between the Druim na Saille pelitic group and the Glen Garvan psammitic group. Individual calc-silicate bands often show zonation of minerals within the bands.

The/

The calc-silicate rocks usually contain abundant quartz. The other minerals present occur in variable quantities, and are labradorite (An 58-68), hornblende, a diopsidic pyroxene, garnet, clinozoisite, sphene, and occasionally biotite. The hornblende is always pleochroic in shades of green.

g. Hornblende schists (Harry pp. 290-291, Dalziel pp. xxxiv-xxxvii).

The hornblende schists form bands, lenses, and blocks which may reach 1 m. in thickness, and have a foliation which has been folded by the second fold movements. They represent pre-metamorphic intrusions and occur both in the metasediments and in the granitic gneiss. They have a foliation produced by alignment of hornblende crystals and are often coarse-grained, the hornblende crystals reaching 3.5 mm. in length. They are occasionally cut by thin pegmatitic veins which end abruptly against the surrounding rock when this is granitic gneiss.

The chief constituent mineral of these rocks is a deep green hornblende which often forms 70% of the rock. Plagioclase is always present, forming 10-20% of the rock, but its composition is variable. In hornblende schists surrounded by granitic gneiss it is An 26-38, but in hornblende schists surrounded by other metasediments it is An 28-63. Quartz, garnet, and biotite are usually present in amounts less than 10%. The usual accessories are sphene, epidote, and ore minerals.

The/

The granitic gneiss adjacent to the hornblende schists usually contains more epidote, sphene and ore minerals and less potash felspar, than normal granitic gneiss, and may contain a little hornblende as well.

2. PETROGRAPHY OF ANALYSED ROCKS

The petrography of the various analysed samples of the same rock type is so similar that the features which differ from rock to rock can be conveniently tabulated.

a. Granitic gneisses: All of these rocks contain quartz, oligoclase An 22-28, potash felspar, biotite, apatite, zircon and ore minerals. Table 1 lists the darkest colour shown by the biotite, the degree of development of perthitic texture, and the presence or absence of garnet, muscovite, and hornblende.

b. Striped and psammitic schists: These are listed in Table 2, which gives the colour of the biotite, the presence or absence of muscovite and garnet, and the nature of the dominant felspar.

c. Pelitic schists and oligoclase gneisses: These are listed in Table 3, which gives the colour of the biotite, the presence or absence of muscovite, garnet, and potash felspar, and the approximate composition of the plagioclase.

TABLE 1.

Petrography of analysed granitic gneisses.

Sample No.	Colour of Biotite.	Presence of Garnet.	Presence of Muscovite.	Presence of Hornblende.	Degree of Perthitic Texture.
G11	B				1
G12	B				1
G13	B				3‡
G21	Bl				2
G22	B				1
G23	B				2
G31	R				2
G32	B				3
G33	B				1
G41	B				3
G42	B				1
G43	B				1
G51	B	+			1
G52	B	+			2
G53	B				2
G54	B				3
G55	B				2
G56	B		+		1
G61	B			+	2
G62	B				1
G63	B				1
G64	Bl				2
G65	Bl				1
G71	Bl				1
G72	Bl		+		1
G73	B				3
G74	B				1
G81	B				2
G82	B	+			1
G83	R	+			1
G91	R				2
G92	B	+			1
G93	B				2
G96	B			+	2
G97	R	+			2
G98	B				1

Continued overleaf.

TABLE 1. (Contd.)

Sample No.	Colour of Biotite.	Presence of Garnet.	Presence of Muscovite.	Presence of Hornblende.	Degree of Perthitic Texture.
G101	B				1
G102	B		+		2
G103	B*				3
G105	B				2
G111	B				1
G112	R				1
GP112	B				3
G113	B				1
G116	B				2
G117	R				1
G121	B				2
G122	B				2
G123	B				2
G131	B			+	1
G132	B				1
G133	B				1
G141	B				2
G142	B				2
G151	B				1
G152	B				1
G161	B				1
G162	B				2
G163	B				1
G171	B		+		3
G172	R				3
G173	B				3
G181	B1				2
G182	B				1
G183	B	+			2
G184	B				2
G185	B				3
G186	R				3
G187	B				2
G191	B				3
G192	B				2
G193	B				2

Continued overleaf.

TABLE 1. (Contd.)

Sample No.	Colour of Biotite.	Presence of Garnet.	Presence of Muscovite.	Presence of Hornblende.	Degree of Perthitic Texture.
G201	B		+		1
G202	B				2
G203	B				3
G211	B				1
G212	B				1
G213	B				1
G221	R				3‡
G222A	R		+		1
G223	B				1
G231	R				2
G232	B*				3
G233	R				2
G241	B	+	+		2
G242	B*	+			2
G243	B*				3
G244	R				3
G251	B		+		2
G252	R*				3
G253	B	+			2
G254	B				2
G261	R	+	+		1
G262	R*	+			1
G263	B	+			2
G271	R	+	+		3
G271A	R	+	+		1
G272	B		+		1
G273	B1		+		2
G274	B				3‡
G281	B	+			1
G282	R	+			1
G283	B	+			2‡
G284	B		+		3
G291	B				2
G292	R		+		3
G293	B*				2

For explanation of symbols see overleaf.

TABLE 1. (Contd.)

Biotite colours:- B = Pleochroic from yellow to dark umber brown.
Bl = Pleochroic from yellow to black or greenish black.
R = Pleochroic from flesh colour to deep brick red.

Muscovite listed as present only when it is definitely not an alteration product
+ = present; space blank = absent.
* Biotite almost all chloritised.

Perthitic textures:- 1 = perthite lamellae in microcline and antiperthite lamellae in plegioclase quite coarse (0.02 - 0.05 mm.)

2 = perthitic textures recognisable from distinct mottled extinction, Lamellae visible under higher powers.

3 = Mottled extinction poorly developed or absent. Lamellae barely visible or invisible.

‡ These rocks are very rich (over 50%) in plagioclase and poor in microcline (less than 15%).

TABLE 2.

Petrography of analysed psammites.

Sample No.	Colour of Biotite	Presence of Muscovite.	Presence of Garnet.	Dominant Felspar.
Q187	R	+		Plagioclase
K191	B	+		Plagioclase
SK211	R	+		K felspar
SK221	R	+		Plagioclase
SK222	R			Plagioclase
SK241	B	+		K felspar
SK242	R	+		Plagioclase
K271	R	+		Plagioclase
K281	R			Plagioclase
SK291	B			Plagioclase
S42	B	+		K felspar
S52	R			K felspar
S71	B			plagioclase
S77	R	+		K felspar
S96	R	+		K felspar
S101	B			K felspar
S116	R	+		Plagioclase
S121	R			Plagioclase
S141	R	+	+	Plagioclase
S142	R	+	+	K felspar
S151	R	+		Plagioclase
S162	* *	+	+	K felspar
S191	R			Equal.
S213	R	+		K felspar
S252	B			Plagioclase
S261	R			K felspar

Sample Q187 is from a thin psammitic band within the Druim na Saille pelitic group. Samples bearing the letter K are striped schists from the Beinn an Tium striped group, and those bearing the letters SK are psammitic schists from that group. Samples bearing the letter S only are psammites from the Glen Garvan psammitic group.

* * = No fresh biotite in slide.

Other symbols as in Table 1.

TABLE 3.

Petrography of analysed pelites.

Sample No.	Colour of Biotite.	Presence of Muscovite.	Presence of Garnet.	Presence of K felspar.	Composition of Plagioclase.
P42	B	+			An 24
P53	R	+			An 30
P81	R	+			An 28
P96	R	+			An 30
P102	R		+		An 30
P125	R		+		An 32
P132	B			+	An 32
P142	R	+			An 26
P161	R				An 28
P166	R	+	+		An 30
P181	R	+			An 30
P192	R	+		+	An 24
P223	R			+	An 26
P233	B	+			An 26
P273	B	+	+		An 30
P291	B1	+			An 32

Figures for An content of plagioclase only approximate.

Explanation of symbols as for Table 1.

d. Other analysed rocks are listed and briefly described below:

G222 and G262A: Concordant pegmatites about 8-10 cm. thick together with their biotitic selvedges (up to 8 mm. thick). The pegmatite veins themselves are very coarse grained (grain size about 4 mm.) and are almost free of dark minerals (though G222 contains muscovite). The selvedges are very rich in biotite (forming 40-50% of the selvedge), which reaches 4 mm. in length, and contain 2 - 5% of garnet, but are poor in potash felspar.

G274A: An aplite from a vein cutting granitic gneiss. This is a medium-grained rock (about 0.5 mm.), rather poorer in dark minerals than most granitic gneisses, but containing abundant muscovite. The micas are randomly arranged and do not give a foliation. Neither the plagioclase nor the potash felspar, which are present in roughly equal amounts, is strongly perthitic, but myrmekitic quartz-plagioclase intergrwoths are common.

GS 166: Though probably derived from a psammite, this rock is unusually coarse grained (average grain-size about 3 mm.). In hand specimen it looks like an unfoliated very leucocratic granite. It is poor in biotite and contains equal proportions of quartz, potash felspar, and plagioclase, plus 5 - 10% of muscovite/

muscovite. The biotite is reddish brown in colour.

L116: Calc-silicate or calcareous psammite. This rock is fine grained (about 0.3 mm.) and consists of about 50.% quartz and 45% plagioclase An 45-50, together with 2 - 3% each of biotite and garnet. The quartz and plagioclase crystals have an almost granualr texture and the garnets, which reach 1 mm., are usually rounded and have few inclusions. The biotites, which are strongly chloritised, are in roughly parallel alignment.

S172: Alternate bands 1 - 2 cm. thick of impure psammite and calc-silicate make up this rock. The psammitic ribs have an average grain size of 0.4 mm. and consist of about 20% of dark brown biotite in parallel alignment, and about 40% each of quartz and plagioclase An 50 which are moderately rounded and flattened parallel to the foliation. The calc-silicate ribs are finer grained and consist of hornblende crystals about 0.5 mm. in size aligned with their long axes in the foliation plane, set in a matrix consisting of fairly rounded grains of quartz and plagioclase An 70 with an average grain size of 0.1 mm.

3. MODAL ANALYSIS.

An attempt was made to check the chemical compositions of the analysed rocks against their modal compositions as determined by/

by point counting. However, due to the unsatisfactory results obtained for the first group of rocks, only 49 modes, all of granitic gneisses, were determined.

Two thin sections of each rock were cut perpendicular to each other and to the foliation of the rock. Before covering, the felspars were stained with sodium cobaltinitrite and barium rhodizonate by the method of Bailey and Stevens (1960). Using a Swift point counter which would only allow measurement areas up to 2.3 x 3.0 cm., 2000 -2100 points were counted on each thin section. The average modes are listed in Table 4.

The sources of error in modal analysis are numerous and have been fully discussed by Chayes (1956). Working on two-felspar granites, he found that the amount of precision obtainable in modal analysis of a homogeneous rock was largely dependent on measurement area and grain size, provided that at least 1500 points were counted on each slide. In the present case, the measurement area could not be increased above about 600 sq. mm., and grain size became the most important factor. Unfortunately, the measure of grain size used by Chayes (1956, pp. 72-75, 89), that is I.C. number, is one liable to be affected by the banding and inhomogeneity present in the Ardgour granitic gneiss but absent from Chayes' homogeneous and unfoliated granites. However, the grain size of the Ardgour granitic gneiss is believed to be similar to that of the average granite or rather less, and in estimating the precision of/

TABLE 4.

Modes of 49 analysed granitic gneisses (average of two counts).

Sample No.	G41	G42	G43	G51	G52	G53	G54	G55	G56	G61	G62	G63	G64	G65
Quartz	31	26	24	32	28	36	21	26	34	30	24	28	29	26
Potash felspar	25	32	34	28	27	25	28	36	20	28	38	36	33	35
Plagioclase	30	31	34	32	32	30	38	33	30	28	27	28	25	29
Biotite	14	11	7.7	9.2	13	7.9	12	3.8	14	12	9.7	9.1	13	9.7
Muscovite	-	-	-	-	-	-	-	-	0.1	-	-	-	-	-
Opaque	0.1	0.1	0.2	tr	tr	1.0	tr	1.3	1.1	1.6	0.2	0.2	0.2	0.1
N.O.A.*	0.7	0.2	0.8	0.3	0.6	0.7	0.3	0.6	0.7	1.4	tr	0.2	0.1	0.3

Sample No.	G71	G72	G73	G74	G81	G82	G83	G91	G92	G93	G96	G97	G98	G101
Quartz	29	32	33	30	21	29	34	38	31	31	38	33	30	34
Potash Felspar	26	25	30	32	34	31	20	32	25	30	33	27	19	42
Plagioclase	32	29	21	24	30	23	30	26	32	30	23	32	39	20
Biotite	13	14	15	15	13	18	15	5.2	13	11	5.6	9.0	13	5.0
Muscovite	0.1	0.5	-	-	-	tr	tr	-	-	-	-	-	-	-
Opaque	-	-	0.2	-	0.2	tr	-	tr	-	-	0.1	-	-	-
N.O.A.*	0.3	0.3	0.5	0.3	0.5	0.4	0.8	0.2	0.5	0.2	0.4	0.5	0.5	-

Sample No.	G102	G103	G105	G111	G112	G113	G116	G117	G121	G122	G123	G131	G132	G133
Quartz	37	27	26	42	34	31	32	30	26	32	38	34	33	33
Potash Felspar	28	37	30	31	33	26	29	27	40	33	23	40	39	16
Plagioclase	28	35	34	23	29	29	28	32	26	32	33	21	25	35
Biotite	7.9	1.6	9.0	4.0	3.4	13	9.1	9.0	7.2	2.9	6.9	4.3	2.6	16
Muscovite	0.7	-	-	-	-	0.1	-	-	-	-	-	-	-	-
Opaque	0.4	-	0.9	0.4	0.9	0.9	1.7	1.3	0.5	0.7	0.2	0.1	0.6	0.2
N.O.A.*	0.1	-	0.5	0.1	0.2	0.7	0.9	0.5	0.3	0.5	0.1	0.6	0.3	0.3

Sample No.	G141	G142	G151	G152	G161	G162	G163
Quartz	38	32	27	29	31	25	34
Potash Felspar	36	24	33	12	33	42	21
Plagioclase	19	31	31	47	30	19	33
Biotite	4.9	13	7.8	9.5	4.6	13	12
Muscovite	-	-	-	-	-	-	-
Opaque	1.4	0.5	0.7	1.1	0.5	1.1	0.3
N.O.A.*	0.5	0.5	0.2	0.5	0.1	0.6	0.2

* N.O.A. = Non-opaque accessories (hornblende, garnet, zircon and apatite).

tr = trace (less than 0.05%).

of modal analysis the granitic gneiss will be compared with granites of I.C. number 40. The standard error predicted by Chayes (1956, p. 76) for counting one thin section, of area 625 sq. mm., for a mineral forming about 30% of the rock should be $\pm$ 2.90 (% mineral). The error for a mineral forming 10% of the rock would be (0.1 x 0.9)/(0.3 x 0.7) of this, that is $\pm$ 1.25, assuming that this error has a binomial distribution.

The average differences between modes of the same rock are shown in Table 5, which shows that the inaccuracy of the modal analyses is only slightly greater than that predicted by Chayes (1956, p. 76) for a rock of I.C. number 40. Nevertheless, the two modes obtained from the same hand specimen often showed considerable discrepancies and often had little relation to the norm of the same rock.

The modal analyses were also compared with the chemical analyses by recalculating the norms on the basis of minerals as close in composition as possible to the analysed minerals separated from the granitic gneiss and surrounding rocks (see Table XIV for fuller explanation of the calculation and Table XV for a list of the modified norms). The effect of analytical error on these recalculated norms was calculated (Table XVI). This error and the error in the modes estimated from the difference between duplicate modes of the same rock were compared with the differences found between the modal and recalculated normative compositions of the same rock. These results/

TABLE 5.

Precision of modal data, calculated from duplicate modes of the same rock.

Mineral.	Greatest difference between 2 modes of same rock.	Average difference between 2 modes of same rock.	Estimated standard deviation for modal analysis.	Predicted error of one mode (Chayes).	Predicted erro of average of two modes (Chayes).
Quartz	21.2	4.27	±3.02	±2.90	±2.04
Potash felspar	34.0	5.11	±3.61	±2.90	±2.04
Plagioclase	18.8	3.94	±2.79	±2.90	±2.04
Biotite	11.1	2.39	±1.69	±1.24	±0.87
Total Dark Minerals*	11.3	2.55	±1.80	±1.24	±0.87

* including muscovite.

TABLE 6.

Comparison of modal and recalculated normative data for 49 granitic gneisses.

Mineral	Average difference (mode-modified norm)	Std.deviation of differences (mode-norm)	Error (as std.deviation) for modal analysis.	Error (as std.deviation) of modified norms due to experimental error.	Predicted std.deviation of difference (mode-norm)	Greatest + and - values of (mode-norm).
Quartz	-1.97	±3.68	±3.02	±1.14	±3.23	+10.4 -12.0
Potash felspar	-0.28	±5.93	±3.61	±1.95	±4.10	+13.2 -14.8
Plagioclase	+2.44	±4.49	±2.79	±1.54	±3.19	+13.2 -7.8
Biotite	+3.18	±3.16	±1.69	±2.10	±2.70	+13.0 -2.8
Total Dark Minerals*	+0.13	±2.54	±1.80	±1.12	±2.12	+7.5 -4.4

* including muscovite.

results are shown in Table 6. The differences between mode and recalculated norm are 20-50% greater than predicted for each mineral, but it is doubtful whether this has any significance.

The difference between the average modal and recalculated normative compositions of the granitic gneiss are due to three causes:

(a) Possible inaccuracies in the analytical methods used; for example, if Al was systematically low, this would give systematically high values for hornblende or low values for muscovite in the modified norm.

(b) Variation in the compositions of the minerals forming the rocks.

(c) The mode is expressed in volume percent, while the modified norm is expressed in molecular percentages. However, the discrepancy due to this cause is quite slight, and Barth (1955, Table 5) has shown that for the minerals used in the present recalculated norms, the one-cation percentage norm is much closer to a volume percentage norm than a weight percentage norm would be.

Estimation of the relative amounts of the various minerals in a rock from a chemical analysis can only be accomplished if the compositions of the minerals in the rock are known. In the present case, the similarity from rock to rock in the compositions/

compositions of the analysed biotites and felspars was thought sufficient for constant mineral compositions (apart from variation in the Fe^{2+}/Mg ratio in biotite and amphibole and An content in plagioclase) to be used in recalculating the norms. The recalculated norms were thought to be at least as good as the modes in estimating the mineralogical compositions of the Ardgour rocks, except that the presence of small quantities of hornblende or muscovite in the recalculated norms of granitic gneisses was realised as usually being due to experimental error in analysis, and did not necessarily indicate the presence of these minerals in the rocks themselves.

Thus the discrepancies between modal and recalculated normative compositions were considered to reach such large proportions as to render modal analysis useless in the study of the compositional variation from place to place within the granitic gneiss, even though modal analysis of a large number of thin sections would be a reasonable way of estimating its average mineralogical composition.

III. SAMPLING AND ANALYTICAL TECHNIQUES

1. SAMPLING

The amount of exposed rock in Western Ardgour is large, and in about 70% of the area an outcrop could be chosen to correspond to a predetermined point on the 6-inch topographic map of the area. It was therefore possible to choose a sampling plan which adequately represented the target population (that is, the total number of samples into which the whole mass of granitic gneiss could be divided). An attempt was made to keep systematic bias in collection to a minimum, and to see that the sample localities were as evenly spread as possible over the granitic gneiss outcrop and left no major part of the outcrop unsampled.

A sufficient number of samples was collected to make it likely that computer treatment of the analyses would yield statistically significant trend surfaces. A considerable amount of random variation due to the inhomogeneity of the rock was anticipated. Also, very little trend surface treatment of chemical analyses has been done. For these reasons, a sampling density in excess of the 4 per square mile (= 1.44 per sq. km.) used by Whitten (1961b) was considered necessary. A sampling density of 2 per sq. km. necessitating the collection of 80 samples was felt to be the minimum number acceptable while a maximum of 120 was set by the length of time available for the chemical/

chemical analysis of the samples.

The idea of collecting samples on a grid was rejected because on the scale proposed (the centres and corners of 1 km. National Grid squares) several important parts of the granitic gneiss would have remained unsampled and more closely gridded sampling would have yielded an impracticably large number of samples. Complete random sampling within the granitic gneiss outcrop would probably have been practicable but was rejected because of the difficulty of deciding how many of the randomly selected points would lie at outcrops of granitic gneiss.

The sampling plan adopted was designed to secure material from the interior and from near the edges of the granitic gneiss body in roughly equal proportions. A point near the centre of the granitic gneiss outcrop (on the summit ridge of Beinn Mheadhoin at the point NM/890700) was chosen, and a line was drawn through it perpendicular to the strike of the granitic gneiss. Similar lines, always keeping perpendicular to the strike, were drawn at half-mile intervals as measured along the major fold axis running through the granitic gneiss outcrop (Figure 2). Depending on the length of these traverses, 2, 3, 4 or 5 granitic gneiss samples were collected along each traverse (Map 1). The samples of pelitic and psammitic rocks were collected along extensions of these traverse lines.

Point samples were collected at each locality, one or two large/

large pieces being taken and trimmed of as much weathered material as possible on the spot. The weights of the granitic gneiss samples ranged from 1 to 2 kg., while the pelite and psammite samples weighed 0.75 to 1.5 kg. Aplites and cross-cutting pegmatites were avoided or removed, but concordant pegmatites less than 5 - 8 cm. in thickness were included in the sample. In two cases, samples of concordant pegmatite (together with its biotitic selvedge) and ordinary granitic gneiss were collected from the same locality.

The size of sample chosen was determined by consideration of the errors involved in sampling a rock where the mineral grains are not randomly arranged, and comparison with the errors involved when, during preparation for analysis, a 200-250 g. grab sample of pieces about 9 c.c. in size is taken from the original sample.

Consider a rock containing by volume 30% quartz, 30% potash felspar, 30% plagioclase, and 10% biotite, with each of the following textures in turn:

(a) Equigranular with grain size 2 x 2 x 2 mm., and all the minerals randomly distributed about the rock.

(b) All the biotite aggregated into randomly distributed folia 50 x 50 x 1 mm. in size.

(c) All the potash felspar aggregated into randomly distributed augen 20 x 20 x 10 mm. in size. The sampling errors can be estimated by calculating the number of crystals, folia/

folia or augen of each mineral in the original sample and the grab sample and assuming that the crystals, folia or augen are randomly distributed in the mass of rock from which the sample comes. Then the binomial theorem can be applied to the crystals, folia or augen and the probable error in the amount of any mineral in the sample can be calculated (compare Moroney 1956 pp. 123-130); where the number of crystals, folia or augen in the sample is large, a normal distribution is approximated. If the whole sample weighs 1500 g. (volume 550 c.c.) and the grab sample consists of 9 pieces each weighing 25 g. (average volume 9 cc.), then the sampling errors, expressed as the standard deviation of the distribution of the percentage of the mineral actually found in the sample, are as shown in Table 7.

Though the size of sample collected is quite adequate for homogeneous rocks, it is too small for inhomogeneous rocks, even though the degree of inhomogeneity occurring in actual rocks never reaches the extreme levels exemplified by textures (b) and (c) above. However, in the case of such inhomogeneous rocks, the error involved in collecting a sample 1.5 Kg. in size is no greater than that incurred during the process of preparation for analysis.

2. PREPARATION FOR ANALYSIS

The samples were scrubbed clean with a stiff brush under-

TABLE 7.

Theoretical errors in sampling and preparation for analysis of rocks with certain hypothetical textures (see p.34).

Mineral	Postulated Texture	Error in original sample.			Error in c.225g. grab sample.			Error in duplicate analyses of same rock.	
		as volume % mineral	as weight % K_2O	as weight % FeO	as volume % mineral	as weight % K_2O	as weight % FeO	Wt.% K_2O.	Wt.% FeO.
Quartz	a	0.0019			0.015				
Plagioclase	a	0.0019			0.015				
K felspar	a	0.0019	0.00029		0.015	0.0021		0.05	0.02
Biotite	a	0.0004	0.00004	0.00009	0.003	0.0003	0.0007	0.05	0.02
Biotite	b	0.64	0.06	0.14	0.65*	0.06	0.14	0.05	0.02
K felspar	c	2.1	0.31		2.8*	0.41		0.05	0.02

* Corrected for the splitting of folia and augen likely to take place in dividing sample into 2x2x2cm. cubes.

All errors in this table are given as standard deviations of populations whose distributions can be assumed to be normal. The figures in columns 9 and 10 are taken from column 7 of Table 9.

running water, rinsed with deionised water, and left to dry overnight. They were then split into pieces up to 3 x 2 x 2 cm. with a hydraulic rock splitter. Fine particles, and pieces containing weathered material and aplite or pegmatite veins were discarded. The rest of the sample was spread out on a piece of glazed paper, mixed thoroughly by hand, and a grab sample of 200-250 g. (7-10 pieces) taken. The remainder of the material was retained in a large paper envelope. A piece 5 x 4 x 3 cm. in size was sectioned.

The 200-250 g. grab sample was pounded down to pass a steel 12 mesh sieve using a tungsten carbide pestle and a plate of the same material. The pestle was brought down vertically and no attempt was made to grind the sample at this stage. Care was taken to see that no chips were lost and so little dust was produced as possible. When all the material from this process had passed the sieve, a sixth of it (35 - 45 g.) was taken for grinding to 100 mesh by means of an automatic sample divider, and the remainder was retained in a paper packet.

The 35 - 45 g. from the sample divider was ground by hand on an agate mortar to pass a 100 mesh bolting silk sieve. Care was taken to sieve the material often, to lose no material from the mortar while grinding, and to produce as little dust as possible. When it had all passed the sieve, the powder was put into a 100 g. glass bottle, shaken well, and dried overnight at 110°C.

In/

In the case of the rocks selected for mineral separation, the 12 mesh material was ground to 100 mesh in an automatic agate mortar. The 100 mesh powder was washed with water in a beaker and the cloudy water containing rock flour decanted. Decantation was continued until the supernatant liquid became clear on standing for 1 - 1½ minutes. The powder was then washed twice with acetone and dried at 110°C. Enough material was ground to provide 100 - 120 g. of washed powder.

Samples for analysis must be ground carefully to keep errors to the minimum. Some cnntamination by steel from the hydraulic rock splitter, which might affect the results for Fe, Mn, and P, and by Si from the agate mortar, cannot be avoided , but when normal care is taken during the crushing and grinding process this contamination is impossible to detect. There is probably no contamination from the tungsten carbide; any contamination from this source could only add W, C, Co and Ti. Contamination by previously ground samples is negligible when a long series of rocks of the same type is being prepared for analysis. Before beginning grinding, however, the whole bench space was washed and wiped down and the walls were wiped as clean as possible. Between samples of the same rock type the glazed paper and sieves were brushed thoroughly. When changing from one rock type to another, a new piece of bolting silk was put in the 100 mesh sieve and all the glazed paper was changed. To minimise the contamination by rock flour and dust falling

on/

on to the sample being ground, it is advisable to work in a room provided with an extractor fan. Errors arising from loss of rock flour, oxidation of material by frictional heat on prolonged grinding, and from the fact that some minerals are ground down more rapidly than others can be minimised by sieving at frequent intervals when using the agate mortar, and by shaking the powder well before analysis. The errors arising from the fact that the whole of the original sample is not ground down to 100 mesh cannot be ignored. As shown in Table 7, when the rock is homogeneous the errors incurred by taking a 200-250 g. grab sample are very slight, but in foliated or porphyritic rocks they can become quite considerable. The errors involved in taking a 35-45 g. portion of 12 mesh material using a sample divider are negligible.

The precision of analytical methods is usually insufficient to tell whether the analysed material has suffered slight contamination during grinding. Even where contamination has been gross, it can only be assessed by subjective comparisons with previous analyses of rocks which are mineralogically and petrologically similar. It is therefore essential to maintain the best possible precautions against contamination, since an analysis which is seriously in error due to contamination may escape detection amidst a large number of good analyses.

3. MINERAL/

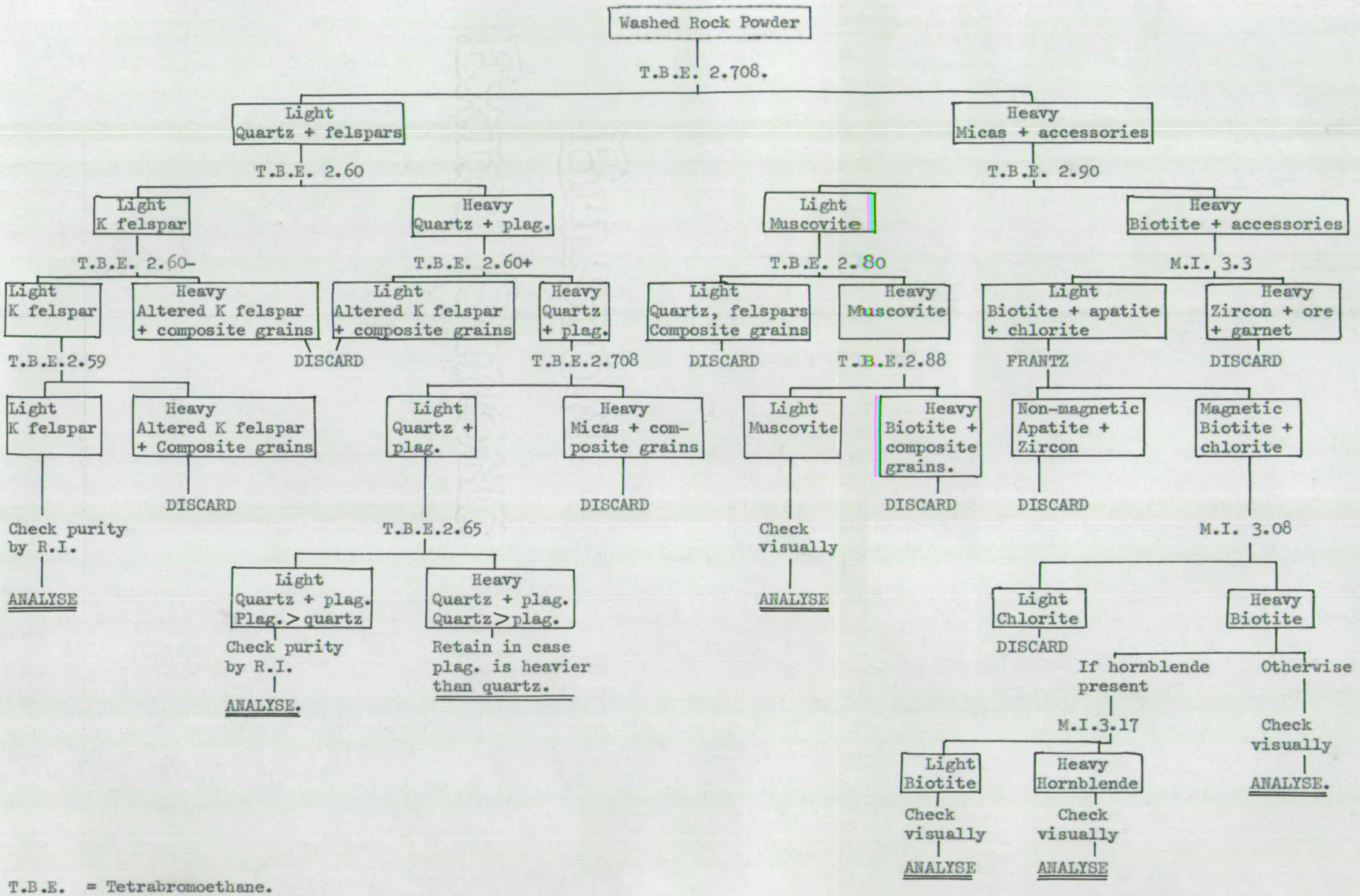
3. MINERAL SEPARATION

Nineteen rocks, including granitic gneisses, pelites, and psammites, were selected for mineral separation. Plagioclase, biotite and where possible potash felspar and muscovite were separated from all of these rocks. The separations were made by serial use of tetrabromoethane and methylene iodide, diluted to varying densities with acetone, and a Frantz Isodynamic separator. To check the densities of the heavy liquids, specific gravity beads were used, except when separating the (quartz + plagioclase) fraction into two parts, when the liquid's density was checked with a crystal of quartz (rock crystal). The method of separation is shown schematically in Table 8. It was often necessary to repeat a given separation until a fraction of sufficient purity was obtained and to adjust the density of the liquid slightly to obtain a better separation.

The purities of the felspar separations were checked against a mixture of butyl digol and 1-bromo-naphthalene having a refractive index of 1.458, purification being continued until all of the grains in the potash felspar separations had refractive indices less than 1.458, and until all of the grains in the (quartz + plagioclase) separations had refractive indices in excess of 1.458 as judged by the Becké test. The other mineral separations were inspected with a binocular microscope. Difficulty was found in estimating the purity of the biotites because/

TABLE 8.

Procedure followed in mineral separation.



T.B.E. = Tetrabromoethane.
M.I. = Methylene iodide (di-iodomethane).

Figures refer to the specific gravity of the liquid used at each stage.

because the dark colour and platy habit of the biotite led to underestimation of the amounts of light-coloured impurities. Contamination of the biotites by apatite was suspected even after the final purification, and the biotites were therefore analysed for P_2O_5. The presence of quartz in the plagioclase meant that the Na_2O, CaO and K_2O values for these separations had to be corrected. This was done on the assumption that the quartz was pure SiO_2 and the plagioclase contained the correct formula weight proportions of Al_2O_3 and SiO_2.

The errors due to contamination by other minerals were thus allowed for or kept to a minimum, though a slight contamination of the plagioclase by composite grains of potash felspar and quartz seems inevitable. In rocks of the types dealt with here, zoning is very slight and there is little likelihood that the material discarded when purifying a separation includes grains of the mineral sought which differ measureably in composition from the average of the purified fraction of that mineral.

4. CHEMICAL ANALYTICAL METHODS

The methods of analysis used were selected by consideration, not only of the precision and accuracy obtainable, but also of their feasibility in terms of the number of analyses which could be completed in the time available. The rocks and minerals were analysed for Si, Ti, Al, Fe (total), Fe^{2+}, Mn, Mg, Ca, Na, K, P, H and F (the last-named being determined in the micas only/

only) by "rapid" wet chemical methods. X-ray fluorescence spectroscopy was used to determine Rb, Sr, Zr and Ba, but with the equipment then available X-ray fluorescence spectroscopic methods were not really feasible for major element analysis.

Wet Chemical Methods: Four kinds of technique were used in the determination of major and minor elements in the rocks and minerals, (a) spectrophotometry using a Unicam SP500 spectrophotometer, (b) flame photometry using EEL Model A and Unicam SP900 flame photometers, (c) volumetric methods (d) gravimetric methods. The determinations of Si, total Fe (expressed as Fe_2O_3), Al, Mn, Ti, P, and F were all most conveniently done by spectrophotometric methods, except that the Si determinations on the biotites were done gravimetrically.

Determination of Si in whole rocks was done by spectrophotometry, using the heteropoly-blue form of silicomolybdic acid. The latter was produced by reduction of the yellow form of silicomolybdic acid with SO_2. This method has the advantage of being rapid and, under strictly controlled conditons, of producing a stable colour, but the maximum precision obtainable is only 0.95% (see Table 9). Most of the error arises in reading the galtanometer scale of the spectrophotometer. It was felt, however, that a more precise method for Si was unnecessary when dealing with rocks containing considerable quantities of quartz.

The /

The determination of Si in the biotites was done gravimetrically. After fusion with Na_2CO_3 and evaporation of the solution with H Cl, any remaining soluble SiO_2 was precipitated by addition of gelatin. This method is longer but more precise than the colorimetric method (see Table 10), even though some loss of precision and accuracy may result through incomplete precipitation of the SiO_2. The greater precision was felt to be necessary in the case of the biotite analyses, because in calculating the mineral formula the amount of Si^{4+} affects the proportion of Al^{3+} in 4-coordinated sites and hence the total occupancy of the 6-coordinated sites.

The spectrophotometric method used for Al was that of Mullen and Riley, which uses the complex with 8-hydroxyquinoline. By adding α-α_1-dipyridyl, hydroxylamine, $BeSO_4$ and sodium acetate, and adjusting the pH to 4.95 before extraction with 8-hydroxyquinoline dissolved in chloroform, interference was limited to that of titanium. A correction was applied to allow for the Ti effect. The precision of this method is low, the minimum error being $\pm$ 1.24% (Table 9), but no better wet method for Al seems to have been devised.

Spectrophotometric determination using the red colour produced by complexing with sodium citrate and adding ammoniacal thioglycollic acid is a very reliable and quick method of determining total iron in rocks and minerals. The precision is excellent at the 2 - 5% Fe_2O_3 level, and good even in biotites with/

with 25-30% total iron (as Fe_2O_3) (see Tables 9 and 10).

Oxidation of a strongly acid solution of the rock with periodate converts all the Mn in the rock to permanganate, which can be conveniently determined spectrophotometrically. Cr, the only element which interferes, could safely be assumed to be virtually absent from the rocks and minerals analysed.

For spectrophotometric determination of Ti, the yellow pertitanic acid produced by adding H_2O_2 to a sulphuric acid solution of the sample was used. Addition of H_3PO_4 prevented interference by Cr or Mn. The Tiron method could have been used for the whole rock analyses (0.1 - 1.0% TiO_2) but the precision of the peroxide method is quite adequate (see Table9), and it is more convenient to determine Mn, Ti, and P together on the same portion of sample.

A molybdenum blue complex similar to that used for Si was used in the spectrophotometric determination of P. Ascorbic acid, however, was used as the reducing agent. For the quantities in the rocks and minerals studied, the precision is quite good, and the method is more suitable than the Baadsgaard and Sandell method, which is preferable when the materials studied contain over 0.25% P_2O_5.

Only biotite and muscovite were analysed for F. Steam distillation of an acidified Na_2CO_3 fusion of the mineral was used to liberate the fluorine which was determined spectrophotometrically by the quantitative decolorisation of a Zr-SPADNS complex/

complex (Peck and Smith 1964). The method is moderately long, but appears to be very precise. A 4% correction has to be made for loss of fluorine to the glassware.

Flame photometry was used to determine Na and K. The sample solutions were diluted so as to contain 5 - 15 ppm Na_2O or K_2O (as sulphate) and the acid concentration was kept at 0.1 - 1.0% H_2SO_4. An EEL Model A flame photometer was used and standard solutions containing 5 - 25 ppm Na_2O or K_2O (as sulphate) in 0.1% H_2SO_4 were read at the same time as the sample. For determinations of K in the plagioclases, standards containing 2, 3, 4 and 5 ppm K_2O were employed and the 7.5 ppm K_2O solution was set to give a transmission of 100. The short length of the galvanometer scale and the slight unsteadiness of the instrument mean that high degrees of precision are not attainable, but the results are adequate for the purpose they are meant to serve. An analogous method was used to determine Li in the biotites.

The determination of Mg and Ca was done by titration with ethylene diamine tetra-acetic acid (E.D. T.A.). Before titration, Fe, Al, and Ti were removed by double precipitation with NH_4OH from a solution containing NH_4Cl. The calcium was titrated at a pH of 12 using KCN to complex heavy metals and triethanolamine to complex any Al left in the solution. Patton and Reeder's reagent was used as indicator. Ca and Mg together were titrated at a pH of 9, again after addition of KCN and triethanolamine, this time using eriochrome black T as indicator. Mg was calcula-

ted/

ted by difference. With such small amounts of Ca and Mg in the rocks, this method is not very precise (see Tables 9 and 10), and in the case of the biotites the large quantity of Mg tended to interfere with the titration of Ca. However, gravimetric methods were too long to be feasible and the additional precision was not felt to be needed.

In the case of the potash felspars, where there is very little Ca, an attempt was made to determine Ca with a Unicam SP 900 flame spectrophotometer. Using a 0.2 g. sample, R_2O_3 was precipitated as for the titration method. The transmission of the sample was determined at 423 mμ, and a correction for background obtained by averaging the transmissions at 418 and 428 mμ was subtracted. Standards containing 0, 2.5, 5 and 7.5 ppm CaO and matched for background with the felspars by containing 280 ppm K_2O, 22 ppm Na_2O, and 1 ppm Sr were read in the same way. On the 0.2 g. samples this method agreed quite well with the titration method, but for both methods precision is low at these levels of Ca content (see Table 10).

Ferrous iron was determined by titration with dichromate using sodium diphenylamine sulphonate as indicator. The rocks and minerals analysed required only 3 - 5 minutes simmering with H_2SO_4 and HF for complete solution. This method is very precise but is believed to be only 96% accurate. A method involving solution in cold HF, oxidation of the Fe^{2+} by VO_4^{3-} and titration

of/

TABLE 9.

Precision of determinations on granitic gneisses.

Element	Average weight % of oxide	Calculated precision as % of amount in rock.	Calculated precision as weight of oxide %	No. of pairs of duplicate analyses	as weight of oxide %		
					Average difference between repeat determinations.	S.D.‡	2x S.D.‡
Si	71.78	±0.95	±0.70	27	0.46	±0.33	±0.67
Ti	0.59	±0.97	±0.006	26	0.015	±0.011	±0.022
Al	13.12	±1.24	±0.16	26	0.22	±0.16	±0.32
Fe (total)*	4.13	±1.00	±0.04	27	0.056	±0.040	±0.080
Fe^{2+}	3.02	±1.02	±0.03	21	0.027	±0.020	±0.040
Mn	0.073	±1.31	±0.0009	26	0.0043	±0.003	±0.006
Mg	0.59	±8.52	±0.051	27	0.053	±0.037	±0.074
Ca	1.48	±2.71	±0.041	27	0.054	±0.038	±0.076
Na	3.18	±3.72	±0.11	25	0.085	±0.060	±0.12
K	4.47	±2.82	±0.13	26	0.071	±0.050	±0.10
P	0.164	±1.26	±0.0021	23	0.0099	±0.007	±0.014
H	0.52	±4.2	±0.02	107	0.045	±0.032	±0.064
Total	99.82		±1.35□				±1.48□

* as Fe_2O_3.

□ Obtained by adding together all entries in column above except those for Fe^{2+} and Mg. Variations in these determinations do not affect the total of the analyses, since Mg is determined by difference and Fe(total) = (Fe^{2+} x 1.1) + Fe^{3+}.

‡ S.D. = standard deviation; 2xS.D. = level of error exceeded in only 5% of cases.

TABLE 10.

Precision of determinations on separated minerals.

Determination	Average weight % of oxide	Calculated precision as % of amount in mineral	Calculated precision as weight of oxide %	Number of pairs of duplicate analyses	As weight of oxide %			
					Average difference between repeat analyses	Standard deviation	Standard error of mean□	2x Standard errors of mean.
Si (biotite-gravimetric)	34	±0.10	±0.03	5	-	-	-	-
Ti (biotite)	3.5	±1.09	±0.039	3	-	-	-	-
Al (biotite)	17	±1.28	±0.21	5	-	-	-	-
Fe (total) (biotite)*	27	±1.02	±0.28	18	0.18	0.13	0.09	0.18
Fe^{2+} (biotite)	22	±0.43	±0.095	18	0.12	0.09	0.06	0.12
Mn (biotite)	0.4	±1.25	±0.004	3	-	-	-	-
Mg (biotite)	5	±1,47	±0.075	4	-	-	-	-
Ca (biotite)	0.25	±7.57	±0.018	4	-	-	-	-
Li (biotite)	0.16	±9.25	±0.01	3	-	-	-	-
Na (biotite)	0.22	±11.35	±0.02	3	-	-	-	-
K (biotite)	9	±3.50	±0.32	18	0.104	0.074	0.052	0.104
H (biotite)	3	±1.72	±0.05	18	0.094	0.067	0.047	0.094
F (biotite)	0.28	±1.5	±0.004	4	-	-	-	-
Ca(plag.+ quartz)	3	±0.74	±0.022	5	-	-	-	-
Na(plag.+ quartz)	6	±2.77	±0.17	19	0.092	0.066	0.046	0.092
K (plag.+ quartz)	0.35	±4.35	±0.015	19	0.019	0.014	0.010	0.019
Ca (K-felspar -titration)	0.12	±7.6	±0.009	3	-	-	-	-
Ca (K-felspar-flame photometry)	0.12	±8.25	±0.010	3	-	-	-	-
Na (K felspar)	1.5	±4.18	±0.06	15	0.048	0.034	0.024	0.048
K (K felspar)	14.5	±2.37	±0.35	15	0.167	0.119	0.084	0.167

* as Fe_2O_3

□ Standard error of mean = standard deviation x $1/\sqrt{2}$.

of the excess VO_4^{3-} with ferrous ammonium sulphate, was tried, but gave very erratic results.

The determination of H was done gravimetrically. A mixture of 1 g. powder and 3 g. flux (2 parts PbO + 1 part $PbCrO_4$) was heated for five minutes in a Pyrex test-tube and the water given off was collected in the cooled upper part of the test-tube. A 10% correction was added to allow for the escape of any uncondensed steam. All determinations were done in duplicate. Precision is not good, and the method was repeated until concordant results were obtained (less than 0.1% H_2O apart for $H_2O < 1\%$ and 0.1% difference for every 1% H_2O thereafter).

5. ERRORS IN WET METHODS.

The errors involved in analysis by wet methods may be divided into two groups: (a) instrumental errors, including the errors in the graduation of flasks and pipettes as well as the errors involved in reading burettes and the galvanometer scales of spectrophotometers and flame photometers, and (b) errors arising from the nature of the methods used, for example enhancement or depression of the emission of Na in a flame by other elements, adsorption of Ca^{2+} and Mg^{2+} on the R_2O_3 precipitate, and loss of SiO_2 in the gravimetric method through incomplete precipitation. Errors belonging to group (a) can be calculated quite easily and affect only the precision and not the accuracy. (Errors belonging

to group (a) can be calculated quite easily and affect only the precision and not the accuracy.) Errors belonging to group (b) can affect the accuracy of the methods quite considerably and often affect the precision as well. Unfortunately they are much more difficult to calculate and this can only be done if a detailed study of the method is made.

The methods used here have all been previously checked against G-1 and W-1, except for the flame photometric method for calcium. A single determination of Ca in G-1 by flame photometry gave 1.33% CaO, compared with the recommended value of 1.36% (Stevens and Niles 1960, corrected for Sr). Also Si was occasionally determined in G-1 by the colorimetric method. The average of 11 determinations ranging from 72.54 to 73.09 % SiO_2 was 72.77%. This is in good agreement with the recommended value of 72.70% (Stevens and Chodos 1960).

The instrumental error involved in the methods above was calculated, and this was compared with the precision calculated from the results of the repeat determinations which were made during analysis. The calculation of instrumental error was based on the tolerances for pipettes, burettes and graduated flasks given in British Standards Institution specifications BS1583, 846, and 1792 respectively (quoted in Gallenkamp 1964, pp. 626, 87, and 327). These tolerances are reproduced in Table 11.

The 5-place balance used for the molybdenum blue method for Si/

TABLE 11.

Tolerances for pipettes, burettes and graduated flasks, taken from B S 1583, 846 and 1792 (quoted by Gallenkamp 1964, pp.626, 87 and 327).

Pipettes - Grade A.

Size (ml.)	2	3	5	10	15	20	25	50	100
Tolerance (ml.)	±0.01	±0.015	±0.015	±0.02	±0.025	±0.03	±0.03	±0.04	±0.06

Graduated flasks - Grade B.

Size (ml.)	25	50	100	200	250	500	1000
Tolerance (ml.)	±0.06	±0.1	±0.15	±0.3	±0.3	±0.5	±0.8

Burettes - Grade A.

Size (ml.)	50
Tolerance (ml.)	±0.05

Si was assumed to have an error ± 0.01 mg. at the 7-8 g. level, and the 4-place balance used for all other methods was assumed to have an error of ± 0.1 mg. at all weights from 0 to 30 g.

It was assumed that a 50 ml. burette, could be read to ± 0.02 ml. and that the end-point could be judged to the nearest drop, that is 0.03 ml. The burette was not read till one minute after the end of the titration to minimise drainage errors, which were not included in the calculation of total error.

The galvanometer scale of the spectrophotometer was read to the nearest 0.001 logarithmic absorption units and the instrument reading was, as far as possible, kept between 0.250 and 0.650 logarithmic units. The precision of the instrument was taken as ± 0.001 logarithmic absorption units.

The EEL flame photometer has a galvanometer scale graduated from 0 to 100 in steps of 2. The sample was always read between two standards differing in Na_2O or K_2O concentration by 2.5 ppm, and this difference usually corresponds to 10 - 12 scale units in the range 5-15 ppm Na_2O or K_2O. It was assumed that the Na_2O or K_2O concentration could be obtained with a precision of ± 0.25 ppm, that is, the sample and standard readings would not be in error by more than ½ scale unit.

In the present case, it is possible to estimate the precision of the wet methods from the repeat analysis which was made of every fourth rock, and from the fact that all Na and K determinations on the felspars and all Fe^{2+}, total Fe, K and H determinations on the biotites/

biotites were done in duplicate. The amounts of the various elements in the granitic gneisses are sufficiently similar for the differences between each pair of determinations to be used as an estimate of the precision of the method at a concentration level equivalent to the average concentration of the element in question in the granitic gneiss. The best way to express the precision of the method is as a standard deviation. This may be calculated as follows:

Consider a pair of analyses of one rock for one oxide. Let the results be

x_1, x_2; let the arithmetic mean be $\bar{x}$ and the number of determinations n. Let s be the standard deviation of an infinite number of determinations of one oxide in one rock by the same method.

Then the best estimate of $s^2 = \frac{\sum(x - \bar{x})^2}{n - 1} = \frac{(x_1 - x_2)^2}{2}$

Therefore, the best estimate of $s = \frac{x_1 - x_2}{\sqrt{2}}$

The precision of the method may thus be estimated by averaging $(x_1 - x_2)/\sqrt{2}$ for all the pairs of duplicate analyses.

An insufficient number of pelites or psammites was analysed in duplicate to give a separate estimate of the precision of the methods, but for most of these rocks the amounts of the various elements are sufficiently similar to the amounts of the same elements/

elements in the granitic gneisses for the same estimates of precision to be used. The chief exception to this is in the case of the most quartz-rich psammites, where the amounts of Mn, Mg, Ca and P are often very low, and the precision drops considerably.

The precision of the methods as applied to the analysed minerals can only be reliably estimated where all the minerals of a given kind have been analysed in duplicate for a given element. In these cases, not only can the error of a single determination be calculated, but the standard error of the mean of the two determinations is found by dividing the standard error of a single determination by $\sqrt{2}$.

The errors of the various methods are presented in Tables 9-10. The value (2 x standard deviation) gives the level of error which would be expected, from the repeat determinations made, to be exceeded in only 5% of all determinations. This level is compared with the calculated error which is regarded as the maximum possible error which can arise from the weighing, pipetting, dilution, and from the reading of the instrument. In most cases, the actual error is considerably greater than the calculated error, and this is largely due to the numerous incalculable sources of error. The tables show that whereas most of the error in the gravimetric method for Si is non-instrumental and the method can give good precision at all levels of Si concentration, the error in the colorimetric Si method is almost all instrumental, and the precision obtained in the whole rock analyses is almost the/

the maximum possible.

6. X-RAY FLUORESCENCE SPECTROSCOPY

X-ray fluorescence spectroscopy was used for determination of Rb , Sr, Zr and Ba in rocks and Rb, Sr and Ba in minerals. The equipment used consisted of:

X-ray generator	- Philips PW 1010/30.
Goniometer	- Philips PW 1050/25.
Electronic circuit panel	- Philips PW 1051/30.
Universal vacuum spectrograph	- Philips PW 1540.
Spectrograph discriminator cabinet	- Philips PW 1523.

A tungsten target was used throughout for generation of X-rays. In all cases, the 100-mesh powder used for analysis was placed in the sample holder on a film of Mylar ("melinex") covering its base. The operating conditions are listed in Table 12.

The peaks used in these determinations are all known to be free from interference by elements likely to be present in appreciable quantities in rocks. To set up the apparatus for a given element it was necessary to determine the location of the peak using a pure compound of the element to be determined, and to adjust the optimum settings of attenuation, channel width, and amplitude to maximise the peak/background ratios and stability of counting.

Counting/

TABLE 12.

Operating conditions of X-ray spectrograph.

Element	E.H.T. generator		Element Peak and Order	Analysing Crystal	Wavelength, as 2θ of crystal		Preparation	S.C.V*	Vacuum?
	Voltage, kV	Current mA			Peak	Background			
Rb	40	20	Kα1 (1)	Topaz	39.60	42.40	Mylar	850	No
Sr	40	20	Kα1 (1)	Topaz	37.63	38.63	Mylar	850	No
Zr	40	20	Kα1 (1)	Topaz	33.70	35.00	Mylar	850	No
Ba	40	20	Kα1 (1)	Topaz	16.37	15.57 and 17.17	Mylar	850	No

* Scintillation counter voltage.

Counting time was 16 seconds and 6 to 8 counts each of peak and background were averaged for each determination. The determinative curves (Figures 3 - 6) were drawn by irradiating a number of international standard materials and plotting the peak/background count ratio against the accepted values for the standards shown in Table 13. The samples were then irradiated and the peak/background ratio calculated. To allow for any secular variation in the count level, the USGS standard granite G-1 was irradiated with each group of three samples, and the peak/background ratios for the samples were adjusted proportionally to give a constant peak/background ratio for G-1. The amounts of the element in question in the samples were then read off the graph against the corrected peak/background ratio.

The main sources of error in X-ray fluorescence determinations are:

(i) Errors in the accepted trace element values of the international standard materials used to draw the working curves. Many of these materials have only recently become available and few workers have determined their trace element concentrations. The USGS standard samples G-1 and W-1 have, however, been analysed repeatedly, and the amount of agreement between the results of different workers varies considerably for different elements. The most complete lists of determinations on G-1 and W-1 are given in Ahrens and Fleischer (1960) and Fleischer and Stevens (1962), and these are supplemented by the determinations of Taylor and Kolbe (1964/

TABLE 13.

Trace element values of international standard materials used to set up working curves.

Standard		ppm Rb	ppm Sr	ppm Zr	ppm Ba
U.S. Geological Survey	Granite G-1	220*	250-280‡	185*	1250*
	Diabase W-1	22*	175-210‡	110*	170*
	Granite G-2	173x	500x		2100x
	Granodiorite GSP-1	250x			
	Andesite AGV-1	64x	670x		1400x
	Dunite DTS-1	0x			
U.S. Bureau of Standards	Milford Granite 4983		145*	105*	2000*
	Silica Brick 102		20*	240*	50*
	Opal Glass 91			47°	70*
	Burnt Refractory 76			620°	
Canadian Association for Applied Spectrocopy	Syenite Sy-1		205*		330*
Johnson Matthey - "Specpure" SiO_2			<5	<5	
Dr. B.G.J. Upton	Alkalifelspar from 40528				4800■
	Alkali felspar from 40521	120□			

* From Taylor and Kolbe (1964).

° From Filby (1964)

‡ From Ahrens and Fleischer (1960 p.99).

□ From Upton (1964, p.43).

■ Upton, B.G.J., unpublished.

x Price, N.B., and Angell, G.R., unpublished.

(1964), Filby (1964), and Kaye (1965). The precision of the values of Taylor and Kolbe (1964), which, apart from Sr in G-1 and W-1, have been used wherever possible, is quoted as 7% for Rb, 8% for Sr, 7% for Zr and 5% for Ba. The accuracy of their results cannot be evaluated quantitatively, but in the case of Rb and Ba the results for G-1 and W-1 compare well with those recommended by Fleischer and Stevens (1962), though in the case of Sr and Zr agreement is less good.

(ii) Non-linearity of the relationship between the concentration of the trace element and the peak/background ratio. Clarke (1965), however, found that for Ba this effect only becomes serious at about 10,000 ppm Ba.

(iii) Subjective errors in drawing the graphs.

(iv) Errors due to the random distribution in time of the X-ray photons given off by the atoms of the element in question in the sample and variation in the intensity of the X-rays produced by the tube. For these two reasons, count stability with the instrument used was low, and this was probably the most important source of error, as it affected the results for each sample differently. Most of the other sources of error affected all the results to a similar extent.

(v) Errors due to scattering of X-rays. 100 mesh material was used, whereas grinding to 200 mesh would have reduced scattering considerably. This error, however, affected all results to the same extent.

(vi) /

(vi) Background effect. The level of counts was much higher for some materials, such as felspars, than for others, such as biotites, and no tests were made to see whether the same peak/background ratio at different count levels did in fact correspond to the same concentration of the element being determined.

The actual quantitative measurement of error in the case of trace elements is difficult. Precision is poor below about 50 ppm Rb, 100 ppm Sr, 100 ppm Zr, or 500 ppm Ba, but improves proportionately with the increase in concentration. Accuracy decreases, however, because of the lack of natural standards at levels above about 300 ppm Rb, 1000 ppm Sr, 600 ppm Zr, and 5000 ppm Ba.

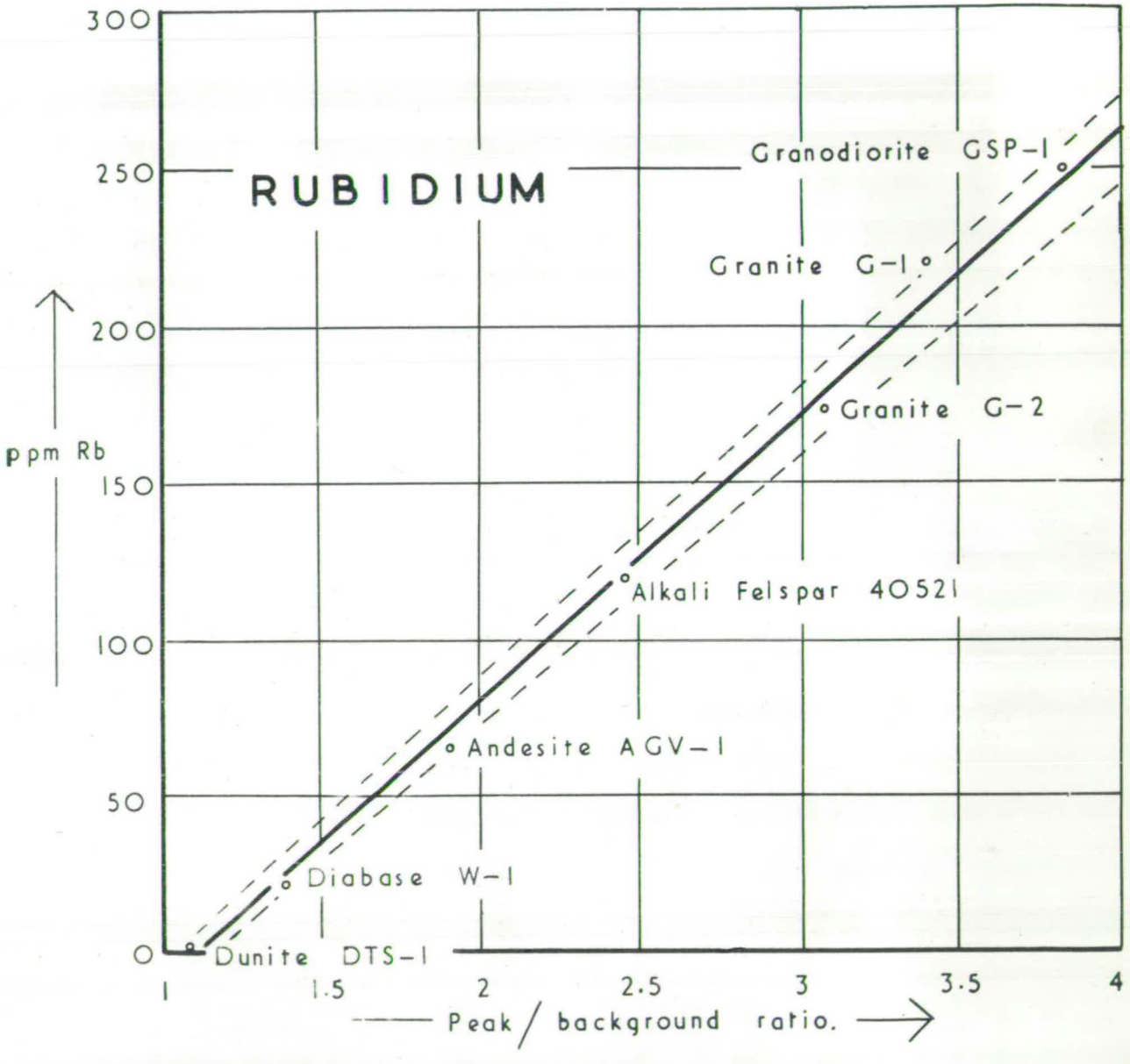
The error involved in the calculation of each peak/background ratio appears to be about 1% (calculated by adding together the standard errors of the means of peak and background counts). This error is doubled when the ratio is corrected by comparison with G-1. Translation of these errors into ppm of trace element is difficult, but the dotted lines on the graphs (Figures 3 - 6) give an approximate idea of the level of error believed to exist in these determinations. For example, in determining Ba, a peak/background ratio of 1.15 would indicate 50-120 ppm Ba, 2.00 would indicate 1620-1820 ppm Ba, and 3.50 would indicate 4500-4930 ppm Ba.

FIGURES 3 - 6.

Working curves for the determination of certain elements by X-ray fluorescence spectroscopy.

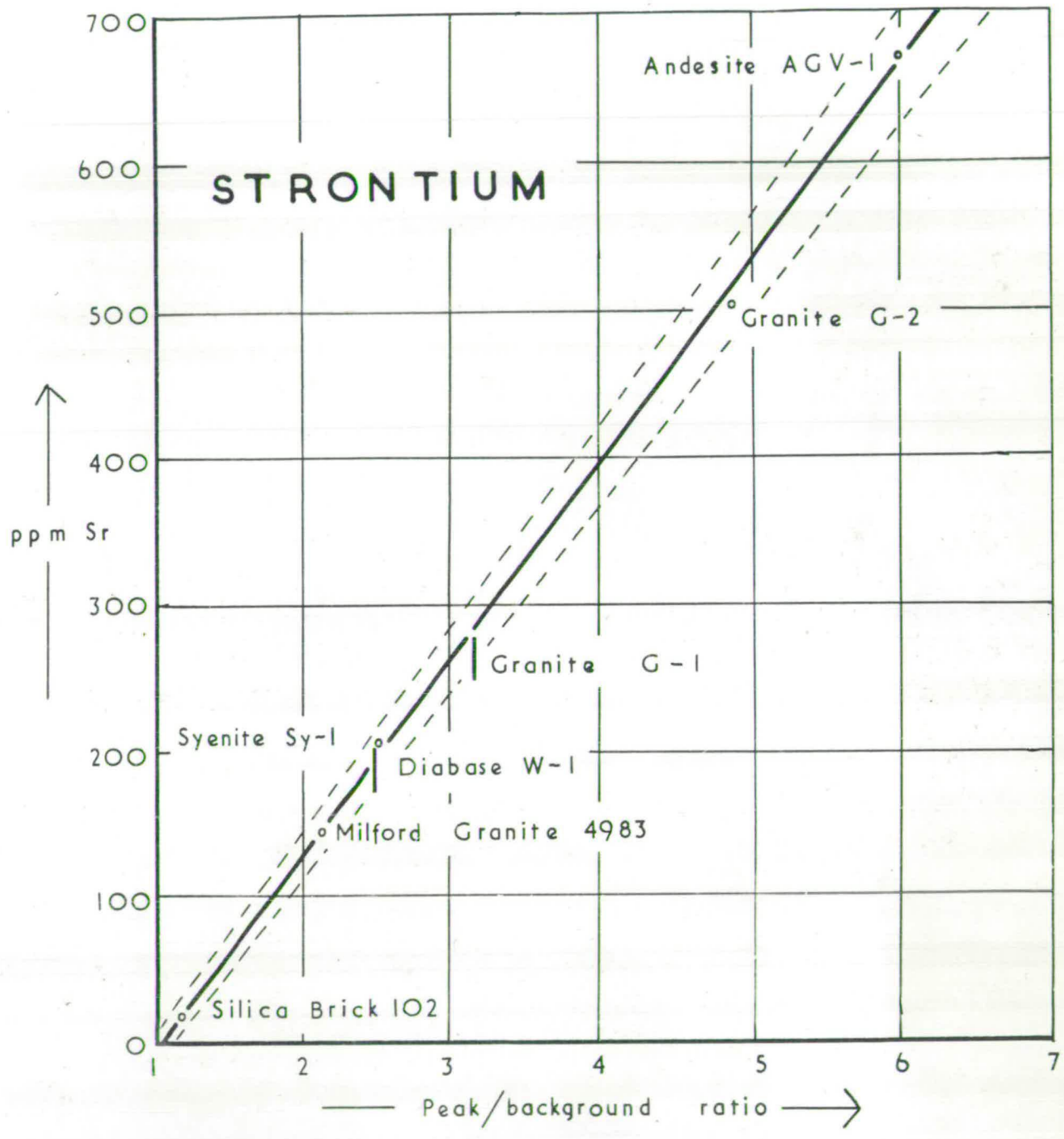
Figure 3 - Rubidium
Figure 4 - Strontium
Figure 5 - Zirconium
Figure 6 - Barium

The international standard materials used to set up these curves and the recommended values for their Rb, Sr, Zr, and Ba contents are listed in Table 13.



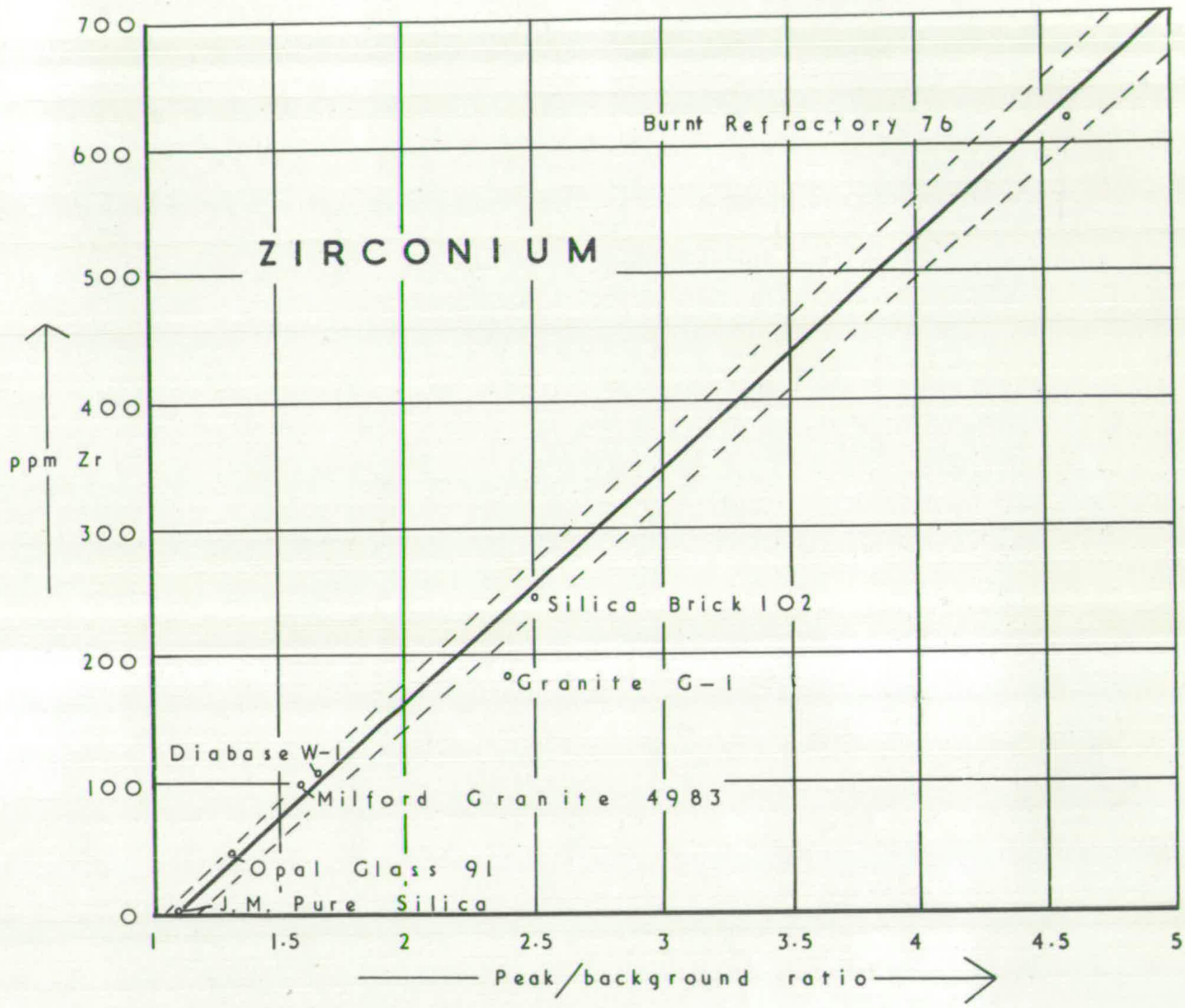
Full line — Determinative curve.
Broken lines — Postulated limits of experimental error.

FIG. 3.



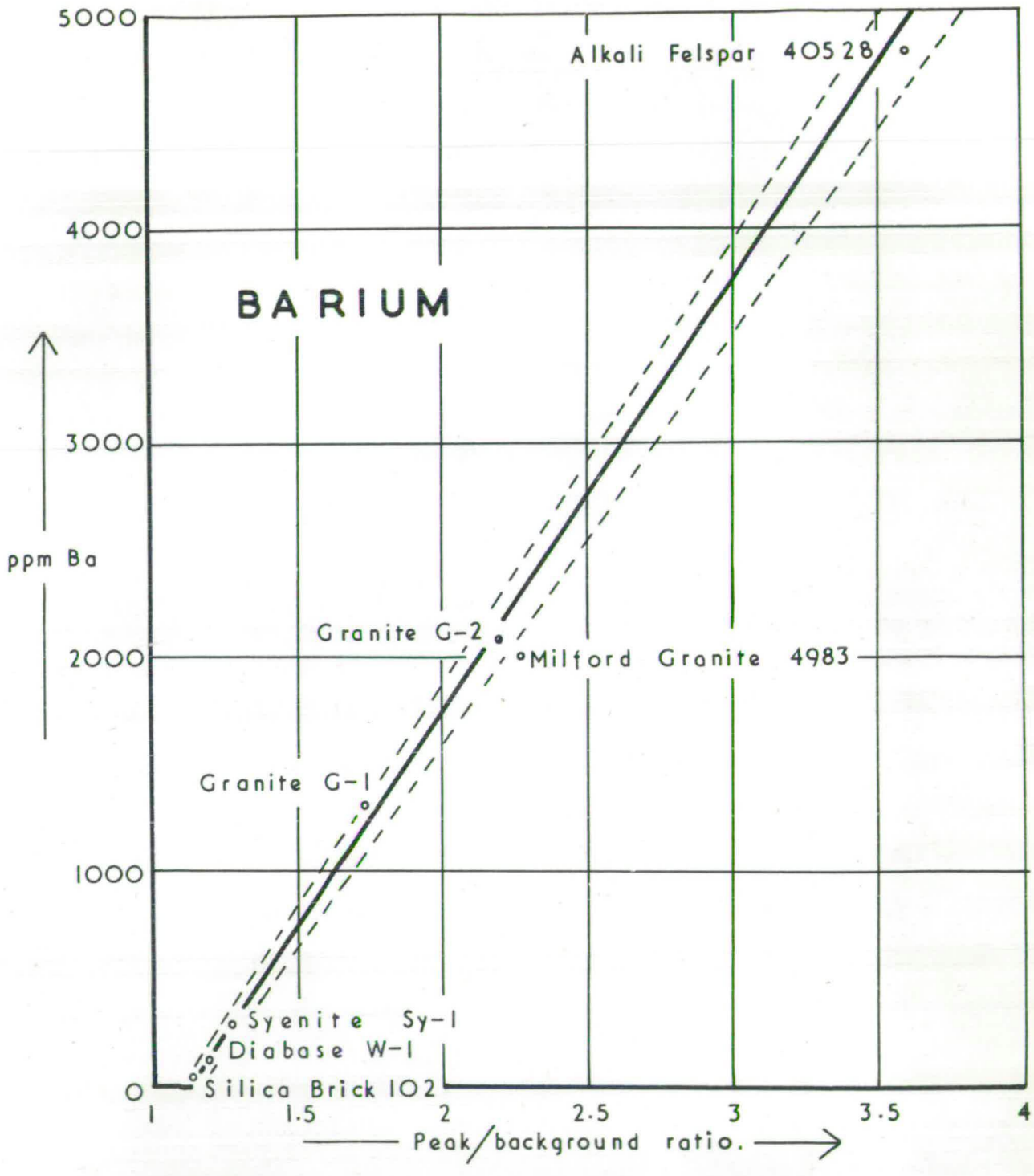
Full line — Determinative curve.
Broken lines — Postulated limits of experimental error.

FIG. 4.



Full line — Determinative curve.
Broken lines — Postulated limits of experimental error.

FIG. 5.



Full line — Determinative curve.
Broken lines — Postulated limits of experimental error.

FIG. 6.

IV. THE CHEMICAL COMPOSITION OF THE GRANITIC GNEISS

The chemical compositions of the analysed samples of granitic gneiss are expressed in weight percentages and 1-cation percentages in Tables I-VII, where they are divided on a geographical basis into seven groups (Fig. 8). The compositions of the granitic gneiss samples are also expressed as Eskola molecular norms, first using the C.I.P.W. sequence of normative minerals (Table XIII), and then using a sequence of normative minerals based on the mineral analyses presented in Chapter Six. The sequence of normative minerals and their compositions are listed in Table XIV and the norms are given in Table XV. The effect of experimental error on the precision of the values quoted in both types of norm is shown in Table XVI; these values are based on twice the standard deviations for experimental error (Table 9, last column) and correspond to a 95% confidence level. Tables I-VII and XIII-XVI will be found in the Appendix.

1. COMPARISON OF THE COMPOSITION OF THE GRANITIC GNEISS WITH THE RESULTS OF EXPERIMENTAL WORK ON ANATEXIS.

The first experimental work to give any indication of the likely composition of anatectic melts was that of Tuttle and Bowen (1958), whose work established the composition of the isobaric liquidus minimum in the system $KAlSi_3O_8$ - $NaAlSi_3O_8$ - SiO_2 - H_2O for values of p_{H_2O} from 500 to 4000 bars. They showed that/

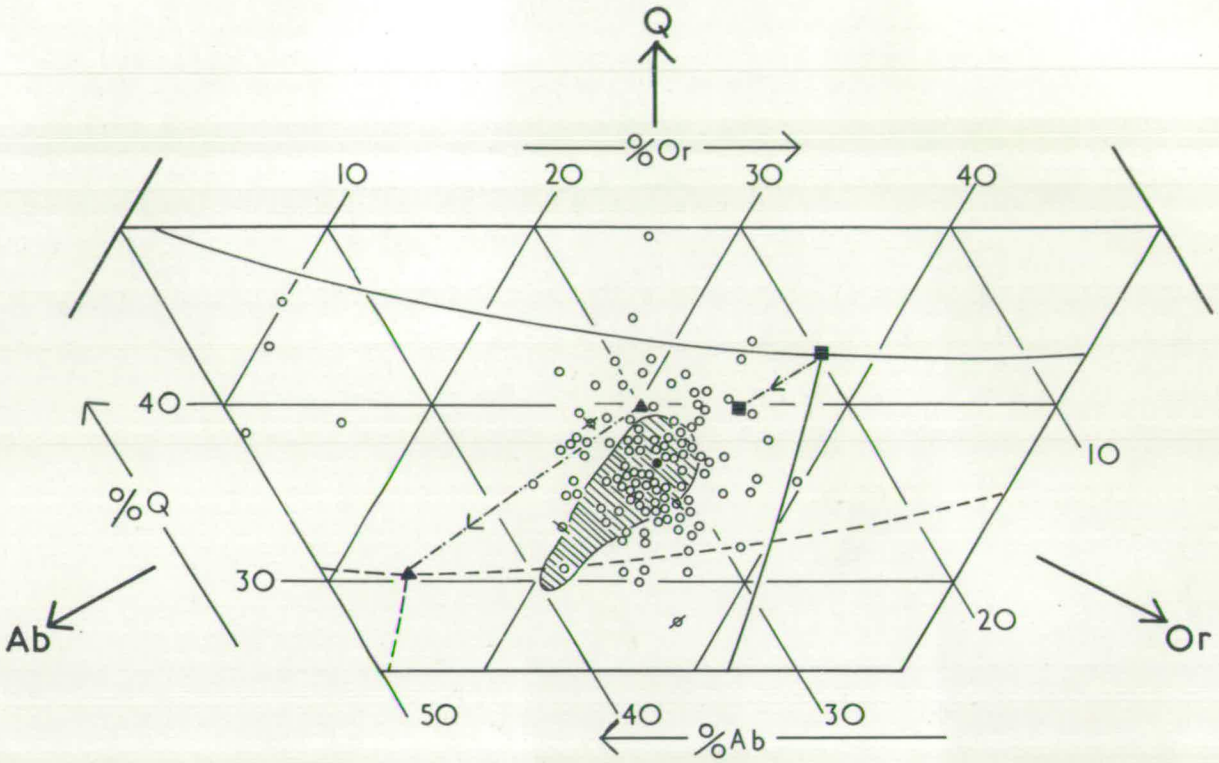
that the temperature of the liquidus increased much more slowly away from the minimum on moving along the quartz-felspar boundary than on moving along lines of equal Ab/Or ratio (Tuttle and Bowen 1958, Figs. 22-5 and 38-40). The composition of the ternary liquidus minimum was shown to be similar to that of the bulk of previously analysed volcanic and plutonic rocks containing over 80% of normative (Q + Or + Ab) (Figs. 41-42). Experimental work on samples of the Quincy and Westerly granites showed that they commenced melting at c.660°C at 4000 bars p_{H_2O}.

The Ardgour granitic gneiss analyses are plotted on a Q - Or - Ab diagram in Fig. 7 (only part of the triangle is shown). Most of the analyses plot fairly close to Tuttle and Bowen's isobaric liquidus minima for 2000 - 4000 bars p_{H_2O}, but are displaced from it in a direction opposite to that of the Ab corner, as are the majority of the plutonic rocks shown in Tuttle and Bowen (1958), Fig. 42.

More recent work with a bearing on the composition of anatectic melts and the temperature of their formation is summarised in Winkler and von Platen (1961) and von Platen (1965). Working at p_{H_2O} = 2000 bars on greywackés and mixtures of clay with varying amounts of NaCl and $CaCO_3$, Winkler and von Platen (1961) found that, on heating, a partial melt began to form at 660 - 740°C and at a temperature of 810°C, 40-80% of the original material was molten. The composition of the first liquid/

FIGURE 7.

Comparison of analyses of Ardgour granitic gneiss with experimental work in the system $NaAlSi_3O_8$ - $KAlSi_3O_8$ - SiO_2 - H_2O.



Q, Ab, and Or are expressed as weight percentages. Only part of the triangle is shown.

- ∘ – Individual granitic gneiss analysis.
- • – Average granitic gneiss composition.
- ⊗ – Concordant pegmatite.
- ⌀ – Granite-like rock believed to be derived from psammite.
- ✕ – Aplite.

Primary phase boundaries in Ca-free system for p_{H_2O} = 4000 bars (Tuttle and Bowen 1958).

Displacement of isobaric liquidus minimum with increase in p_{H_2O} from 500 to 4000 bars (Tuttle and Bowen 1958).

Primary phase boundaries for Ab/An 3·8 at p_{H_2O} = 2000 bars (von Platen 1965).

Suggested analogous movement of von Platen's liquidus minimum on increasing p_{H_2O} from 2000 to 4000 bars.

Most common composition of granitic rocks (Tuttle and Bowen 1958).

FIG. 7.

liquid to form depended on the relative proportions of K, Na, and Ca in the starting material, but lay slightly on the quartz-rich side of Tuttle and Bowen's (1958) experimentally determined quartz-felspar boundary on the isobaric liquidus at 2000 bars, and well to the Or-rich side of the liquidus minimum composition. As the temperature increased, the liquid composition moved nearer to the point representing the starting material, which was usually considerably more sodic than the first-formed liquid.

Additional experimental work using mixtures of obsidian and anorthite and vapour phases of water and 0.05 molar HCl at vapour pressures of 2000 bars (von Platen 1965) showed that the composition of the liquidus minimum in the system Q - Or - Ab varies systematically with the Ab/An ratio in the norm of the starting material. This variation was also deduced from previous experimental work by Kleeman (1965) whose Fig. 3 illustrates the melting behaviour of compositions lying on the silica-saturated surfaces of the tetrahedron $KAlSi_3O_8$ - $NaAlSi_3O_8$ - $CaAl_2Si_2O_8$ - SiO_2 at 5000 bars P_{H_2O}, by projecting from SiO_2 on to the Or - Ab - An face of the tetrahedron.

The normative Ab/An ratio (as weight%) for the average Ardgour granitic gneiss is 4.4. Von Platen's experimental cotectic line and eutectic point for obsidian + 8% An (Ab/An = 3.8) with H_2O alone (1965, Figs. 3 & 4) are included in Fig. 7. If the position of this eutectic point alters with increase in P_{H_2O} in

a/

a similar way to the eutectic in the pure system $NaAlSi_3O_8$ - $KAlSi_3O_8$ - SiO_2 - H_2O it will lie very close to the average composition of the Ardgour granitic gneiss at p_{H_2O} = c. 5000 bars. The temperature of beginning of melting for (obsidian + 8% An) was found to be 690 - 700°C. By analogy with the behaviour of the Quincy and Westerly granites (Tuttle and Bowen 1958, Table 12, p. 82.), this would be reduced to 660 - 670°C at p_{H_2O} = 4000 bars.

Von Platen also showed that a partial melt of granitic composition and rich in the K-felspar component is produced by partial melting at 2000 bars p_{H_2O} of metasedimentary rocks lacking potash felspar but containing a potassium-rich phase such as biotite or muscovite.

The evidence presented above shows that the proportions of salic constitutents in the Ardgour granitic gneiss are those which would be expected if the granitic gneiss crystallised from a liquid formed by partial melting of some other material at the lowest possible temperature. It also shows that if the granitic gneiss after formation was heated to temperatures at which partial melting could commence, almost all of the quartzo-felspathic material of the granitic gneiss would become liquid within a very small range of temperature, if the requisite amount of water was present (a granitic liquid at 660°C and p_{H_2O} = 5000 bars contains 8 - 9% water (Tuttle and Bowen 1958, Figs. 26, 27, pp./

pp. 56-57)). The implications of these conclusions will be considered in Chapter Seven (pp.152-153).

2. VARIATION IN COMPOSITION OF THE GRANITIC GNEISS FROM PLACE TO PLACE WITHIN ITS OUTCROP.

The variation in composition of the granitic gneiss with locality was examined in order to compare it with the variation in composition within their outcrop of typical intrusive granitic bodies such as those studied by Whitten (many papers, especailly 1961b) and others. Unfortunately, nearly all of these workers used modal analyses as the basis for their statistical studies. For the reasons stated on pages 27-31, modal analysis was considered too imprecise to give useful results when applied to rocks so inhomogeneous on a hand-specimen scale as the granitic gneiss. Chemical analyses were considered not only to be more precise, but to give more information which might be useful in determining the origin of the granitic gneiss. The major element data were used in the form of one-cation percentages, because it was felt that this would give a truer estimate of the chemical variation within the granitic gneiss.

The variation in the composition of the granitic gneiss was studied statistically in two ways: (a) the granitic gneiss samples were divided on a geographical basis into seven groups and the means and standard deviations for each element within each group calculated and compared; (b) a trend surface analysis was carried/

carried out for each major element using a programme for the Atlas computer developed by Dr. M.R. Osborne (Computer Unit, Edinburgh University) in connection with a stratigraphical study conducted by Dr. E.K. Walton (Geology Department, Edinburgh University).

(a) Division of granitic gneiss into seven groups. The areas covered by the groups are listed below and shown in Fig. 8.

Group I - The southern tongue of granitic gneiss south of Sgurr a'Chaoruinn.

Group II - The large expanse of granitic gneiss lying south-west of Sgurr Dhomhnuill.

Group III - The western tongue of granitic gneiss west of National Grid line 1/86.

Group IV - The belt of granitic gneiss from Druim Garbh and Lochan Dubh.

Group V - The part of the granitic gneiss outcrop from Meall Mor to Meall nan Damh.

Group VI - The northern slopes of Meall nan Damh as far as the National Grid line 7/785.

Group VII - The northernmost part of the granitic gneiss outcrop lying east of Glenfinnan.

The chemical analyses of the individual samples forming Groups I - VII are listed in Tables I - VII respectively (Appendix).

FIGURE 8.

Division of Ardgour granitic gneiss samples into seven regional groups.

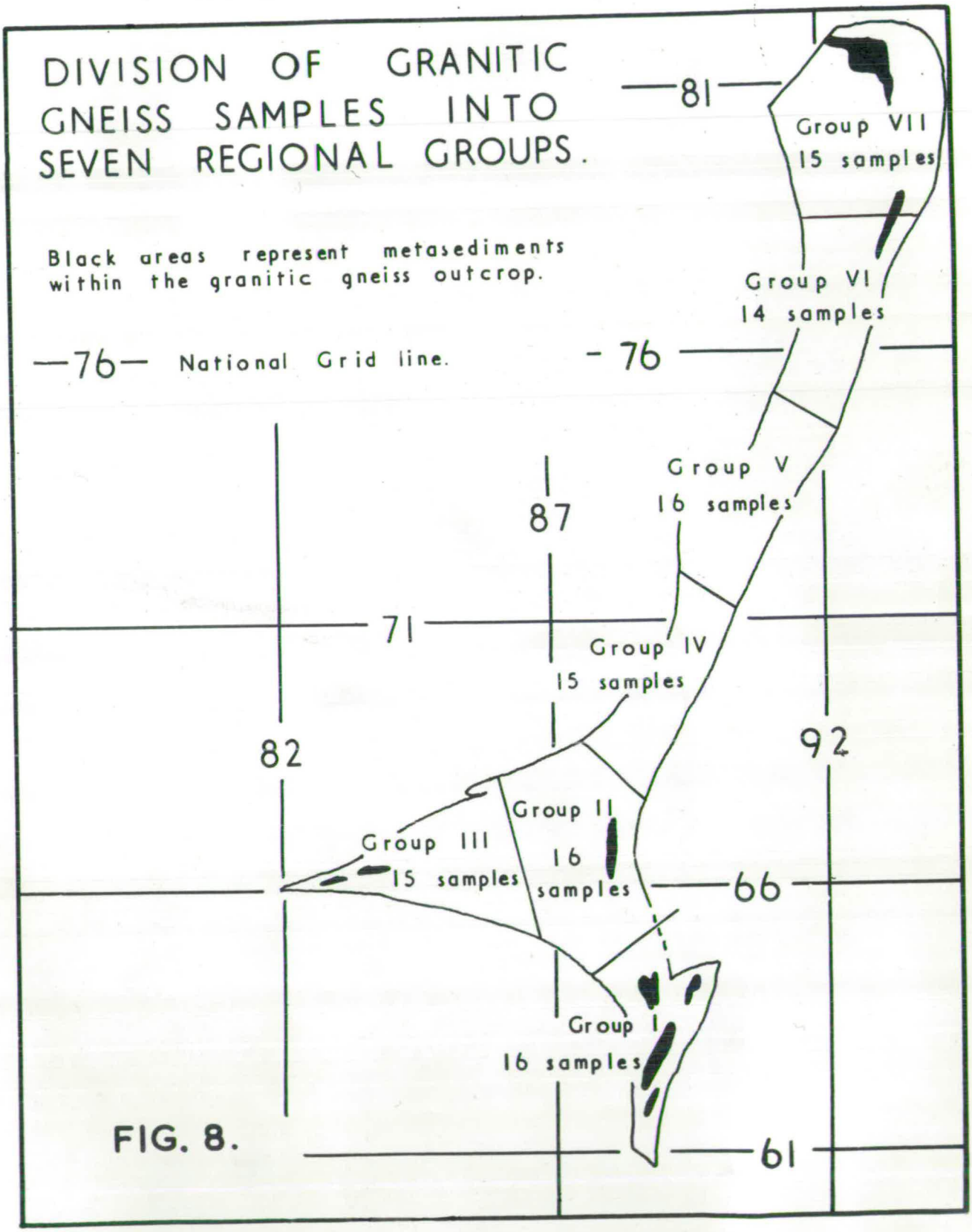

DIVISION OF GRANITIC GNEISS SAMPLES INTO SEVEN REGIONAL GROUPS.
Black areas represent metasediments within the granitic gneiss outcrop.
—76— National Grid line.
—81—
Group VII
15 samples
Group VI
14 samples
- 76
Group V
16 samples
87
71
Group IV
15 samples
82
92
Group II
16
samples
Group III
15 samples
66
Group I
16 samples
61

FIG. 8.

The mean compositions and the standard deviations for each group and for the granitic gneiss as a whole are given in Table 14. In order to avoid distortion of the pattern of regional variation in composition, the samples G13 (Group IV), G221 (Group VI), G274 and G283 (both Group VII), which contain very little potash felspar and unusually large amounts of plagioclase, were ignored in the calculations for Ca, Na, K, Rb, Ba and Sr whether or not the value for any one of these elements seemed anomalous compared with the values for that element in the neighbouring granitic gneisses. Sample 103 (Group II) was ignored in the calculations for Rb, Ba and Sr because of the anomalously high Ba and Sr content of this rock in which it resembles the sample GS166 (Table XII, Appendix).

The standard deviations in Table 14 were calculated (Hoel 1947, pp. 10-13 and 129) using the formula

$$\sigma = \sqrt{\frac{n}{n-1}\left[\frac{\sum(x^2)}{n} - (\bar{x})^2\right] - s_e^2}$$

where $\sum(x^2)$ is the sum of squares of the values of an element in the group.

$\bar{x}$ is the arithmetic mean of the values of that element.

n is the number of samples in the group.

s_e is the standard deviation which would be found in repeat analyses of the same sample (taken from Table 9).

and/

TABLE 14.

(a) Average compositions of granitic gneisses.

	Group I	Group II	Group III	Group IV	Group V	Group VI	Group VII	All granitic gneisses.
Si^{4+}	69.47	69.04	68.99	67.74	67.84	66.48	67.05	68.12
Ti^{4+}	0.37	0.38	0.41	0.43	0.42	0.45	0.45	0.41
Al^{3+}	14.11	14.32	14.32	14.77	14.60	15.52	15.16	14.67
Fe^{3+}	0.80	0.74	0.54	0.71	0.78	0.68	0.67	0.70
Fe^{2+}	1.55	1.84	2.24	2.36	2.29	2.80	2.82	2.26
Mn^{2+}	0.046	0.050	0.055	0.063	0.063	0.066	0.072	0.059
Mg^{2+}	0.66	0.75	0.81	0.81	0.83	0.98	0.96	0.82
Ca^{2+}	1.52	1.43	1.34	1.67	1.48	1.49	1.60	1.50
Na^{+}	5.62	5.74	5.54	5.96	6.08	5.90	5.64	5.78
K^{+}	5.77	5.59	5.66	5.37	5.47	5.44	5.52	5.55
P^{5+}	0.089	0.114	0.121	0.139	0.150	0.165	0.139	0.130
Fe^{2+}/Fe^{3+}	2.35	2.52	2.77	3.07	3.06	3.48	3.49	2.96
Rb	169	181	187	178	180	168	169	176
Ba	768	813	759	816	814	902	929	825
Sr	140	148	138	155	149	187	179	155
Zr	372	376	400	455	442	499	538	438

(b) Standard deviations, corrected for experimental error and small sample effect.

Si^{4+}	1.82	1.83	1.55	0.66	0.84	0.76	0.76	1.64
Ti^{4+}	0.12	0.10	0.07	0.05	0.04	0.06	0.06	0.08
Al^{3+}	0.70	0.94	0.56	0.33	0.35	0.34	0.29	0.71
Fe^{3+}	0.32	0.33	0.14	0.30	0.23	0.14	0.18	0.26
Fe^{2+}	0.68	0.57	0.55	0.37	0.33	0.48	0.51	0.67
Mn^{2+}	0.015	0.014	0.015	0.007	0.009	0.006	0.016	0.015
Mg^{2+}	0.33	0.30	0.19	0.16	0.10	0.12	0.21	0.24
Ca^{2+}	0.44	0.27	0.19	0.26	0.20	0.16	0.22	0.28
Na^{+}	0.49	0.60	0.45	0.45	0.37	0.46	0.40	0.46
K^{+}	0.68	0.67	0.65	0.57	0.50	0.46	0.24	0.56
P^{5+}	0.031	0.042	0.038	0.017	0.018	0.013	0.053	0.041
Fe^{2+}/Fe^{3+}	0.65	0.80	0.66	0.34	0.30	0.55	0.48	0.65
Rb	36	26	43	19	20	5	8	27
Ba	230	238	139	118	82	122	102	164
Sr	43	43	31	32	23	54	44	41
Zr	157	136	106	87	77	77	91	121

Rb, Ba, Sr, Zr as p.p.m. (weight), all other elements as 1-cation percentages, G13, G221, G274 and G283 excluded from Ca , Na , K , Rb, Ba and Sr calculations; G103

and σ is the standard deviation of the population from which the samples were drawn

s_e was taken to be 0.07 $\bar{x}$ for Rb and Zr, 0.08 $\bar{x}$ for Sr and 0.05 $\bar{x}$ for Ba for each group. These values are those found by Taylor and Kolbe (1964) for optical spectrographic methods but Figs. 3 - 6 show that the X-ray fluorescence methods used in the present study are only slightly more precise.

The statistical significance of the difference in group means was tested by the analysis of variance technique (Moroney 1956 pp. 371-8 and Hoel 1947 pp. 154-8). The values of Snedecor's variance ratio F, for which tables are available (Yule and Kendall 1950, Appendix Table 5, pp. 668-670), were calculated using the formula:

$$F = \frac{\text{Between groups mean square}}{\text{Within groups mean square}} = \frac{\dfrac{\sum\left[n_i(\bar{x}_i)^2\right] - n(\bar{\bar{x}})^2 - k s_e^2}{k - 1}}{\dfrac{\sum(x^2) - \sum_1^k\left[n_i(\bar{x}_i)^2\right] - (n-k)s_e^2}{n - k}}$$

where n = the total number of granitic gneiss samples

k = the number of groups

n_i = the number of samples in each group (if this varies from group to group $n_i(\bar{x}_i)^2$ must be calculated separately for each group before summation).

$\bar{x}_i$ = the arithmetic mean of the values of x for each group.

$\bar{\bar{x}}$ = the arithmetic mean of all the values of x for the whole suite of samples.

and/

and s_e = the standard deviation which would be found in a series of repeat analyses of the same sample.

The values of F are listed in Table 15 and the level of significance given refers to the probability that the differences in mean composition could have arisen by chance in a homogeneous population having the degree of variability found in the analysed granitic gneisses.

The subtraction of part of the sum of squares due to experimental error from the between groups sum of squares as has been done above may not be statistically correct, but even if the whole correction were to be subtracted from the within groups sum of squares, the significance levels attained by the values of F for the various elements wold not be changed. It is logical that the inclusion of the experimental correction should increase the statistical significance of the differences between group means, because the effect of experimental error is to tend to conceal any significant geographical trend of variation beneath a blanket of random variation.

The differences in standard deviation from group to group were tested for statistical significance using a test described by Hoel (1947, pp. 208-212). This test involves the calculation of a "likelihood ratio" μ according to the formula.

$$\mu = \frac{\sigma_1^{(n_1-1)} \cdot \sigma_2^{(n_2-1)} \cdots\cdots \sigma_k^{(n_k-1)}}{\left[\frac{(n_1-1)\sigma_1^2 + (n_2-1)\sigma_2^2 + \ldots + (n_k-1)\sigma_k^2}{n-k}\right]^{\frac{n-k}{2}}}$$

where/

TABLE 15.

Analysis of variance for granitic gneiss analyses.

	Total sum of squares	Between Groups sum of squares	Within groups sum of squares	Between Groups. Variance	D.F.	Within Groups. Variance	D.F.	F	Significance Level.
Si^{4+}	288.5733	123.3286	165.2477	20.5552	6	1.6525	100	12.439	0.1%
Ti^{4+}	0.7288	0.1017	0.6271	0.0170	6	0.00627	100	2.711	5 %
Al^{3+}	54.5406	20.5266	34.0140	3.4212	6	0.3401	100	10.059	0.1%
Fe^{3+}	6.8833	0.6763	6.2070	0.1127	6	0.0621	100	1.815	not significant.
Fe^{2+}	48.2123	19.7405	28.4718	3.2901	6	0.2847	100	11.556	0.1%
Mn^{2+}	0.0244	0.0081	0.0163	0.00135	6	0.000163	100	8.282	0.1%
Mg^{2+}	3.9807	1.1319	2.8488	0.1887	6	0.0285	100	6.621	0.1%
Ca^{2+}	8.0095	1.0136	6.9959	0.1689	6	0.0729	96	2.317	5 %
Na^{+}	24.7645	3.4680	21.2965	0.5780	6	0.2218	96	2.606	5 %
K^{+}	32.8422	1.6800	31.1622	0.2800	6	0.3246	96	0.863	not significant.
P^{5+}	0.1688	0.0574	0.1114	0.00957	6	0.00111	100	8.622	0.1%
$(Fe^{2+}+Fe^{3+})$	44.2563	12.1420	32.1143	2.0237	6	0.3211	100	6.302	0.1%
Rb	71511	4593	66918	765.5	6	704.4	95	1.087	not signif
Ba	2762105	340142	2412963	56690	6	25494	95	2.224	5 %
Sr	174672	30008	144664	5001	6	1523	95	3.284	1 %
Zr	1550606	362548	1188058	60425	6	11881	100	5.086	0.1%

D.F. = Degrees of freedom.
G13, G221, G274 and G283 excluded from Ca^{2+}, Na^{+}, K^{+}, Rb, Ba and Sr calculations.
G103 excluded from Rb, Ba and Sr calculations.
All values in this table include experimental error correction.
Rb, Ba, Sr, Zr in ppm. by weight.
All other elements in 1-cation percentages.

where $n_1, n_2, \ldots n_k$ are the numbers of samples in groups 1, 2 .... k.

$\sigma_1, \sigma_2, \ldots \sigma_k$ are the standard deviations of the populations from which the analysed samples of groups 1, 2, ..., k were drawn, calculated as in Table 14.

and the other symbols have the same meanings as in the formula for F.

The quantity $q = \dfrac{-2 \log_e \mu}{1 + \frac{1}{3(k-1)} \cdot \sum_1^k \left(\frac{1}{n_i} - \frac{1}{n}\right)}$ then has a distribution very close to that of χ^2 with (k - 1) degrees of freedom (Hoel 1947 pp. 134 and 246). This test is applicable whether the individual group means are the same or not, but it does assume that the populations forming the individual groups are normally distributed about their respective means. With the size of group used in the present study, this assumption is difficult to prove, but appears to be approached fairly closely for most of the elements studied (but see the comments on Mg, K and P on pp. 83, 85). The results of the test of homogeneity of variances (variance = (standard deviation)^2) are listed in Table 16.

It should be remembered that the two tests described above do not take into account the fact that the areas represented by Groups I - VII form a S - N progression along the granitic gneiss outcrop, and that a gradual S - N increase in a quantity may be quite/

TABLE 16.

Test of homogeneity of variances for granitic gneisses.

	Including experimental error correction.			Uncorrected for experimental error.		
	$\log_{10}\mu$	q	Level of significance	$\log_{10}\mu$	q	Level of significance
Si^{4+}	-7.4259	36.47	0.1%	-6.4684	29.15	0.1%
Ti^{4+}	-5.4046	24.36	0.1%	-5.1131	23.04	0.1%
Al^{3+}	-8.3223	37.51	0.1%	-6.4286	28.97	0.1%
Fe^{3+}	-4.8213	21.73	1 %	-4.4572	20.09	1 %
Fe^{2+}	-2.1849	9.85	Not significant.	-2.1799	9.82	Not significant.
Mn^{2+}	-4.7870	21.58	1 %	-4.1850	18.86	1 %
Mg^{2+}	-6.4950	29.27	0.1%	-5.5433	24.98	0.1%
Ca^{2+}	-4.6379	20.88	1 %	-4.4492	20.03	1 %
Na^{+}	-0.9466	4.26	Not significant.	-0.8661	3.90	Not significant.
K^{+}	-3.2985	14.85	5 %	-3.1454	14.16	5 %
P^{5+}	-8.8234	39.77	0.1%	-8.0507	36.28	0.1%
$(Fe^{2+}+Fe^{3+})$	-4.5165	20.36	1 %	-4.4739	20.16	1 %
Rb	-14.0670	63.32	0.1%	-6.3878	28.76	0.1%
Ba	-5.9612	26.83	0.1%	-5.0380	22.68	0.1%
Sr	-2.6243	11.81	Not significant.	-2.2094	9.94	Not significant.
Zr	-3.0703	13.84	5 %	-2.4570	11.07	Not significant.

$$q = \frac{-2.\log_e \mu}{1 + \frac{1}{3(k-1)} \sum_{1}^{k} \left(\frac{1}{n_i} - \frac{1}{n} \right)}$$

Rb, Ba, Sr and Zr in ppm. by weight.
All other elements in 1-cation percentages.
G13, G221, G274 and G283 excluded from Ca^{2+}, Na^{+}, K^{+}, Rb, Ba and Sr calculations.
G103 excluded from Rb, Ba, and Sr calculations.

quite important without being statistically significant according to the tests used. However, the high degree of statistical significance obtained for the variation in amount of certain elements increased the likelihood that trend surface analysis would produce significant results.

The results listed in Table 14 demonstrate that there is a real increase in Ti, Al, Fe^{2+}, Mn, Mg, Na, P, Ba, Sr and Zr and a decrease in Si and K in a northerly direction within the granitic gneiss, while Fe^{3+}, Ca, and Rb remain roughly constant. In addition to this trend, there is a remarkable northward increase in homogeneity (Table 16) for all elements except Na, Sr, and possibly Fe^{2+}. It is notable that the maximum homogeneity is usually observed in Group V or Group VI, while in some cases, for example P, Group VII is as inhomogeneous as Group I or Group II. This may be connected with the large amount of unmigmatised metasediment occurring within the granitic gneiss outcrop in the districts represented by Groups I and VII.

The similarity in the behaviour of P and Zr to that of Fe^{2+}, Mn, Mg, and Ti suggested that the proportions of apatite and zircon might be correlated with the amount of biotite in the rock. To test whether such a relation existed, the correlation coefficients for P and Mg, Zr and Mg, and P and Zr were calculated, Mg being chosen to represent biotite because

it/

it was unlikely to enter any of the other phases in the rock in significant quantities. These correlation coefficients were found to be +0.79, +0.73 and +0.63 respectively, which indicate an extremely significant positive correlation between the three variables. It was concluded that the amounts of biotite, apatite, and zircon in the granitic gneiss samples are closely related to each other.

The effect of the variation in chemical composition of the granitic gneisses on their mineralogical composition is shown in Table 17, where the average compositions of each group are converted into norms. Table 17b shows that only in Groups VI and VII is there excess Al over the amount required to convert all of the C.I.P.W. normative hypersthene into biotite, and comparison with Table 1 shows that the majority of the muscovite- and garnet-bearing granitic gneisses belong to Groups VI and VII. It is notable, however, that none of the granitic gneisses contain normative garnet (Table XV), and this, together with the rounded shape of the garnets in the granitic gneiss and the lack of any obvious relation between their inclusions and the foliation of the surrounding rocks makes it possible that the garnets may not be in equilibrium with their surroundings. The presence of normative hornblende in so many of the granitic gneisses of Groups I to V when so very few of them have modal hornblende may be due to a slight infelicity in the choice of normative biotite composition, or to a tendency to incomplete extraction in the chemical method for Al. It should be noted that a deficiency of 0.1% Al_2O_3 in the analysis will produce an extra 0.6% of hornblende in/

TABLE 17.

Average compositions of granitic gneiss expressed as Eskola molecular norms.

(a) Using C.I.P.W. normative minerals.

	Group I	Group II	Group III	Group IV	Group V	Group VI	Group VII	All granitic gneisses
Q	31.07	30.64	30.64	28.44	28.32	26.77	27.47	29.16
Or	28.85	27.95	28.30	26.85	27.35	27.15	27.60	27.75
Ab	28.10	28.75	27.70	29.80	30.40	29.15	28.20	28.90
An	6.80	6.30	5.80	7.30	6.25	6.25	6.85	6.50
C	-	0.47	0.80	0.46	0.55	1.63	1.22	0.70
Di	0.08	-	-	-	-	-	-	-
Hy	2.94	3.78	4.86	4.86	4.74	6.28	6.30	4.76
Mt	1.20	1.11	0.81	1.08	1.17	1.07	1.04	1.07
Il	0.74	0.76	0.82	0.86	0.84	0.92	0.92	0.82
Ap	0.23	0.28	0.30	0.35	0.38	0.40	0.38	0.33
$100 \times \frac{Fe^{2+}}{(Fe^{2+}+Mg)}$ in Hy	56	60	67	67	65	69	69	66

(b) Using high-grade metamorphic minerals.

	Group I	Group II	Group III	Group IV	Group V	Group VI	Group VII	All granitic gneisses
Q	32.38	32.57	33.22	30.86	30.73	30.16	30.87	31.64
Kfels.	33.32	29.84	28.22	27.46	27.92	23.68	24.73	27.94
Plag.	25.96	28.14	27.42	29.82	29.72	31.01	30.11	29.05
Musc.	-	-	-	-	-	0.91	0.08	-
Biot.	3.67	6.81	9.57	8.16	8.38	12.94	12.98	9.02
Horn,	3.17	1.25	0.57	2.37	1.78	-	-	1.00
Mt.	0.85	0.81	0.49	0.66	0.79	0.32	0.28	0.73
Il.	0.44	0.33	0.25	0.32	0.30	0.54	0.54	0.27
Ap.	0.23	0.28	0.30	0.35	0.38	0.40	0.38	0.33
% An in plag.	19	19	20	21	17	20	23	20

G13, G221, G274 and G283 not included in these calculations.

in the modified norm.

The samples G13, G221, G274 and G283, which are remarkable for their unusually high plagioclase content and low potash felspar content, come from the northern, otherwise more homogeneous part of the granitic gneiss outcrop. Apart from those caused by this characteristic, however, the geochemical features of these rocks are very similar to those of the other granitic gneisses. Even when these samples are excluded from consideration, the northward increase in homogeneity of the granitic gneisses in Na, Ca, and K is much less than their increase in homogeneity in most of the other major elements (Table 16). Hence when G13, G221, G274 and G283 are taken into consideration, the reality of this increase in homogeneity becomes very doubtful.

The ratios $(Fe^{2+} + Fe^{3+})/Fe^{2+} + Fe^{3+} + Mg)$ and K/Rb show no clear-cut N-S trend (Table 18), but K/Ba, Ca/Sr, and (Ca + Na + K)/Sr show a definite, though irregular, tendency to decrease inna northerly direction.

The pattern of variation found in the Ardgour granitic gneiss may be briefly compared with that found by Mercy (1960 a) in the Older Granodiorite of Donegal. The degree of variability from sub-area to sub-area was not studied but a definite trend of compositional variation was established. This consisted of an ESE-WNW increase in $(Na + K) / (Na + K + Fe^{2+} + Fe^{3+} + Mg)$, K/Ca and Q/(Q + F + P) (von Wolff symbols) and a decrease in Na/K, while $Mg/(Fe^{2+} + Fe^{3+})$ remained constant in an ESE-WNW direction but/

TABLE 18.

Variation in certain ratios within the granitic gneiss.

	Group I	Group II	Group III	Group IV	Group V	Group VI	Group VII	All granitic gneisses.
Na/K	0.974	1.027	0.979	1.110	1.112	1.085	1.022	1.041
$(Fe^{2+}+Fe^{3+})/(Mg+Fe^{2+}+Fe^{3+})$	0.781	0.775	0.774	0.791	0.787	0.780	0.784	0.783
$(Na+K)/(Na+K+Fe^{2+}+Fe^{3+}+Mg)$	0.788	0.770	0.754	0.742	0.745	0.715	0.712	0.745
K/Ca	3.80	3.91	4.23	3.22	3.70	3.65	3.45	3.70
Q/(Q+F+P)	0.303	0.300	0.302	0.278	0.276	0.262	0.269	0.285
K/Rb	234	210	207	207	209	222	223	216
K/Ba	51.6	46.8	51.1	45.2	46.2	41.3	40.7	46.1
Ca/Sr	76.1	68.1	67.8	76.1	70.0	55.8	62.7	68.2
(Ca+Na+K)/Sr	521	474	510	469	488	383	407	463

$(Na+K)/(Na+K+Fe^{2+}+Fe^{3+}+Mg)$, K/Ca, Na/K and $(Fe^{2+}+Fe^{3+})/(Mg+Fe^{2+}+Fe^{3+})$ calculated from 1-cation percentages.

K/Rb, K/Ba, Ca/Sr, and (Ca+Na+K)/Sr calculated from weight percentages.

Q/(Q+F+P) calculated from von Wolff values (as defined by Barth (1955)).

but increased in a NNE direction. There is a similarity between this trend and that established in the Ardgour granitic gneiss, with the southernmost part of the Ardgour mass corresponding to the westernmost part of the Donegal granodiorite, and in both cases the behaviour of the ratio $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ (or $Mg/(Fe^{2+} + Fe^{3+})$) is unrelated to the other aspects of the regional trend of composition variation.

However, the absolute values of $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ are widely different in the two bodies (0.41 - 0.54 in Donegal and 0.77 - 0.79 in Ardgour), and the Ardgour trend in respect of Na/K and K/Ca (Table 18) is much more erratic than the Donegal trend. Also, the total range of composition over which the Ardgour granitic gneiss varies is much less than the range of composition shown by the Donegal granodiorite (compare Tables 14, p.61, and 18 with Mercy, 1960 a, pp. 110 - 116). This is true, even though the values of the various ratios shown in Table 18 are obtained by averaging 13-16 analyses whereas the points on Mercy's diagrams (1960 a, Figs. 1 - 4) represent the averages of 4 or fewer analyses, because the area covered by the Older Granodiorite of Donegal, is considerably greater than that of the Ardgour granitic gneiss. Averaging of the composition of the Donegal granodiorite over areas of the size represented by the Ardgour groups would not appreciably lessen the apparent geochemical variation of the Donegal granodiorite.

The/

The petrogenetic significance of the trend of composition which is so marked in the Older Granodiorite of Donegal (Mercy, 1960a), and is much less prominent in the Ardgour granitic gneiss (Tables 14 and 18), is uncertain. The type of variation described in both cases could have been produced in at least three ways:

(i) fractional crystallisation of a granodioritic or tonalitic liquid by the early crystallisation of ferromagnesian minerals and relatively calcic plagioclase in large quantities causing an enrichment in quartz and potash felspar in the later stages.

(ii) addition to, and/or reaction with a granitic liquid of varying proportions of pelitic metasediments, followed by fairly complete homogenisation.

(iii) partial melting of pelitic or greywacké-type metasediments along a rising scale of temperature. As shown by Winkler and von Platen (1961) this would produce a granitic liquid where the temperature was lowest and a tonalitic liquid where it was highest.

The field and petrographic studies of Pitcher (1952) indicate that the Older Granodiorite of Donegal formed either by partial mobilisation of country rock granitised in place or by piecemeal stoping and marginal assimilation of pelitic country rock by granitic magma. Mercy (1960 a) inclined to the former view, but considers the evidence insufficient to justify speculation as to the mechanism of this granitisation.

Thus/

Thus the variation in composition from place to place of a granitic body is very difficult to use as an indicator of its origin. However, the very small amount of regional variation in composition in the Ardgour granitic gneiss compared with the Older Granodiorite of Donegal suggests that the rôle of a liquid phase, if present, in the formation of the Ardgour granitic gneiss may have been different from that in the formation of the Donegal granodiorite.

(b) Trend surface analysis. The variation in a geological quantity from place to place may be regarded as consisting of a regional trend and a residual component measuring local deviations from this trend. The regional trend may be expressed as a function of two or three variables representing geographical position, for example, easting, northing, and altitude (often represented by x, y, and w respectively). The regional trend has so far always been expressed as a polynomial (usually of a degree not higher than the third) in x and y or x, y, and w.

If the geological quantity studied is z, and has been determined for n sample localities, the trend surface is the function of x, y, and w $[f(x, y, w)]$ for which the quantity

$$\sum_{i=1}^{i=n} [f(x_i, y_i, w_i) - z_i]^2$$

is a minimum. (x_i, y_i, w_i, z_i are the values of x, y, w, z at the ith sample locality). Thus the calculation of f(x, y, w) is based on the same idea as the calculation of regression equations (compare Krumbein/

Krumbein, 1959 with Hoel, 1947, pp. 89 - 90), but the algebra required is more complex, involving matrices, and the arithmetic considerably lengthier.

The significance of the trend surface is usually assessed by calculating the percentage of the total sum of squares accounted for by the trend surface. This is given by the expression

$$\frac{\sum_{i=1}^{i=n}\left[f(x_i, y_i, w_i) - \bar{z}\right]^2}{\sum_{i=1}^{i=n}\left(z_i - \bar{z}\right)^2} \times 100$$

where $\bar{z}$ is the arithmetic mean of the values of z. It gives a good idea of the relative importance of the trend and the deviations from it, but to decide whether the trend is statistically significant or whether it could have arisen by chance from a completely random set of z values, the F test (see p.62) is used. If the number of terms in the polynomial (3 for a linear, 6 for a linear + quadratic, and 10 for a linear + quadratic + cubic trend surface) is denoted by t and the number of sample localities by n, the number of degrees of freedom associated with the portion of the sum of squares accounted for by the trend surface is (t-1) and the number associated with the portion not accounted for by the trend surface is (n-t).

The first attempts to apply trend surface analysis to geological problems were made as early as 1955 (Krumbein 1956, see especially the appendix), but they were restricted by the inadequacy of the mathematical treatment as then worked out to the case/

case where all the sample points lay at the corners of a rectangular grid. In these attempts, only stratigraphic variables, such as the thickness of a formation or the depth below sea-level at which a given horizon was encountered in boreholes, were treated.

When the technique of trend surface analysis was extended to non-orthogonal sample localities by Krumbein (1959), the application of this statistical technique to the study of the variation in composition of rocks became possible, though the amount of calculation required to work out the trend surface became so large as to be possible only with the aid of a computer.

Trend surface analysis has been applied to the study of the variation in composition within granitic masses by Whitten (1959, 1960, 1961 a, b, c, d, 1962 a, b, 1963 a) and Dawson and Whitten (1962). Apart from the study of the variation in Na and K in the Malsburg granite (Whitten 1962 b) and the variation in certain chemical ratios in the Mourne Mountains granite G2 (Whitten 1961 b), all of this work has been based on modal analyses obtained by point counting thin sections. The amount of error involved in point counting his thin sections was not calculated by Whitten, but if only one thin section of area 660 mm.^{2} was counted per sample and 2000 points were counted per thin section, as appears to have been the case in Whitten's (1959, pp. 836-838) study of part of the Older Granodiorite of Donegal near Bunbeg, the analytical error/

error reaches quite large values unless the rocks are fairly fine-grained (Chayes 1956, Fig. 10, p. 77). The large increase in the percentage of the sums of squares accounted for by the trend surfaces using the average of four modes per square mile instead of one single mode per square mile (Whitten 1961 a, pp. 1344 and 1351) may be partly due to the decrease in the effect of experimental error on the values for each grid square. The best summary of the application of trend surface analysis to the variation in composition within granitic bodies is given in Whitten (1963 a).

Chayes and Suzuki (1963) have expressed doubts as to the value of trend surface in many geological studies, but though most of their criticism of Whitten's (1962 a) attempt to find the average composition of a granitic body by integrating the trend surface equation with respect to x and y rather than by simply averaging the raw data is valid, there seems (Whitten 1963 b) to be no objection to the use of a trend surface to represent the broad regional variation in composition of a granitic body even if the percentage of the total sum of squares accounted for by the trend surface is low, provided that it is realised that the variation represented by the trend surface may not be the only or even the most important type of variation within the rock mass. The deviations from the trend surface may have a geological significance as in the case of the Thorr district (Whitten 1960), though the exact nature of that significance is often impossible

to determine objectively.

In the present study, trend surfaces were computed over the whole area of outcrop of the Ardgour granitic gneiss for Si, Al, (Fe^{2+} + Fe^{3+}), Mg, Ca, Na and K, all expressed in 1-cation percentages. An Atlas computer was used and the programme employed had already been worked out by Dr. M.R. Osborne (Computer Unit, Edinburgh University) in connection with the stratigraphical research of Dr. E.K. Walton (Geology Department, Edinburgh University).

The easting and northing (x and y in the present nomenclature) were calculated using the National Grid, with an arbitrary origin at NM/820610 (Fig. 8), and a scale of 1.000 unit = 10 km. Despite the fact that the National Grid had to be transferred by hand from the 1 inch 7th Series Ordnance Survey maps on to the 6-inch Ordnance Survey maps of the Ardgour area, the coordinates of the sample localities were estimated to the nearest 10 metres, though the probable error was 20 - 30 metres. The programme used did not allow for the sample localities to be described by a third coordinate representing altitude, but the variation in height above sea level was only 780 metres, compared with variations of 12,000 metres in easting and 21,000 metres in northing. The trend surfaces were calculated using the actual 1-cation percentage values obtained at each of the 107 sample localities, rather than averaging them in groups of/

TABLE 19.

Sample Localities.

x = Easting)
y = Northing) 1 unit = 10 km.

Origin = NM/820610

Sample	x	y	Sample	x	y	Sample	x	y	Sample	x	y
G11	0.726	0.974	G81	0.227	0.494	G151	0.647	0.103	G231	0.947	1.569
G12	0.756	0.968	G82	0.241	0.553	G152	0.673	0.103	G232	1.013	1.557
G13	0.788	0.956	G83	0.234	0.614	G161	0.637	0.032	G233	1.089	1.537
G21	0.686	0.911	G91	0.143	0.497	G162	0.651	0.016	G241	0.944	1.653
G22	0.716	0.898	G92	0.153	0.531	G163	0.670	0.014	G242	1.001	1.649
G23	0.738	0.890	G93	0.142	0.565	G171	0.744	1.056	G243	1.048	1.646
G31	0.648	0.823	G96	0.051	0.508	G172	0.772	1.045	G244	1.100	1.632
G32	0.687	0.801	G97	0.056	0.522	G173	0.803	1.033	G251	0.979	1.731
G33	0.724	0.804	G98	0.100	0.500	G181	0.732	1.134	G252	1.031	1.738
G41	0.595	0.779	G101	0.477	0.409	G182	0.750	1.154	G253	1.143	1.728
G42	0.637	0.737	G102	0.505	0.465	G183	0.782	1.161	G254	1.181	1.727
G43	0.667	0.714	G103	0.562	0.517	G184	0.821	1.138	G261	0.962	1.804
G51	0.506	0.723	G105	0.627	0.563	G185	0.848	1.105	G262	1.044	1.819
G52	0.524	0.693	G111	0.550	0.378	G186	0.752	1.231	G263	1.121	1.860
G53	0.542	0.651	G112	0.596	0.433	G187	0.796	1.194	G271	0.940	1.889
G54	0.579	0.667	GP112	0.706	0.466	G191	0.828	1.255	G271A	0.974	1.893
G55	0.626	0.636	G113	0.661	0.476	G192	0.866	1.220	G272	1.029	1.933
G56	0.525	0.585	G116	0.621	0.387	G193	0.902	1.176	G273	1.096	1.967
G61	0.380	0.443	G117	0.678	0.389	G201	0.882	1.331	G274	1.203	1.961
G62	0.404	0.515	G121	0.622	0.334	G202	0.926	1.306	G281	0.929	1.968
G63	0.428	0.584	G122	0.701	0.344	G203	0.963	1.278	G282	0.973	2.005
G64	0.438	0.636	G123	0.769	0.338	G211	0.916	1.403	G283	1.016	2.051
G65	0.426	0.709	G131	0.639	0.264	G212	0.943	1.370	G284	1.037	2.076
G71	0.315	0.480	G132	0.700	0.261	G213	0.983	1.360	G291	1.157	2.061
G72	0.345	0.545	G133	0.747	0.260	G221	0.937	1.483	G292	1.155	2.088
G73	0.373	0.616	G141	0.649	0.183	G222A	0.989	1.457	G293	1.184	2.066
G74	0.352	0.701	G142	0.687	0.181	G223	1.036	1.425			

TABLE 20.

Equations of trend surfaces.

Arithmetic mean of x = $\bar{x}$ = 0.7245. Let u = x - $\bar{x}$.

Arithmetic mean of y = $\bar{y}$ = 0.9912. Let v = y - $\bar{y}$.

Linear surfaces:-

Si^{4+} = 68.12 - 0.4002 u - 1.491 v.

Al^{3+} = 14.67 + 0.0135 u - 0.7001 v.

$(Fe^{2+}+Fe^{3+})$ = 2.970 - 0.1966 u + 0.7313 v.

Mg^{2+} = 0.8253 - 0.0674 u + 0.1876 v.

Ca^{2+} = 1.538 + 0.3870 u - 0.00788 v.

Na^{+} = 5.869 + 0.5385 u + 0.0717 v.

K^{+} = 5.415 - 0.2721 u - 0.3048 v.

Linear + quadratic surfaces:-

Si^{4+} = 67.80 - 1.109 u - 1.854 v - 6.561 u^2 + 7.018 uv - 0.2000 v^2.

Al^{3+} = 14.70 + 0.6821 u + 0.7009 v + 3.970 u^2 - 3.194 uv - 0.1938 v^2.

$(Fe^{2+}+Fe^{3+})$ = 3.118 - 0.7822 u + 1.094 v - 0.0956 u^2 - 1.719 uv + 0.2213 v^2.

Mg^{2+} = 0.839 - 0.0872 u + 0.2120 v + 0.0798 u^2 - 0.1831 uv - 0.0091 v^2.

Ca^{2+} = 1.579 + 0.4281 u + 2.477 v + 0.9798 u^2 - 2.086 uv + 0.4256 v^2.

Na^{+} = 5.962 + 1.087 u - 0.1160 v 1 0.0230 u^2 + 1.027 uv - 0.6389 v^2.

K^{+} = 5.370 + 0.2321 u - 0.4790 v + 1.227 u^2 - 0.1156 uv - 0.0961 v^2.

Linear + quadratic + cubic surfaces:-

Si^{4+} = 67.87 + 1.724 u - 3.404v - 17.55 u^2 + 1.417 uv - 1.428 v^2
- 28.82 u^3 + 21.59 u^2v - 6.035 uv^2 + 1.523 v^3.

Al^{3+} = 14.68 - 0.3504 u + 1.150 v + 8.819 u^2 - 7.480 uv + 1.183 v^2
- 0.0075 u^3 + 12.10 u^2v - 6.960 uv^2 + 0.6558 v^3.

$(Fe^{2+}+Fe^{3+})$ = 3.109 - 2.593 u + 2.032 v + 5.653 u^2 - 4.973 uv + 0.3978 v^2
+ 24.40 u^3 - 28.34 u^2v + 12.14 uv^2 - 1.932 v^3.

Mg^{2+} = 0.826 - 0.2813 u + 0.4184 v + 0.2370 u^2 + 0.4999 uv - 0.2333 v^2
+ 8.665 u^3 - 14.01 u^2v + 5.980 uv^2 - 0.8793 v^3.

Ca^{2+} = 1.585 + 0.2390 u - 0.0361 v + 4.352 u^2 - 4.610 uv + 0.5867 v^2
+ 6.365 u^3 - 6.638 u^2v + 2.105 uv^2 + 0.3411 v^3.

Na^{+} = 6.028 + 0.0832 u + 0.3138 v - 2.748 u^2 + 2.187 uv - 0.8903 v^2
- 6.730 u^3 + 7.269 u^2v + 0.5594 uv^2 - 0.7446 v^3.

K^{+} = 5.275 + 1.435 u - 0.7957 v + 0.2042 u^2 + 1.052 uv + 0.3664 v^2
- 8.967 u^3 + 15.33 u^2v - 10.88 uv^2 + 1.473 v^3.

TABLE 21.

Significance of trend surfaces.

	Total Sum of Sqq.	Sum of sqq. due to Linear surface.	Sum of sqq. due to Lin + quad. surface	Sum of sqq. due to Lin + quad + cub. surface	% sum of sqq. explained by linear surface.	% sum of sqq. explained by lin + quad. surface.	% sum of sqq. explained by linear + quad + cubic surface
Si^{4+}	298.91	99.498	111.27	119.00	33.29	37.22	39.81
Al^{3+}	57.753	18.369	20.412	21.178	31.82	35.36	36.68
$(Fe^{2+}+Fe^{3+})$	55.569	16.203	17.869	21.513	29.16	32.16	38.71
Mg^{2+}	6.2823	0.9971	1.0069	1.4059	15.87	16.07	22.38
Ca^{2+}	12.224	1.1322	1.7642	2.3598	9.26	14.43	19.31
Na^{+}	46.818	3.5756	5.2435	6.0962	7.64	11.58	13.02
K^{+}	84.467	6.2657	6.8166	8.8180	7.42	8.07	10.44

	linear				linear + quadratic.				lin.+ quad.+ cubic				quadratic only.			
	F	D.F. trend	D.F. residual	Significance level.	F	D.F. trend	D.F. residual	Significance level	F	D.F. trend	D.F. residual	Significance level	F	D.F. trend	D.F. residual	Significance level
Si^{4+}	25.9	2	104	0.1%	12.0	5	101	0.1%	7.13	9	97	0.1%	2.15	3	103	Not signif.
Al^{3+}	24.3	2	104	0.1%	11.0	5	101	0.1%	6.24	9	97	0.1%	1.88	3	103	Not signif.
$(Fe^{2+}+Fe^{3+})$	21.4	2	104	0.1%	9.58	5	101	0.1%	6.81	9	97	0.1%	1.52	3	103	Not signif.
Mg^{2+}	9.81	2	104	0.1%	3.87	5	101	1%	3.11	9	97	1%	0.21	3	103	Not signif.
Ca^{2+}	5.30	2	104	1%	3.41	5	101	1%	2.58	9	97	1%	2.07	3	103	Not signif.
Na^{+}	4.30	2	104	5%	2.56	5	101	5%	1.61	9	97	not signf.	1.53	3	103	Not signif.
K^{+}	4.17	2	104	5%	1.77	5	101	not signf.	1.26	9	97	not signf.	0.24	3	103	Not signif.

D.F. = degrees of freedom.

	cubic only. F	D.F. trend	D.F. residual	Significance level.
Si^{4+}	1.10	4	102	not significant
Al^{3+}	0.53	4	102	not significant
$(Fe^{2+}+Fe^{3+})$	2.73	4	102	5%
Mg^{2+}	2.07	4	102	not significant
Ca^{2+}	1.54	4	102	not significant
Na^{+}	0.56	4	102	not significant
K^{+}	0.67	4	102	not significant.

The F test tests the probability that the observed reduction in the sum of squares could have occurred if the values of Si^{4+}, Al^{3+} ..... K^{+} had been completely randomly distributed.

of 3 or 4 before computing. As shown by Whitten (1961 a, Fig. 6, p. 1341), averaging of results produces very little change in the actual surface computed, and though it increases the percentage sum of squares accounted for by the trend surface, it does not increase the statistical significance of the surface. The sample localities, expressed in terms of x and y are listed in Table 19. The linear, linear + quadratic, and linear + quadratic + cubic trend surfaces are depicted graphically in Figures 9, 10 and 11 respectively, and their equations given in Table 20. The percentages of the total sum of squares accounted for by the various trend surfaces, and their statistical significance are listed in Table 21. The sign of the deviations of the actual values from those predicted by the linear + quadratic + cubic trend surfaces in different parts of the area are depicted in Figure 12.

The results shown in Table 21 indicate that the linear components of the calculated trend surfaces are by far the most important. The quadratic and cubic surfaces by themselves lack any statistical significance and when added to the linear componenet serve only to reduce the significance of the total trend surface. However, the (linear + quadratic + cubic) trend surface is still significant at the 0.1% level for Si, Al, and (Fe^{2+} + Fe^{3+}) and at the 1% level for Mg and Ca. For Na and K, only the linear components of the trend surfaces have any significance at all.

The/

FIGURES 9 - 12.

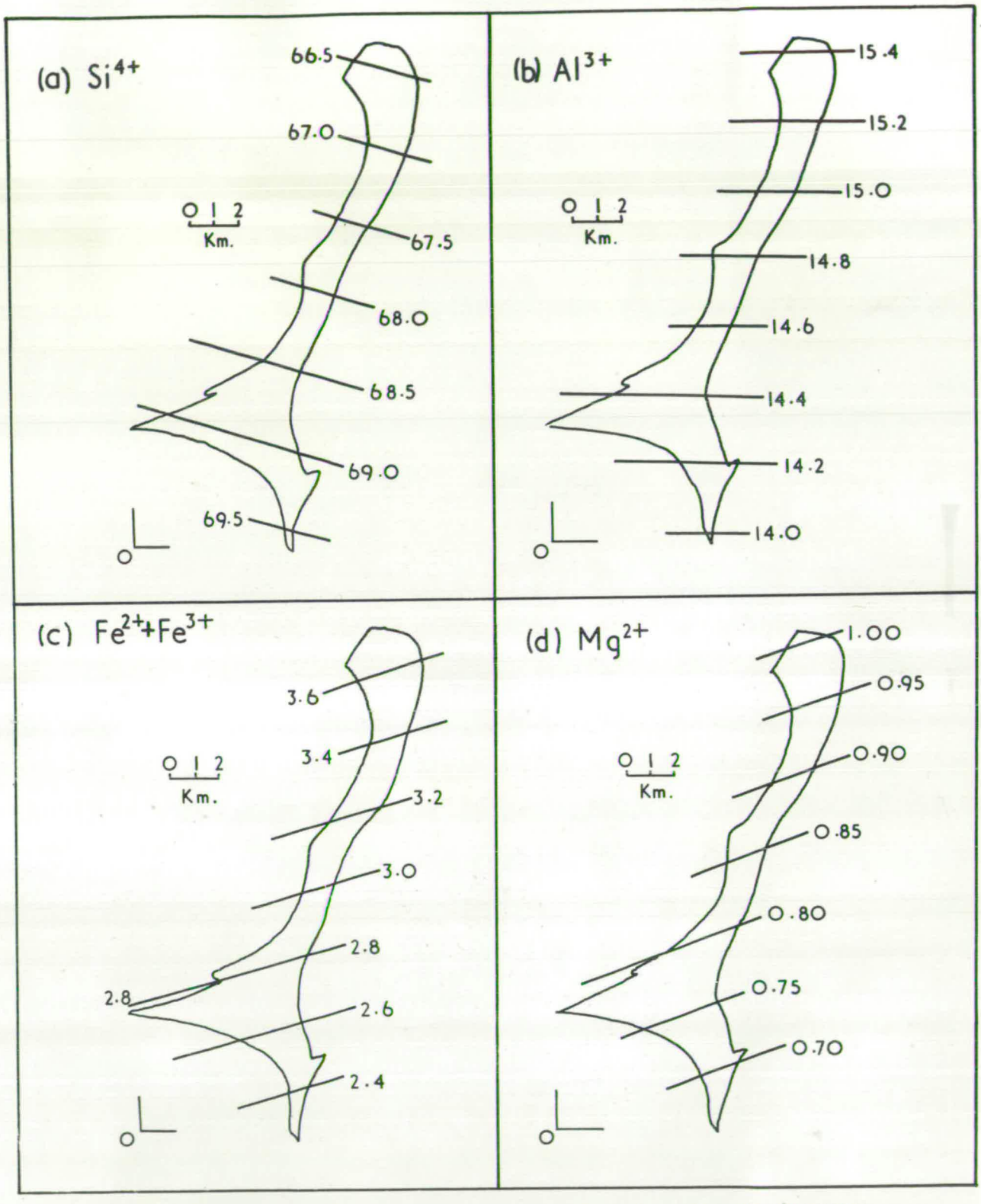
Figure 9 - Linear trend surfaces for compositional variation within Ardgour granitic gneiss.

Figure 10 - (Linear + quadratic) trend surfaces for compositional variation within Ardgour granitic gneiss.

Figure 11 - (Linear + quadratic + cubic) trend surfaces for compositional variation within Ardgour granitic gneiss.

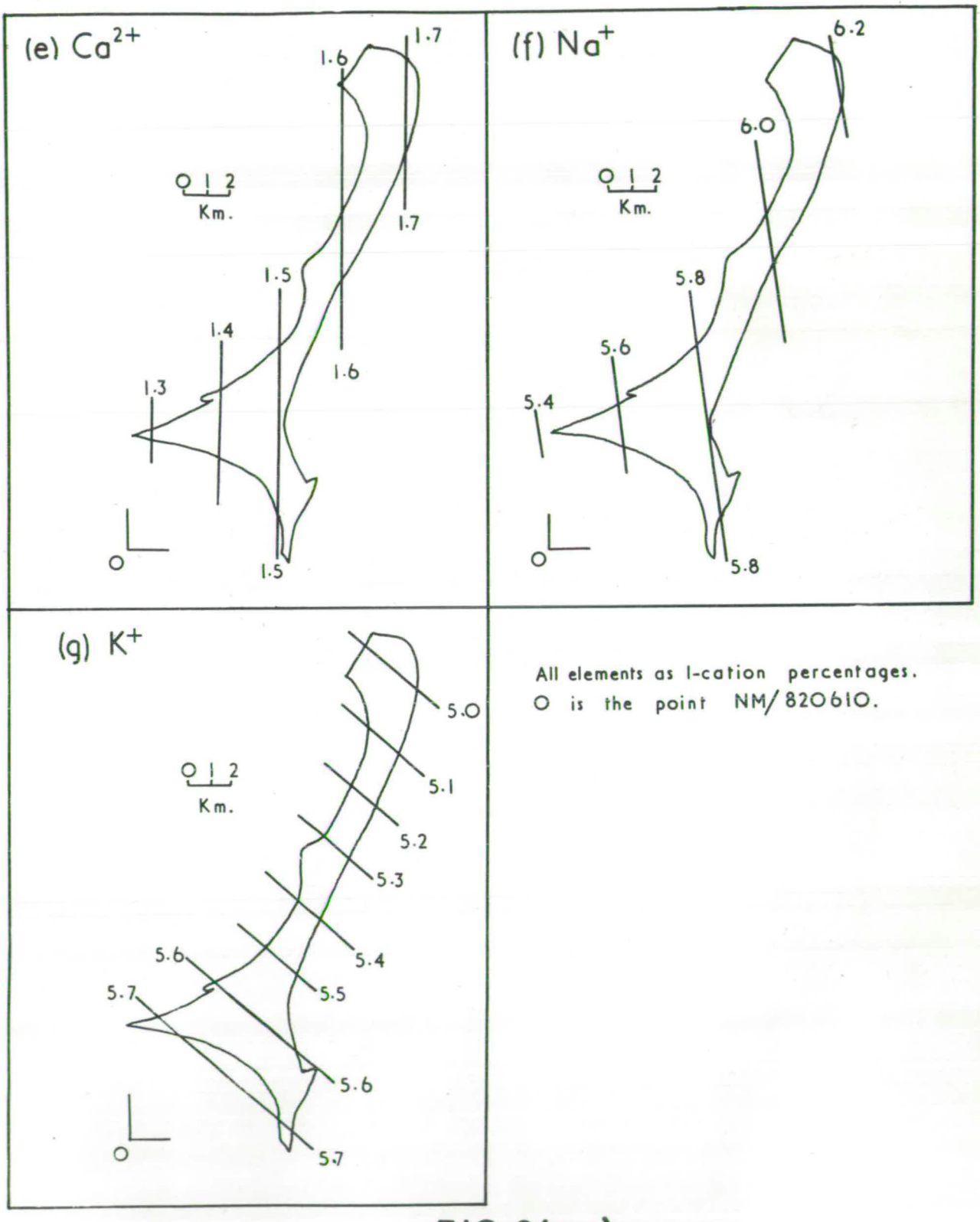
Figure 12 - Deviations from (linear + quadratic + cubic) trend surfaces, Ardgour granitic gneiss.

(a) Si^{4+}
(b) Al^{3+}
(c) $(Fe^{2+} + Fe^{3+})$
(d) Mg^{2+}
(e) Ca^{2+}
(f) Na^{+}
(g) K^{+}



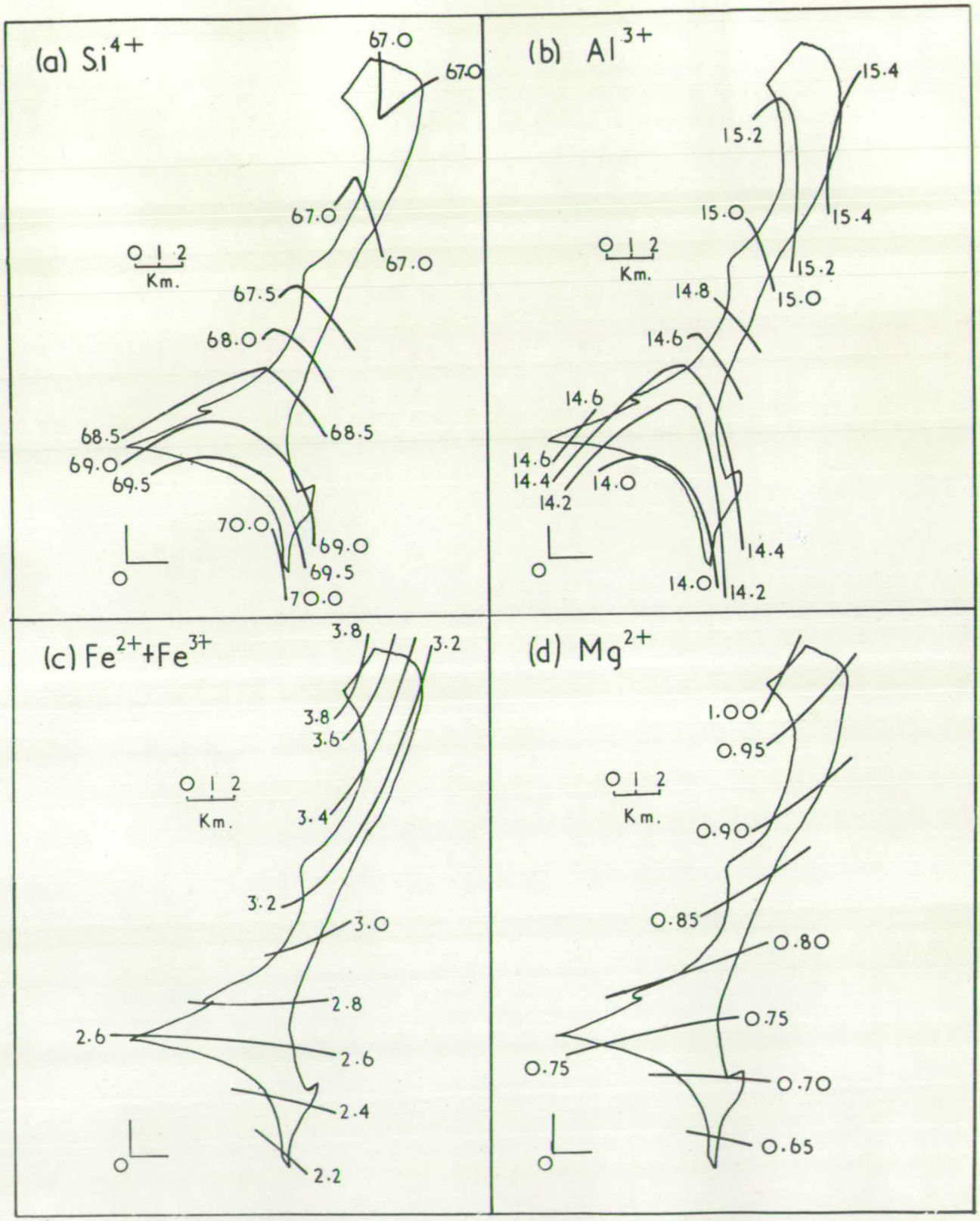
All elements as 1-cation percentages. O is the point NM/820610.

FIG. 9(a-d).



All elements as l-cation percentages.
O is the point NM/820610.

FIG. 9(e-g).



All elements as l-cation percentages. O is the point NM/820610.

FIG. 10 (a-d).

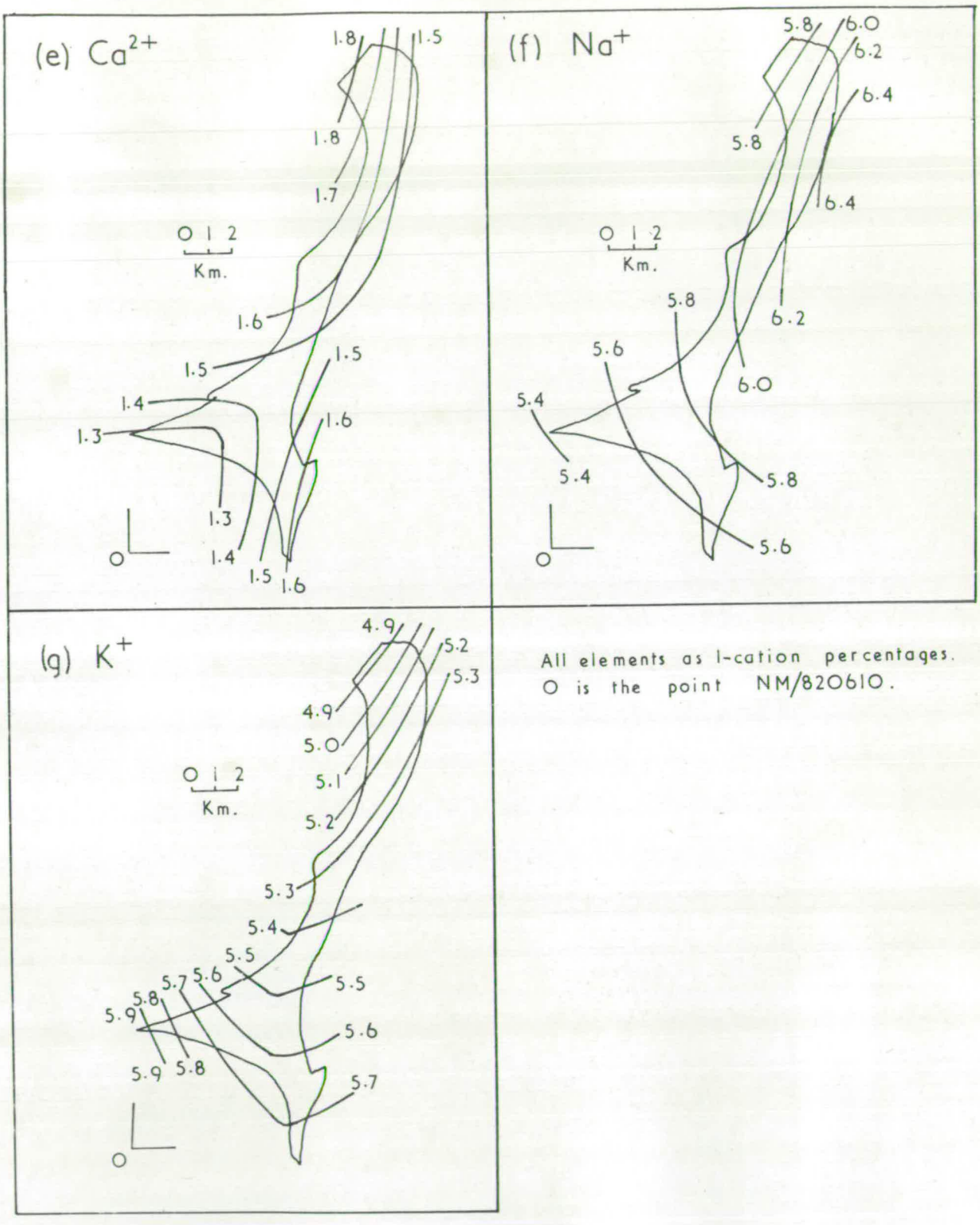
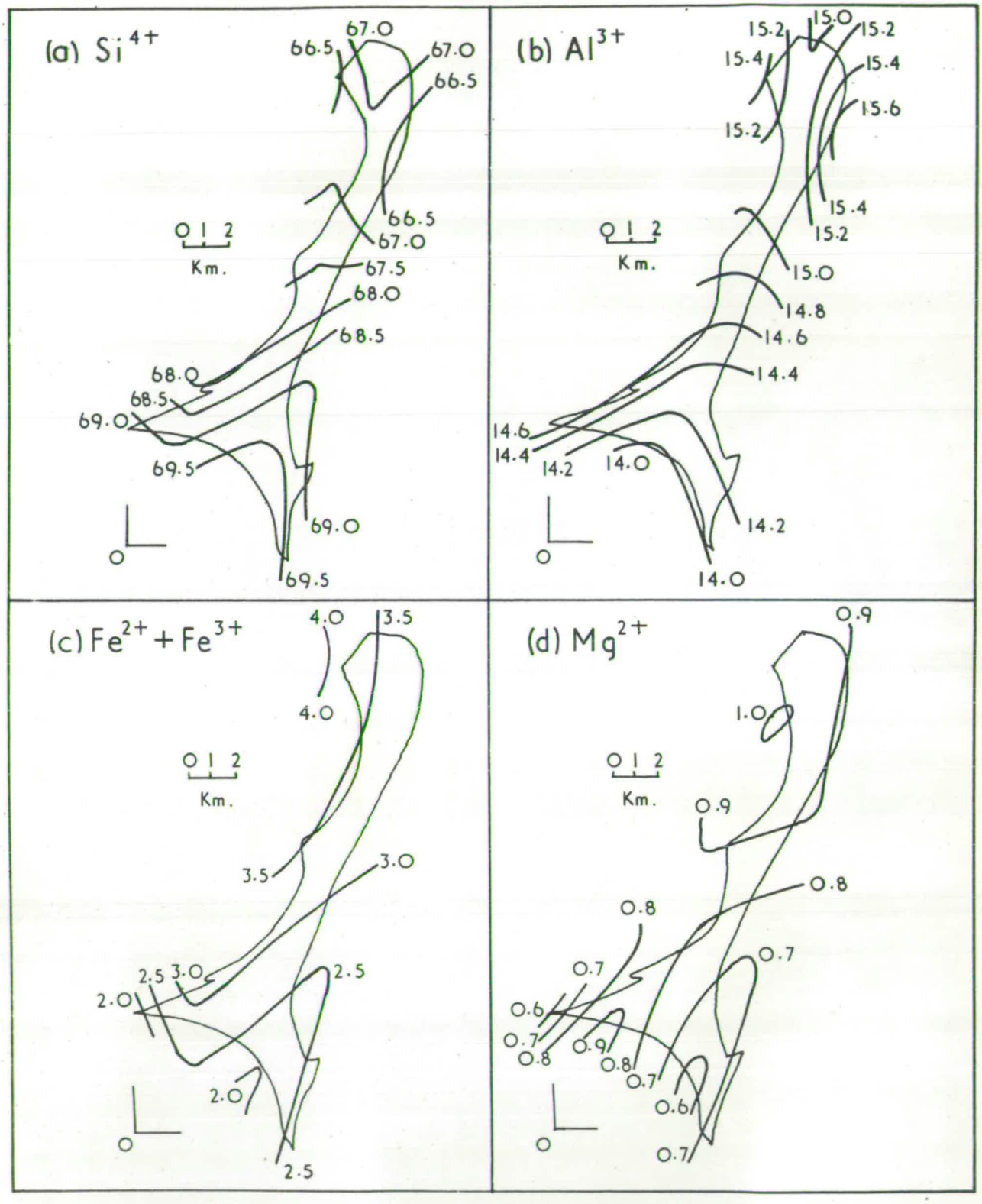

(e) Ca^{2+}
1.8
1.5
1.8
1.7
0 1 2
Km.
1.6
1.5
1.5
1.4
1.6
1.3
1.3
1.4
1.5
1.6
(f) Na^+
5.8
6.0
6.2
6.4
5.8
6.4
0 1 2
Km.
5.8
6.2
5.6
6.0
5.4
5.8
5.4
5.6
(g) K^+
4.9
5.2
5.3
4.9
5.0
0 1 2
Km.
5.1
5.2
5.3
5.4
5.5
5.7
5.6
5.5
5.8
5.9
5.6
5.9
5.8
5.7
All elements as l-cation percentages.
O is the point NM/820610.

FIG. 10 (e-g).



All elements as 1-cation percentages. O is the point NM/820610.

FIG. 11(a-d).

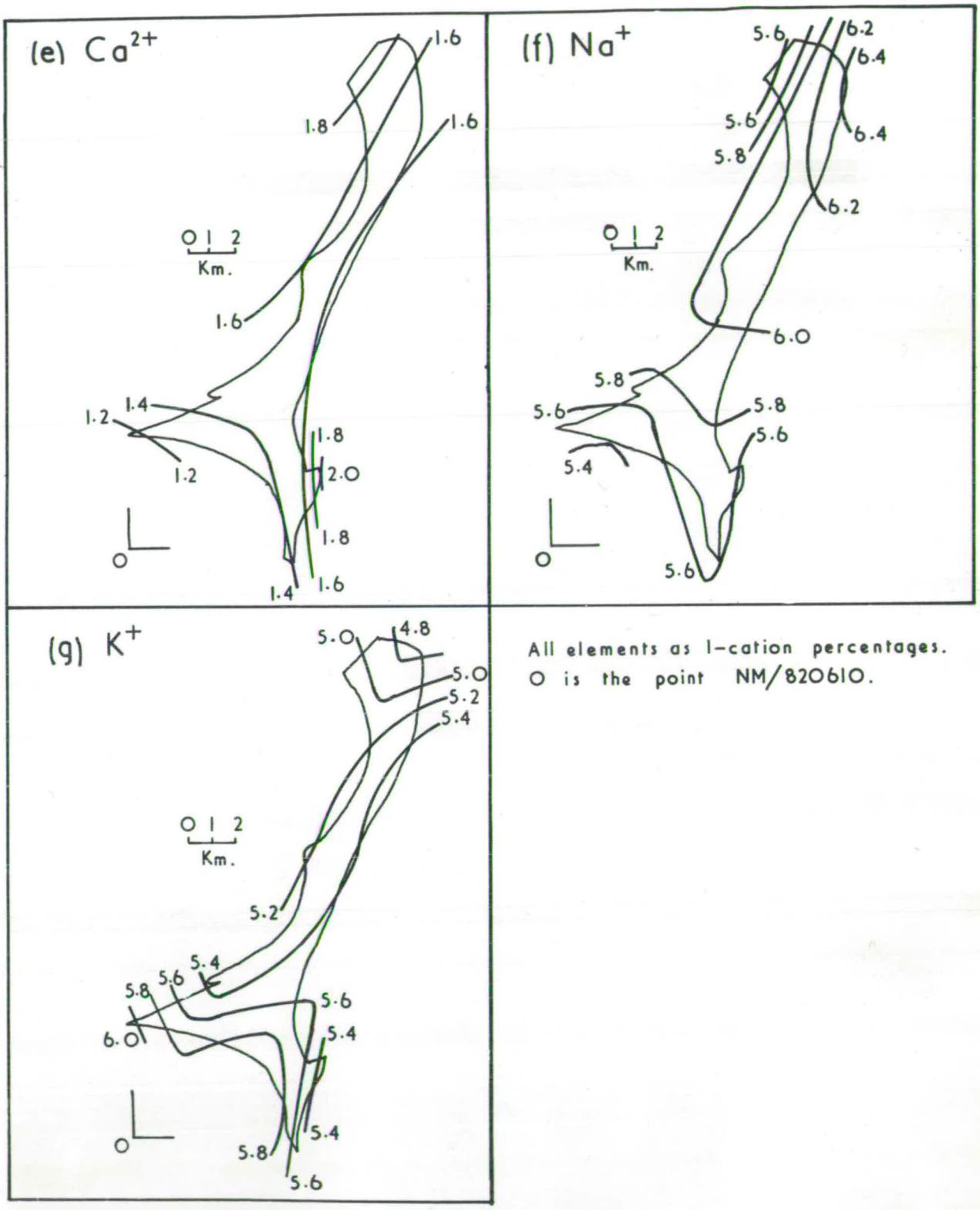
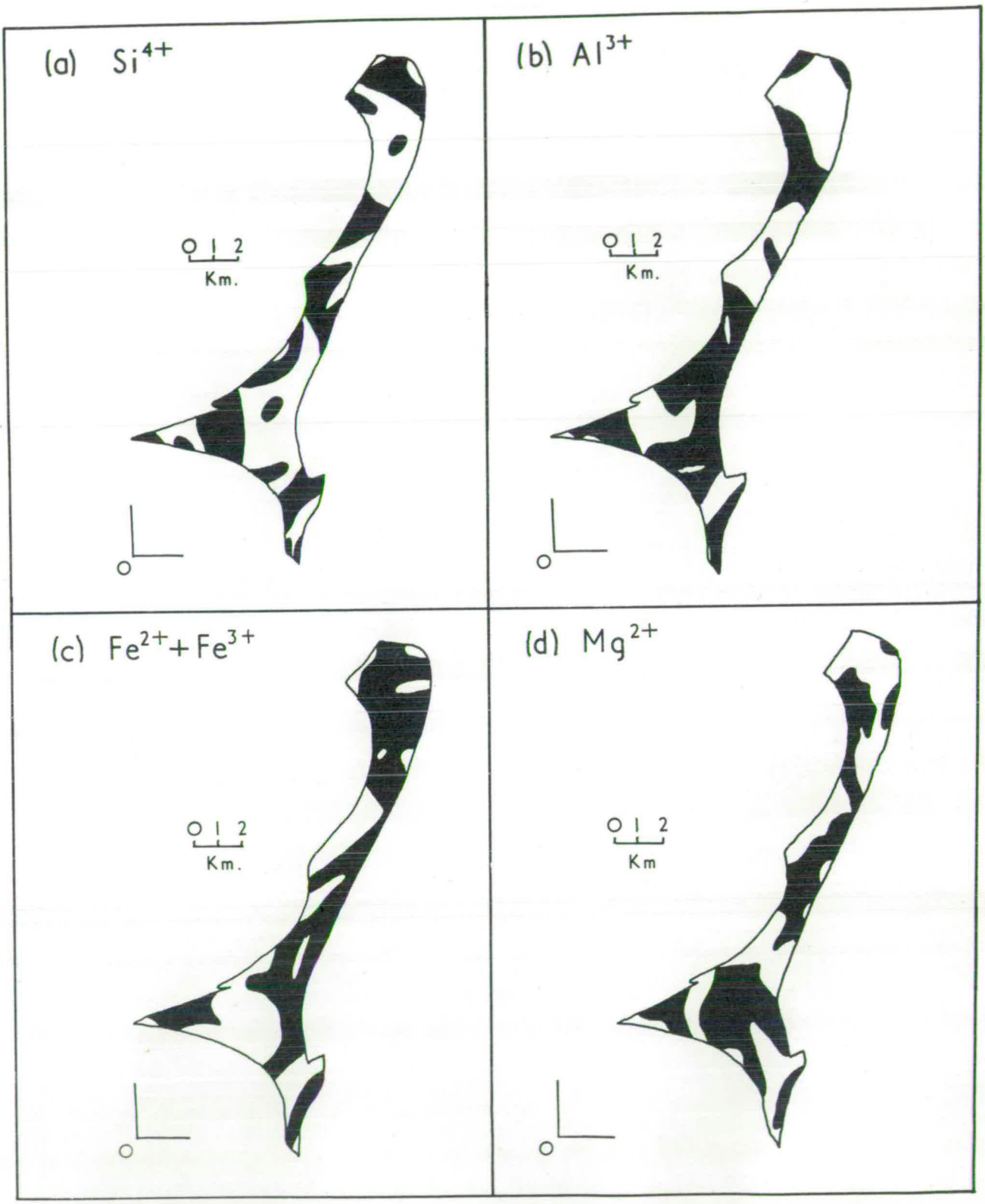

(e) Ca^{2+}
1.6
1.8
1.6
0 1 2
Km.
1.6
1.2
1.4
1.8
1.2
2.0
1.8
1.4
1.6
(f) Na^+
5.6
6.2
6.4
5.6
6.4
5.8
6.2
0 1 2
Km.
6.0
5.8
5.6
5.8
5.6
5.4
5.6
(g) K^+
5.0
4.8
5.0
5.2
5.4
0 1 2
Km.
5.2
5.4
5.8
5.6
5.6
6.0
5.4
5.4
5.8
5.6
All elements as 1-cation percentages.
O is the point NM/820610.

FIG. 11 (e-g).



Black—deviation positive. White—deviation negative. O is the point NM/820610.

FIG. 12 (a–d).

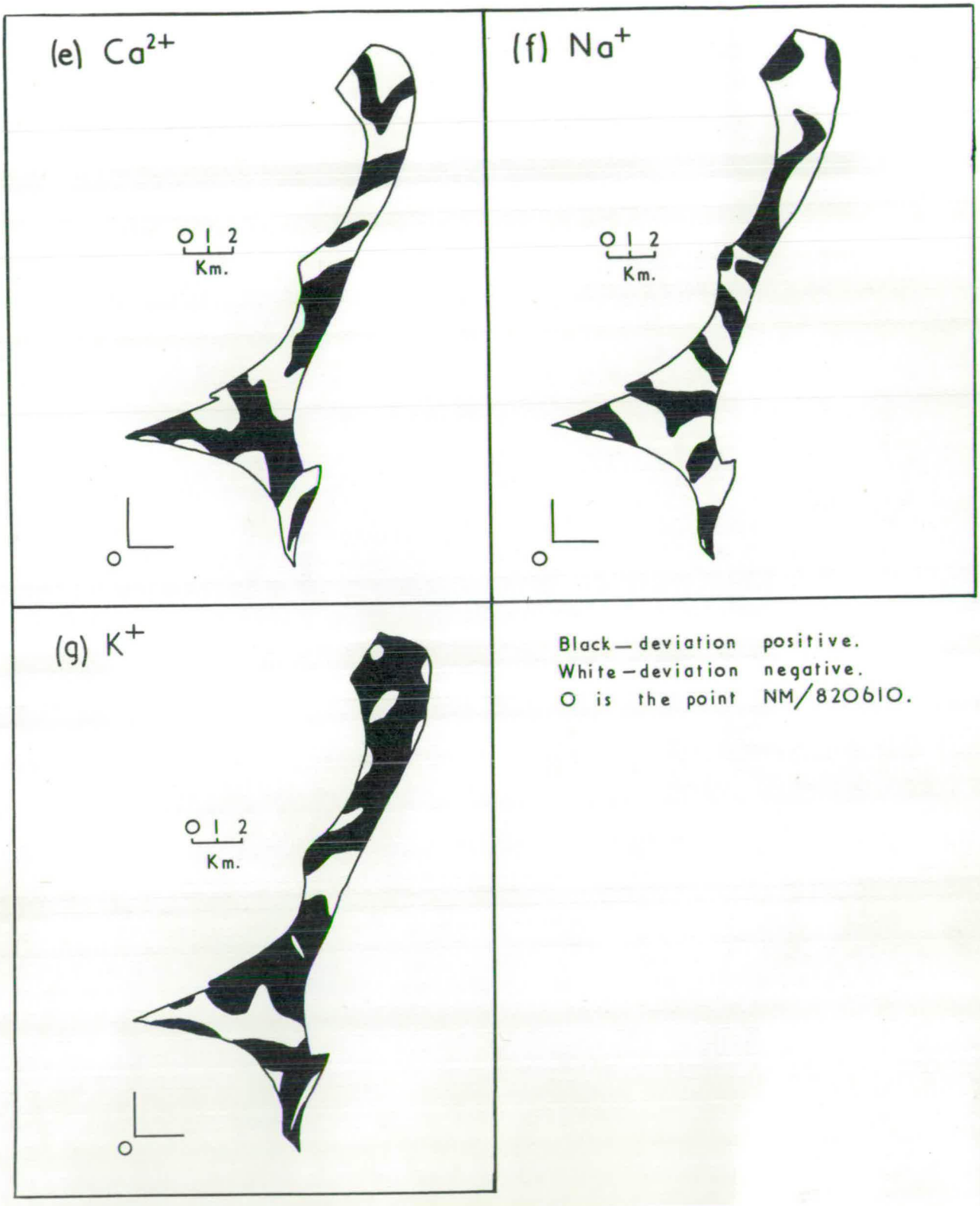

(e) Ca^{2+}
0 1 2
Km.
(f) Na^{+}
0 1 2
Km.
(g) K^{+}
0 1 2
Km.
Black — deviation positive.
White — deviation negative.
O is the point NM/820610.

FIG. 12 (e–g).

The linear trend surfaces for Si, Al, (Fe^{2+} + Fe^{3+}), and Mg have strikes ranging from ENE-WSW to ESE-WNW, whereas the linear trend surfaces for Ca, Na, and K have strikes between N-S and NW-SE. All of these trends except that for Ca produce considerable variations in the average values of the elements in the various groups studied in the previous section (compare Table 14). Despite the large differences in the strike of the linear trend surfaces, the (linear + quadratic + cubic) trend surfaces for (Fe^{2+} + Fe^{3+}) and K are almost exactly opposite in slope to each other everywhere, suggesting a strong degree of negative correlation between the two elements. This is quite probably fortuitous, and the chance that it is fortuitous is quite considerably increased by the fact that (Si + Al + (Fe^{2+} + Fe^{3+}) + Mg + Ca + Na + K) $\doteq$ 100% (Chayes 1960). There is a general similarity of the (linear + quadratic + cubic) surfaces for Si, Al, (Fe^{2+} + Fe^{2+}) and K^{+} though the amount and sense of variation differs. This reflects the northward increase in dark minerals and plagioclase and decrease in quartz and potash felspar. However, the relationship between the quantities of the various minerals and the amounts of the different elements in the granitic gneiss are by no means straightforward; for example, K enters not only into potash felspar but also into micas, Na is an important constituent of potash felspar as well as plagioclase, and Al enters all the minerals found in the rock except quartz in large proportions.

The/

The deviations from the regional (linear + quadratic + cubic) trend surfaces shown in Figure 12 appear to be completely random. Despite the fact that they account for such a large proportion of the total sum of squares (Table 21) there seems to be no particular pattern in the deviations. This contrasts strongly with the situation found in the Thorr district of N.W. Donegal by Whitten (1960), who found that a magmatic granodiorite invading metasediments showed a pattern of variation which, when maps of the deviations from the regional trend were drawn, reflected the stratigraphy of metasediments which had been replaced in situ. It also differs from his findings (Whitten, 1962 b) in the case of the Malsburg granite (Schwarzwald, West Germany) where a strong NW-SE trend of the contours in the deviation map is noticeable, though no likely origin for this structure was suggested.

The work on the Malsburg granite can be more closely compared with the present work than can any of Whitten's other work, because chemical analyses are used instead of modes. The only other case where analyses were used as the basis of a trend surface analysis was that of the Mourne Mountains Granite G2 (Whitten, 1961 b), and in this case the number of sample localities was very small (only 21), and the variables studied were all ratios, of which only Na_2O/K_2O seemed likely to have any petrological significance. In the case of the Malsburg granite, only Na and K were determined, but the number of/

TABLE 22.

Comparison of the Ardgour granitic gneiss with the Malsburg granite.

	Na_2O			K_2O		
	Malsburg	Ardgour	Ardgour*	Malsburg	Ardgour	Ardgour*
Arithmetic Mean	3.74	3.18	3.13	4.74	4.47	4.58
Variance	0.07	0.14	0.09	0.10	0.56	0.23

* Excluding G13, G221, G274 and G283.

		Na_2O Malsburg	Na^+ Ardgour	K_2O Malsburg	K^+ Ardgour
Percentage sum of squares accounted for by trend surface	Linear surface significance level.	3.81 not significant	7.64 5%	4.39 not significant	7.42 5%
	Linear + quadratic	24.88	11.58	21.24	8.07
	significance level	0.1%	5%	0.1%	not significant.
	Linear + quad. + cubic	27.93	13.02	31.87	10.44
	significance level	0.1%	not significant	0.1%	not significant.
	Quad. only	21.06	3.94	16.85	0.65
	significance level	0.1%	not significant	0.1%	not significant.
	Cubic only	3.06	1.44	10.63	2.37
	significance level	not significant	not signif.	1%	not signif.

of analysed samples was similar to that involved in the present study (110 compared with 107) and hence the significance level associated with a given percentage of the sum of squares is the same.

The arithmetic means and variances of the Na, and K results for Ardgour and Malsburg are compared in Table 22, which shows that although the variability of the Na_2O results is similar in the two cases, the K_2O results from Ardgour are much more variable than those from Malsburg, even if samples G13, G221, G274 and G283 are excluded.

The trend surfaces for the Malsburg granite are not dominated by their linear components to nearly such a degree as are the Ardgour trend surfaces, and, despite the considerable increase in the percentage of the sum of squares accounted for when the altitude of the sample localities is considered, the trend surfaces for Na and K in the Malsburg granite using only two coordinates for the sample localities achieve a very high level of significance, comparable to that attained by the trend surfaces for Si, Al, $(Fe^{2+} + Fe^{3+})$, and Mg in the Ardgour granitic gneiss.

3. ELEMENT DISTRIBUTIONS WITHIN THE GRANITIC GNEISS

In this section the elements will be considered individually in relation to the distribution of the experimentally determined/

determined values about the mean, without regard to the location of the samples, though the conclusions reached in Part 2 of this chapter will be taken into account. The distributions of the various elements are shown in Fig. 13, the class boundaries being taken at values $\frac{1}{8}$, $\frac{3}{8}$, $\frac{5}{8}$, $\frac{7}{8}$, $\frac{9}{8}$ ..... standard deviations removed from the mean in either direction. Table 23 gives the arithmetic mean, relative deviation (standard deviation/ mean), and relative skewness $\left(\frac{\Sigma(x-\bar{x})^3}{n} \Big/ \left[\frac{\Sigma(x-\bar{x})^2}{n}\right]^{3/2} \right)$ (Hoel 1947, p. 15) of the distributions, the latter being calculated roughly by replacing each of the values in each class of the histogram by the central value of that class. Histograms such as those of Figure 13 have previously been drawn only for analyses of one particular rock type either chosen randomly to include rocks of many ages and belonging to many different suites, or at best, chosen from rocks from different intrusions belonging to the same province or suite. On the basis of such studies of trace and minor elements Ahrens (1954) proposed that in a given rock-type the distribution of any minor or trace element was lognormal, that is, that the logarithms of the concentrations were distributed normally. Later (Ahrens 1963 a), he extended this hypothesis to all of the major elements in granitic rocks except Na, K, and Si, it being realised that all of the major elements of a rock could not have positively skewed distributions, since they must add up to 100%.

Chayes (1954) pointed out, however, that many other transformations/

TABLE 23.

Dispersion and skewness of element distributions (see p. 79).

Element	$\bar{x}$	C	a_3
Si^{4+}	68.12	2.4%	+0.79
Ti^{4+}	0.41	20.5%	-0.66
Al^{3+}	14.67	5.0%	-0.67
Fe^{3+}	0.70	36.7%	+0.65
Fe^{2+}	2.26	29.7%	-0.49
Mn^{2+}	0.059	27.1%	-0.24
Mg^{2+}	0.82	30.4%	+0.02
Ca^{2+} *	1.54	21.9%	+1.28
Na^{+} *	5.87	11.3%	+1.39
K^{+} *	5.42	16.4%	-2.31
P^{5+}	0.130	32.3%	-0.58
$(Fe^{2+}+Fe^{3+})$	2.96	21.7%	-0.87
Rb *	173	18.9%	-0.58
Ba §	805	24.2%	-0.42
Sr §	159	31.2%	+2.02
Zr	438	28.3%	-0.27
Ca^{2+} □	1.50	18.9%	+0.61
Na^{+} □	5.78	8.7%	-0.05
K^{+} □	5.55	10.1%	-0.10
Rb ○	176	16.6%	-0.04
Ba ○	825	20.4%	+0.13
Sr ○	155	27.9%	+0.88

* All samples.

§ Excluding G103.

□ Excluding G13, G221, G274 and G283.

○ Excluding G13, G103, G221, G274 and G283.

$\bar{x}$ = arithmetic mean.

C = $\bar{x}/s \times 100$ = percentage relative deviation, uncorrected for small sample effect or experimental error.

a_3 = relative skewness = $\frac{\sum(x-\bar{x})^3}{ns^3}$. (see p.79 and Hoel 1947 p.15).

Rb, Sr, Ba and Zr in ppm. by weight, all other elements in 1-cation percentages.

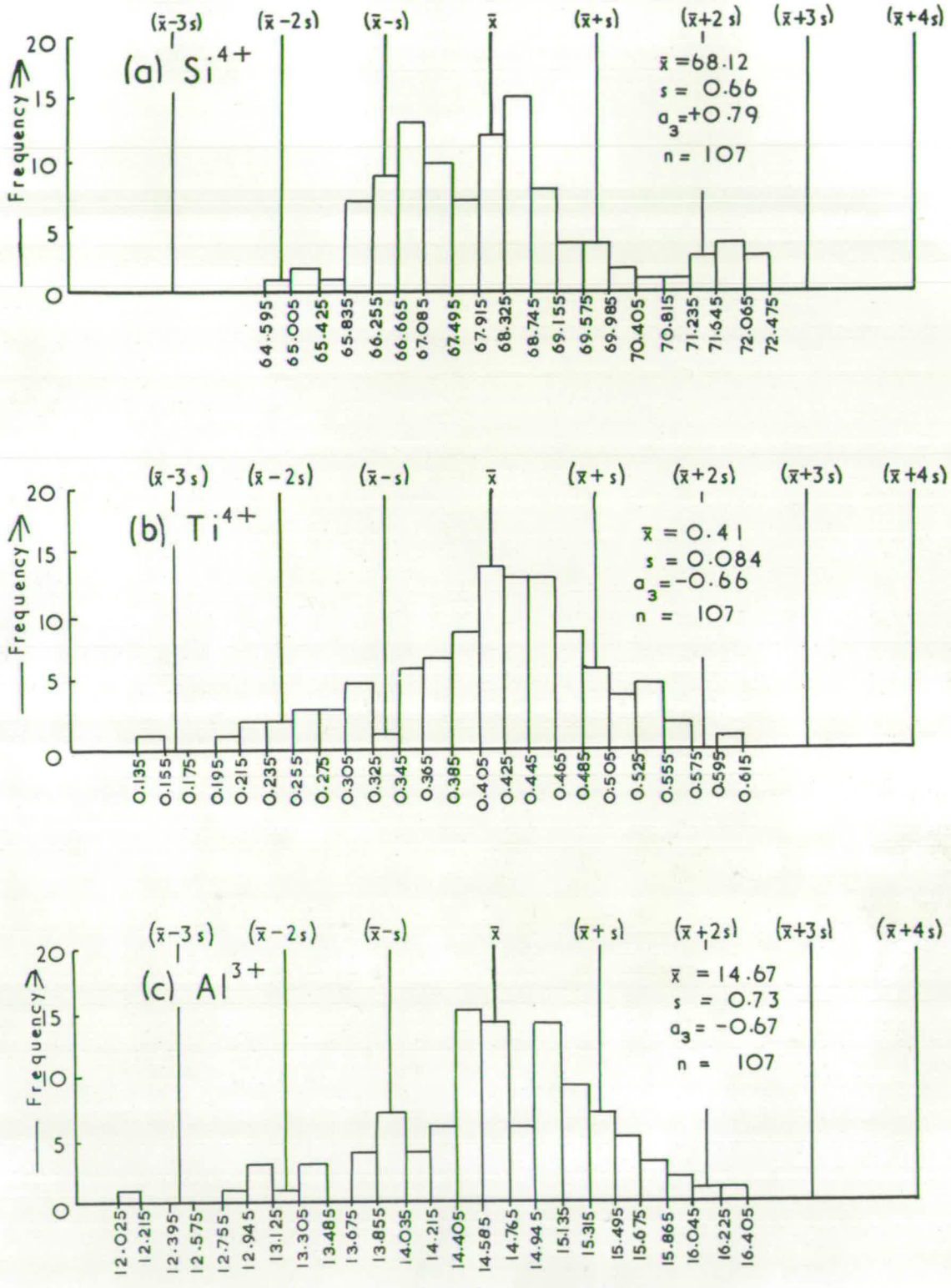
ations besides a logarithmic one would convert the distributions of Ahrens (1954) to near normality, and Shaw (1961) showed that if the relative deviation (corrected for experimental error) was less than 20 per cent, the difference between normal and lognormal distributions was so slight as to be undetectable in results of the type usually obtained. Shaw (1961) considered that a trace element camouflaged in a major rock-forming mineral or found only in an accessory mineral could be expected to have a lognormal distribution, but that no worthwhile prediction could be made about major element distributions. Butler, J. (1964) has also criticised the work of Ahrens on element distributions, and considers that most, if not all of the element distributions found by Ahrens (1954, 1963 a, b, c, 1964) can be explained by combining one of various types of magmatic differentiation trends with the effects of experimental error in increasing the positive skewness or decreasing the negative skewness of a distribution.

The element distributions within one body of rock could be expected to differ to some extent from those found by Ahrens, but it might have been expected that if that body of rock were of igneous origin, the distributions at least of the minor elements largely concentrated in accessory minerals would possess a high positive skewness. The relative deviations found for the elements studied in the Ardgour granitic gneiss are smaller than those found for most of the elements in the/

FIGURE 13.

Element distributions within Ardgour granitic gneiss.

(a) Si^{4+}
(b) Ti^{4+}
(c) Al^{3+}
(d) Fe^{3+}
(e) Fe^{2+}
(f) $(Fe^{2+} + Fe^{3+})$
(g) Mn^{2+}
(h) Mg^{2+}
(i) P^{5+}
(j) Ca^{2+}
(k) Na^{+}
(l) K^{+}
(m) Rb
(n) Ba
(o) Sr
(p) Zr

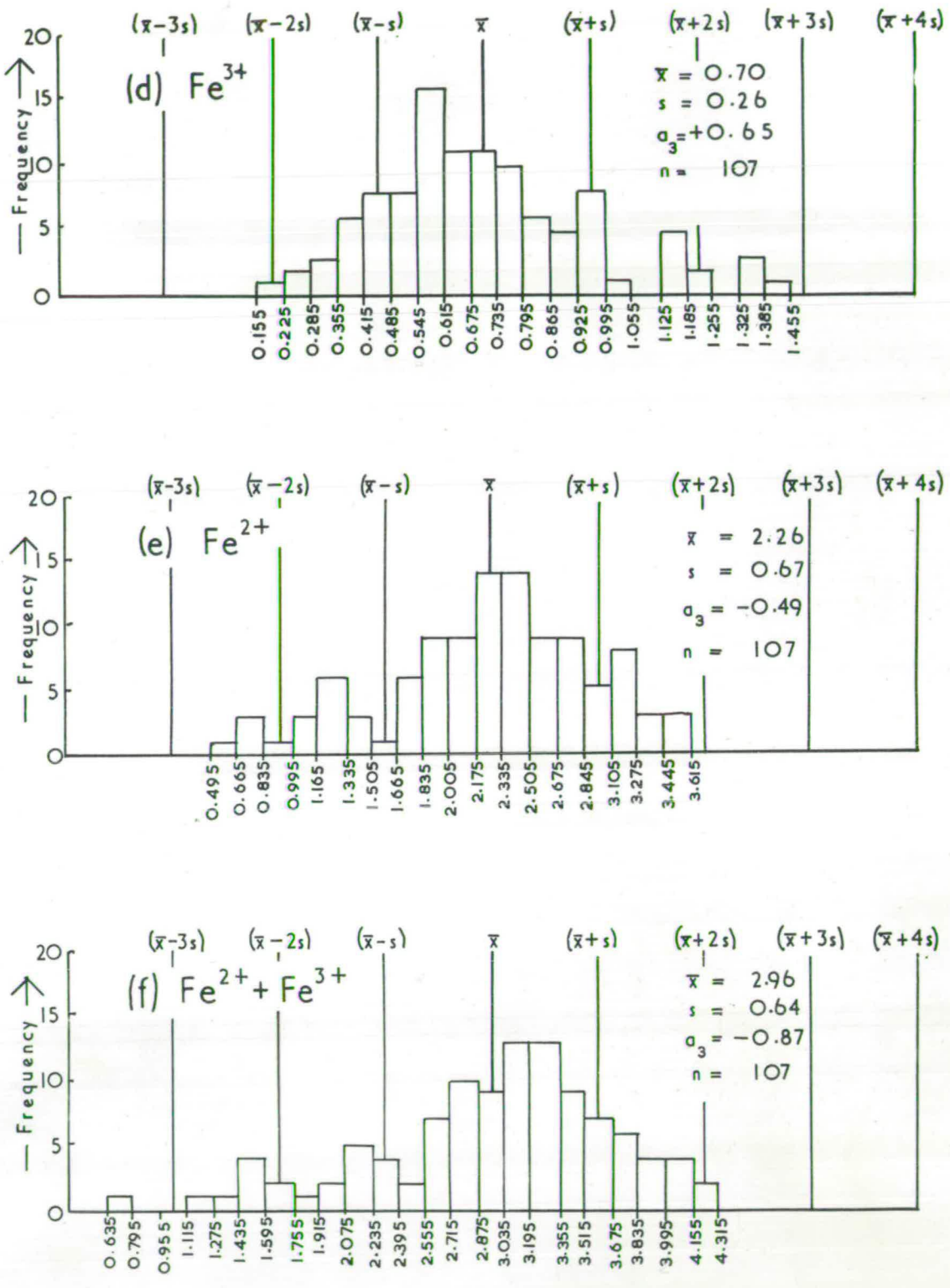


All elements as l-cation percentages.

$\bar{x}$ = arithmetic mean. s = standard deviation (uncorrected).

a_3 = relative skewness. n = number of samples.

FIG. 13 (a–c).

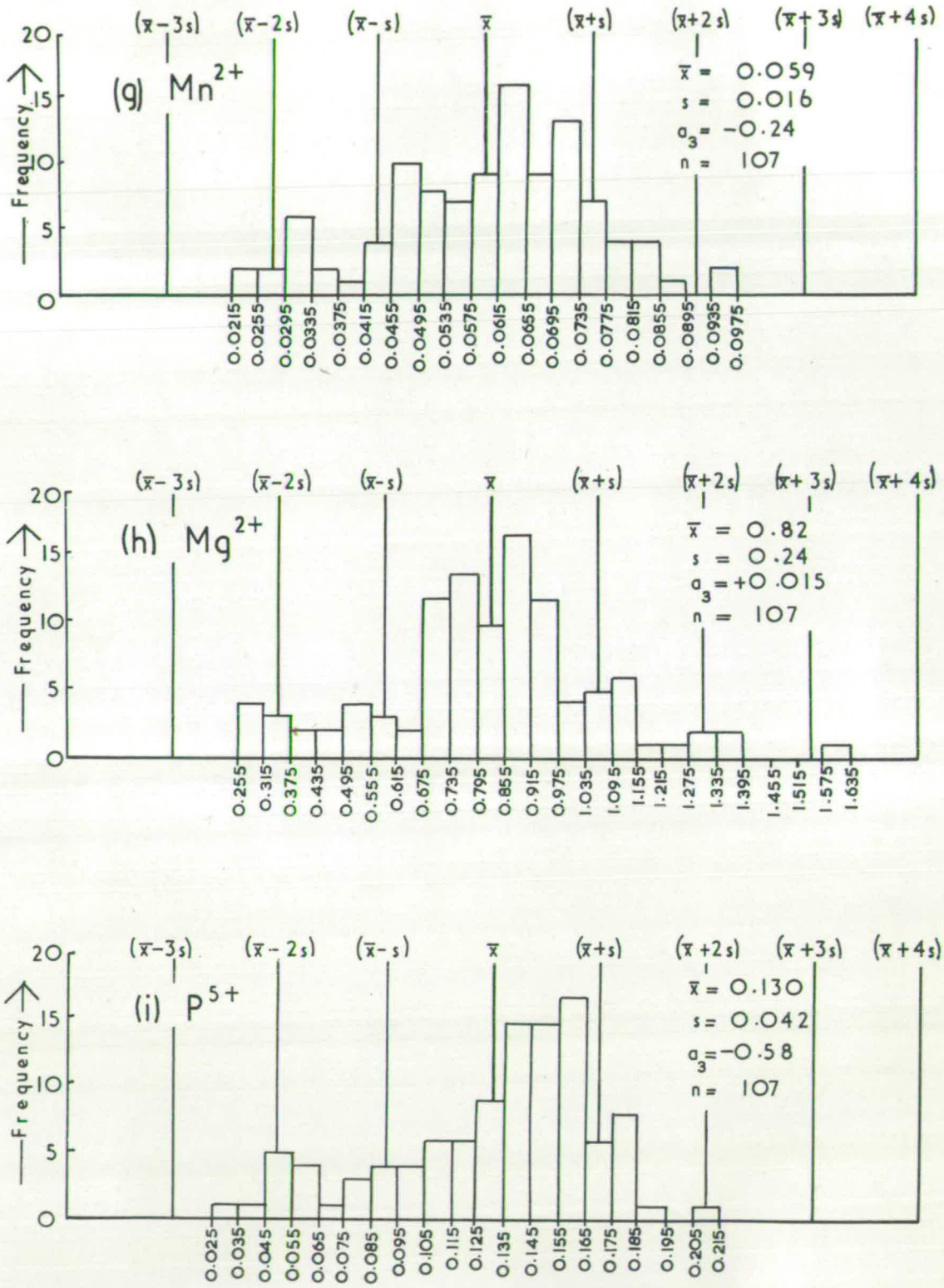


All elements as 1-cation percentages.

$\bar{x}$ = arithmetic mean. s = standard deviation (uncorrected).

a_3 = relative skewness. n = number of samples.

FIG. 13 (d–f).

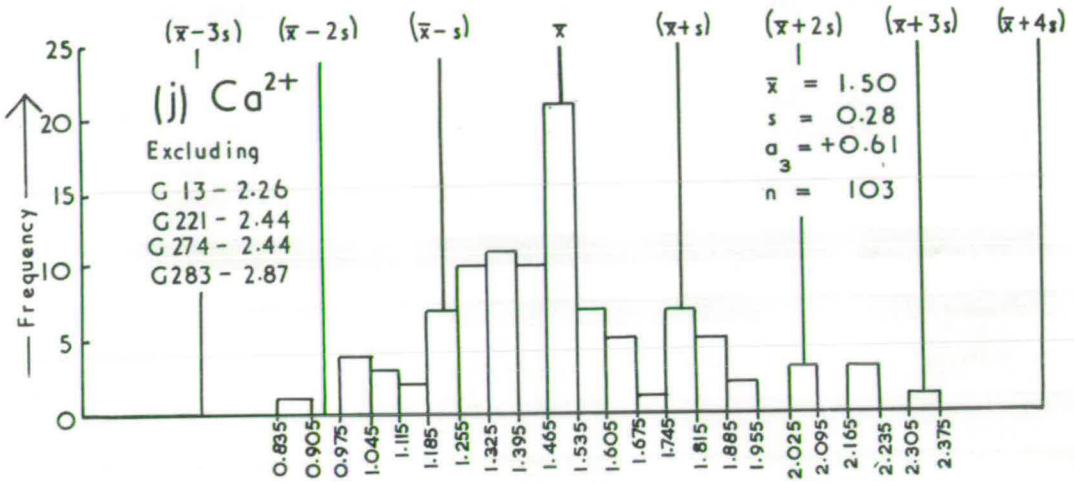


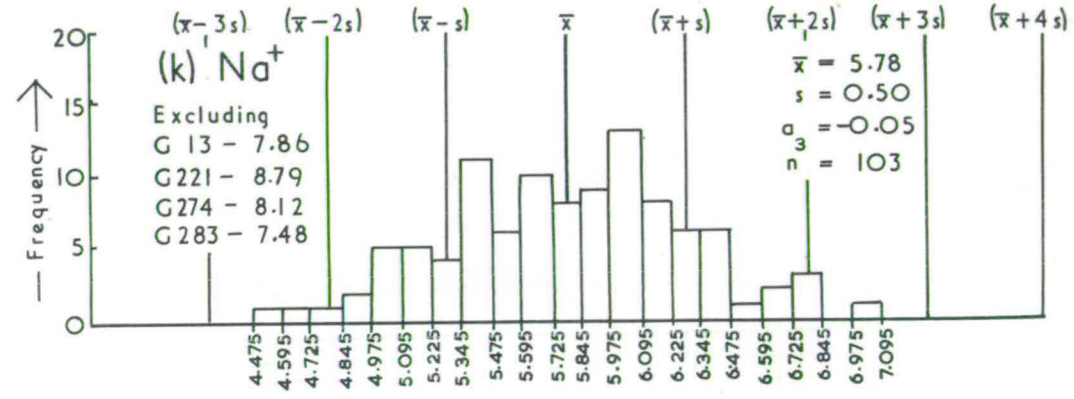
All elements as 1-cation percentages.

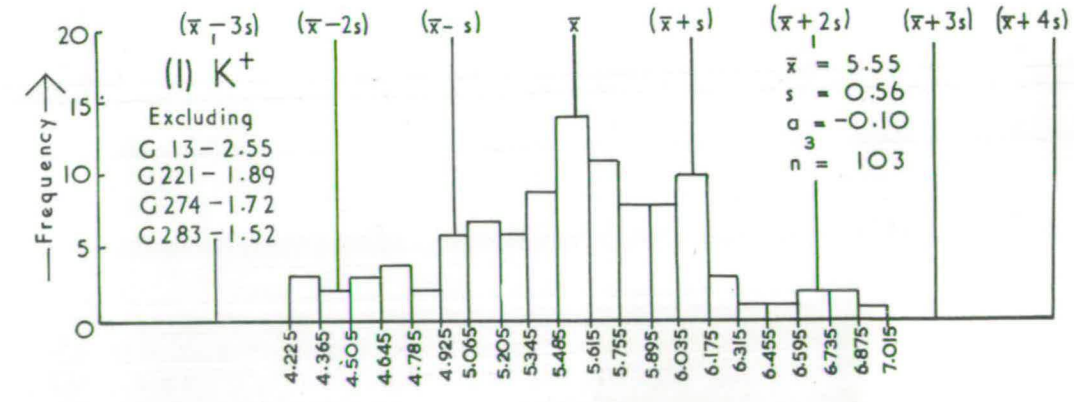
$\bar{x}$ = arithmetic mean. s = standard deviation (uncorrected)

a = relative skewness. n = number of samples.

FIG. 13 (g–i).





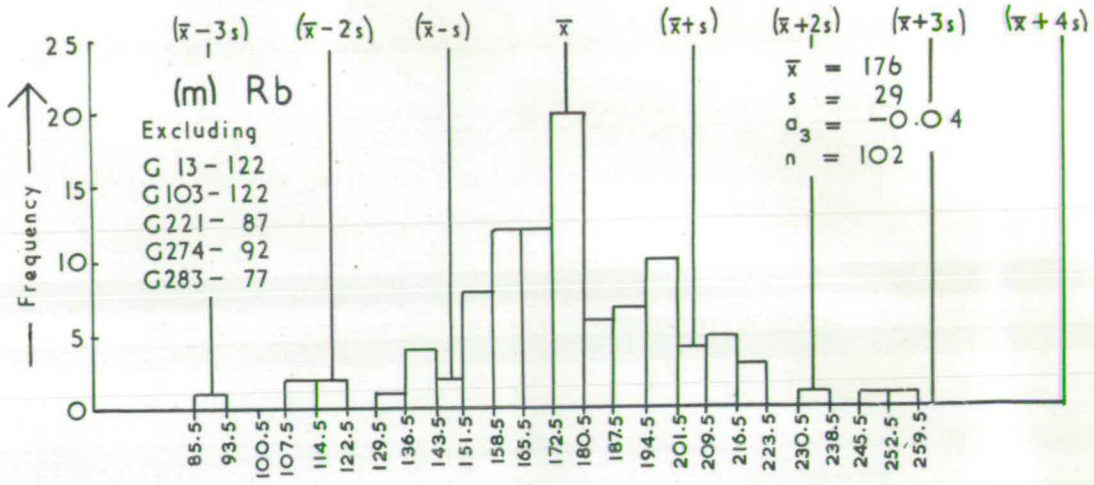


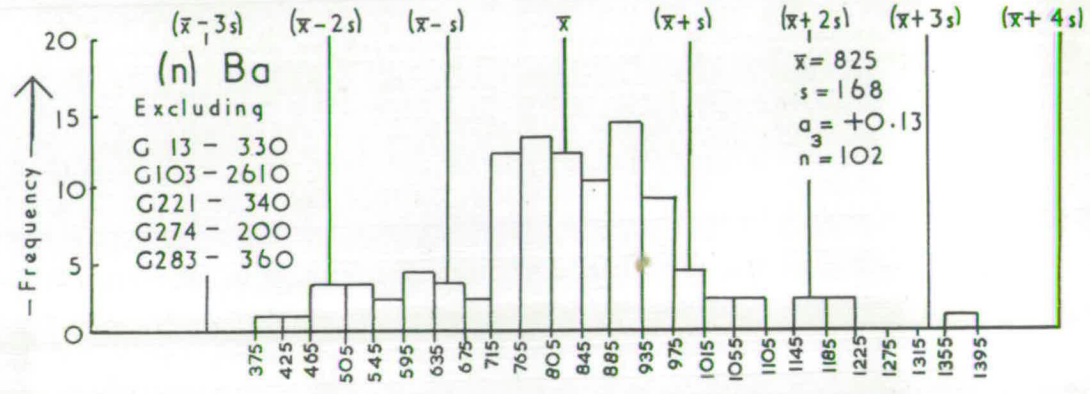
All elements as I-cation percentages.

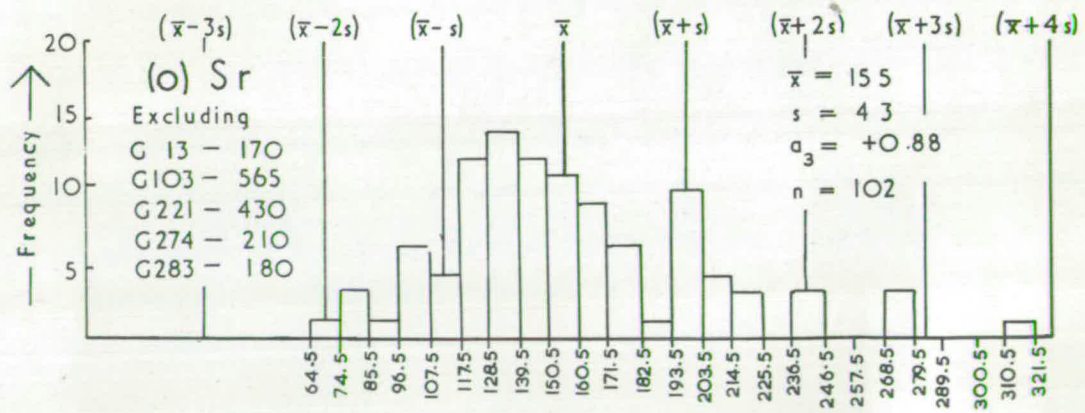
$\bar{x}$ = arithmetic mean. s = standard deviation (uncorrected).

a_3 = relative skewness. n = number of samples.

FIG. 13 (j–l).





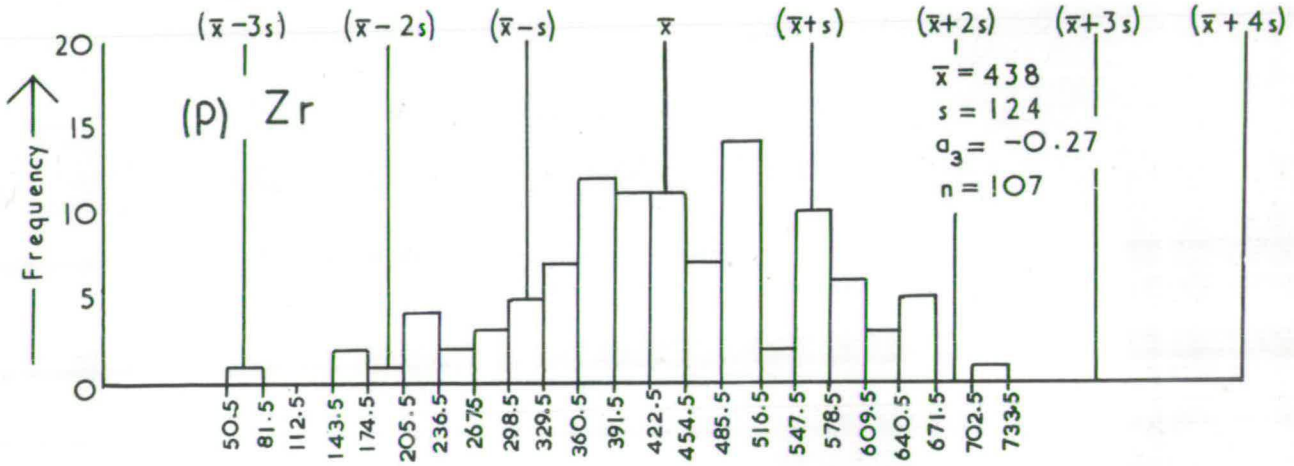


All elements as p.p.m. by weight.

$\bar{x}$ = arithmetic mean. s = standard deviation (uncorrected)

a_3 = relative skewness. n = number of samples.

FIG. 13 (m–o).



Zr given as p.p.m. by weight.

$\bar{x}$ = arithmetic mean. s = standard deviation (uncorrected).

a_3 = relative skewness. n = number of samples.

FIG. 13 (p).

the wide selection of rocks studied by Ahrens (1954). The values of C in Ahrens' Table 2 (1954, p. 64) may be compared with those in Table 23 of the present work. For this reason it is impossible to state categorically that any element has or has not a lognormal distribution in the Ardgour granitic gneiss, but where the distributions have a negative skewness, it is always true that the lognormal distribution is a worse approximation to the actual distribution than the normal distribution would be. The behaviour of the individual elements in the Ardgour granitic gneiss will now be discussed.

(i) Si: As shown in Table 14 (p.61) and Figs. 9a, 10a, and 11a, there is a very definite N-S variation in Si content within the Ardgour granitic gneiss, the lowest Si values being found in the north. In addition, the variation between adjacent samples decreases in a northerly direction. Both of these trends are highly significant (Tables 15, 16 and 21, pp.63-4and74). These two types of variation combine to give Si a distribution which has a considerable degree of positive skewness (Fig. 13a), which is largely accounted for by a few of the most siliceous samples in Groups I and II. This positively skewed distribution contrasts strongly with the negatively skewed distributions found by Ahrens (1963 b, c, 1964 a) for Si in granitic rocks.

(ii) Ti: The distribution of Ti is interesting in that, though the S-N increase in Ti is barely large enough to show statistical/

statistical significance using the F test (Table 15), yet in combination with the highly significant northward increase in the homogeneity of the granitic gneiss with respect to Ti (Table 16), it gives rise to quite a considerable negative skewness in the Ti distribution (Fig. 13(b)). In the granitic gneiss, a large proportion of the Ti is contained in the biotite, and thus it differs from elements such as P and Zr which are almost entirely concentrated in accessory minerals.

(iii) Al: There is a great similarity between the distributions of Al and Ti in the Ardgour granitic gneiss (Fig. 13 (b) and (c)), and the negative skewness of the Al distribution reflects the combination of an increase in the amount of Al and a decrease in the variability in the same direction (Tables 14 and 16). The Al and Ti distributions found here are very different from those found by Ahrens (1963 a, b) and Rooke (1964), where a considerable degree of positive skewness was found for Ti, and a lesser degree of positive skewness for Al.

(iv) Fe^{3+}: This element has a positively skewed distribution in the granitic gneiss (Fig. 13 d). The lack of any appriciable variation in the average amount of Fe^{3+} between the various groups (Tables 14, 15) means that the positive skewness is a property of most of the individual group distributions as well as the distribution of Fe^{3+} within the granitic gneiss as a whole. It may be significant that the positive skewness is associated with a large relative deviation (Table 23). The high level/

level of error involved in Fe^{3+} determinations on the granitic gneiss (with a predicted error of 7% and an actual standard deviation of duplicate analyses of 6% in the Fe_2O_3 determinations (Table 9, p.45), it is the second most imprecisely determined major element in the granitic gneiss) may also for reasons suggested by Butler, J. (1964), be responsible for the positive skewness.

(v) Fe^{2+} and $(Fe^{2+} + Fe^{3+})$: The distributions of Fe^{2+} and $(Fe^{2+} + Fe^{3+})$ are very similar, due to the approximately constant amount of Fe^{3+} in the rocks, and are shown in Fig. 13 (e) and (f). The distributions show quite a high degree of negative skewness which is explained in the same way as that of Al and Ti.

(vi) Mn: As shown in Tables 14 - 16, Mn behaves in a similar way to Fe^{2+} and its distributions is also negatively skewed (Fig. 13 (g)). The skewness is rather less, however, and the distribution bears some resemblance to that of Mg.

(vii) Mg: Despite the extremely significant N-S variation in group means and variances for Mg (Tables 14 - 16), the distribution (Fig. 13 h) is almost exactly symmetrical, though it is rather more "peaked" than a normal distribution. For this to happen, some at least of the distributions of Mg in the individual groups must have a considerable degree of positive skewness. The smallness of the individual groups, however, precludes the possibility of an accurate assessment of this skewness/

skewness. Its cause, however, may reasonably be ascribed to the high level of experimental error for this element (Butler, J. 1964), which decreases any negative skewness and increases positive skewness in any distribution. The large amount of experimental error may also account for the fact that the trend surface for Mg is rather less significant than those for Si, Al, or (Fe^{2+} + Fe^{3+}) (Table 21, p. 74).

(viii) Ca: The distribution of Ca shows a considerable degree of positive skewness (Fig. 13 j) though the average amount of Ca in the granitic gneiss varies little from group to group (Table 14). This skewness is increased greatly (Table 23) if the samples very rich in plagioclase and poor in potash felspar are taken into consideration. The variation in group means is irregular (Table 14) and the significance of the difference in group variances (Table 16)is mostly accounted for by the extreme variability of Group I. Hence the positive skewness of the Ca distribution is not a result of composition trends within the granitic gneiss, and it is doubtful whether it can all be accounted for as a result of experimental error.

(ix) Na: The distribution of Na is almost normal (Fig. 13 k) if samples G13, G221, G274 and G283 are excluded. This agrees with the fact (Tables 14 and 16) that the variability in Na content is roughly constant throughout the granitic gneiss. The inclusion of the four samples mentioned above gives the distribution a large positive skewness and probably accounts to

a/

a large extent for the trend of north-eastward increase in Na found by the computer (Fig. 9 f and Table 21).

(x) K: If samples G13, G221, G274 and G283 are excluded, the distribution of K has a slight negative skewness (Fig. 13 l) which is increased greatly by the inclusion of these samples (Table 23). The northward decrease in group means and variances (Table 14) would have been expected together to give rise to a positive skewness as in the case of Si, but the negative skewness is more in accord with the findings of Ahrens (1963 a, c) and Rooke (1964).

(xi) P: Despite the fact that all other workers (Ahrens 1963 a, Rooke, 1964 and Vistelius, 1962) found that the distribution of P in the granitic rocks studied by them had a high degree of positive skewness, the Ardgour granitic gneiss distribution (Fig. 13 i) shows a considerable degree of negative skewness. Such a distribution can be explained in the same way as those of Ti, Al, Fe^{2+} and Mn, but only if the distributions within each group do not show a high degree of positive skewness. This complete lack of any tendency to a positively skewed distribution for P is absolutely different from what would be expected either in a metasedimentary sequence where accessory minerals tend to be concentrated into widely separated bands rich in them, or in an igneous body where the crystals of an accessory mineral are often clustered together, and where Shaw (1961) predicts that an element entering only into an accessory mineral/

mineral should tend to be lognormally distributed.

(xii) Rb: The distribution of Rb in the Ardgour granitic gneiss is approximately normal (Fig. 13 m) if the samples with anomalous Ca, Na, and K contents are ignored, but if they are included it acquires a moderate negative skewness (Table 23). The resemblance to the distribution of K in the granitic gneiss and to the distribution of Rb in various New England granites (Ahrens 1954) is considerable, and contrasts strongly with distribution of Rb found by Rooke (1964) in various African granites.

(xiii) Ba: In the Ardgour granitic gneiss, the distribution of Ba (Fig. 13 n) also approaches normality if G13, G103, G221, G274 and G283 are left out of consideration, but if G13, G221, G274 and G283 are included, a moderate degree of negative skewness results (Table 23). Thus, provided sample G103 is ignored, Ba behaves very similarly to Rb and K except in its tendency to increase in the northern part of the area (Table 14). The anomalously high Ba and Sr concentrations of G103 resembles those of GS166 (Table XII, Appendix) but no satisfactory explanation of them is available.

(xiv) Sr: The element whose distribution in the granitic gneiss most closely approaches lognormality is Sr (compare Fig. 13 o with Rooke (1964) Fig. 5 where the relative deviation is much higher than in Ardgour). This distribution is similar to, but has a higher skewness than that of Ca in the granitic gneiss.

The/

The complex distribution of Sr in granitic rocks found by Ahrens (1963 b) was not expected for the Ardgour rocks, since, apart from G103, the most Sr-rich granitic gneiss (G221) only contains 430 ppm Sr, and the departure from lognormality found by Ahrens only became important at 500 ppm.

(xv) Zr: The remarks made in connection with the distribution of P are also applicable to the distribution of this element (Fig. 13 p). The negative skewness is less for this element than for P, and this is due to the lower significance of the northward increase in homogeneity of the granitic gneiss with respect to Zr than with respect to P (Tables 14, 16).

These findings are of interest because they relate to one single body of rock whose origin is uncertain. If it is not of magmatic origin and does not show geochemical features similar to those produced by crystallisation differentiation, the experimental error incurred in chemical analysis may not have the effect upon element distributions which has been described by Butler, J. (1964). However, the composition trends within the granitic gneiss have been shown to have a strong influence on the skewness of the element distributions when combined with the general northward decrease in the amount of local variability within the granitic gneiss. Contrary to the expectation of Shaw (1961), the distributions of the trace elements Rb, Ba and Sr, which are camouflaged in major rock-forming minerals, are similar to those of the major elements which they follow, Rb

and/

and Ba resembling K and Sr resembling Ca, whether the samples which are excessively rich in plagioclase and poor in potash felspar are considered or not.

The study of the distribution pattern of various elements within a single body of rock serves to amplify the findings of Rooke (1964). She discovered that whereas all of the elements except Si, Na and K were approximately lognormally distributed in a collection of African granitic rocks of widely differing ages, when this collection was split into an older and a younger group, many of the element distributions within one or other group were found to have a much smaller degree of positive skewness, and some were almost normally distributed. Thus, while the distribution of many elements may approach lognormality in rocks of a given type chosen without regard to place or time, the more restricted the field of study becomes the greater is the tendency for symmetrical or negatively skewed distributions to occur, even in the case of elements (such as P or Zr) which are largely concentrated in the accessory minerals of the rock.

4. SUMMARY OF RESULTS CONCERNING THE CHEMICAL COMPOSITION OF THE GRANITIC GNEISS.

The Ardgour granitic gneiss has the same relative proportions of quartz, potash felspar and plagioclase as would be expected to occur in a partial melt produced from rocks of the same composition as the surrounding metasediments at a pressure of 4000 - 7000/

7000 bars. There is a significant trend of composition within the granitic gneiss, consisting of a northward increase in Al, Fe^{2+}, Mn, Mg, P, Sr, Zr and possibly Ti, Na and Ba and a northward decrease in Si and possibly K and Rb with Fe^{3+} and Ca remaining roughly constant in amount, though in the case of Ca a trend of increase in an eastward direction seems to exist. This trend of composition variation is combined with a general northward decrease in local variability as regards all elements except Na and Sr. As a result, the statistical distributions about the mean of the values obtained for the various elements are different from those found by Ahrens (1954, 1963 a, b, c and 1964) and Rooke (1964). In the Ardgour granitic gneiss, the distributions of Ti, Al, Fe^{2+}, (Fe^{2+} + Fe^{3+}), Mn, P and Zr are all negatively skewed, while those of Si, Fe^{3+}, Ca and Sr have positive skewness and that of Mg is symmetrical. The distributions of Na, K, Rb and Ba are nearly normal if samples G13, G221, G274 and G283 are excluded, but inclusion of these samples produces a considerable skewness which is negative for all but Na.

V. THE CHEMICAL COMPOSITION OF THE METASEDIMENTS

The compositions of the analysed samples of the metasediments surrounding the Ardgour granitic gneiss are given in Tables VIII - XI. The analyses of pelitic rocks are listed in Table VIII, those of psammites from the Glen Garvan psammitic group in Table IX, and those of psammites from the Beinn an Tuim striped group and the Druim na Saille pelitic group in Table X. Table XI contains analyses of two calcareous psammites. The compositions of the analysed metasediments are expressed as Eskola molecular norms, first using C.I.P.W. normative minerals (Table XIII) and then using the sequence of normative minerals based on the mineral analyses of Chapter Six (Table XV). Tables VIII - XI, XIII and XV will be found in the Appendix. Petrographic data on the psammites and pelites are given in Tables 2 and 3 respectively (p.25).

1. COMPARISON WITH OTHER ANALYSES OF MOINE METASEDIMENTS

Until recently very few analyses of Moine metasediments were available, but Mercy (1963) and Butler, B. (1965 a) have improved the situation considerably. All previous published analyses of Moine rocks, except those of calc-silicates and hornblendic rocks, were brought together by Butler, B. (1965 a, Table 3), who studied in detail the variation in composition shown by them and compared it with that found in his own (1965 a, Table 2) analyses/

analyses of Moine micaschists from Ardnamurchan. All but one of Butler's analysed Ardnamurchan micaschists were fine-grained, compact, and evenly foliated; their metamorphism was considered to be indubitably isochemical. Their chemical composition was found to vary depending on their stratigraphical setting. The micaschists forming thin bands in dominantly psammitic or striped schist localities were relatively rich in K and contained little or no (2% or less) modal garnet (types 1 and 2 micaschists) while those forming the bulk of the main "pelitic" groups of the Moines or occurring in a repetition of "pelitic" lithology within the psammitic groups were relatively rich in Na and poor in K and contained over 2% of modal garnet (type 3 micaschists).

The variation in the previously analysed Moine metasediments (Butler, B. 1965 a, Table 3 and Figs. 4-15) was found to be similar to that in his Ardnamurchan rocks though the range of compositions was somewhat greater. No systematic difference between rocks described as "migmatised" and those described as "unmigmatised" was detected, except for the analyses of the Ardgour granitic gneiss (Butler, B. 1965 a, Table 3, D, K, L) included in Butler's study, which differed considerably from the other analyses studied by him. The trend represented by Cheng's sequence of "pelitic" rocks from Bettyhill (Butler, B. 1965 a, Table 3, L', M', N') was shown to involve a much smaller amount of chemical variation than existed among the unmigmatised micaschists. Butler concluded that the analyses available did not/

not prove whether or not the "injected" "pelitic" rocks had undergone a change in their bulk chemical composition during migmatisation.

The analysed "pelitic" rocks from Ardgour (analyses in Table VIII) come from the "zone of injection metamorphism", and show the features believed by Harry (1953, pp. 300-302) to represent the results of the introduction of quartzofelspathic material into these rocks. In Table 24 and Figures 14 & 15, the average Ardgour "pelitic" rock is compared with the average of Butler's Type 3 and Types 1 and 2 mica-schists, both as 1-cation percentages and as norms based on high-grade metamorphic minerals (Table XIV, Appendix). The average Ardgour "pelitic" rock is very similar to Butler's average Type 3 mica-schist though slightly less aluminous. The absence of garnet from the norm of the average Ardgour "pelite" reflects that only P125 and P166 have more than 1 or 2 percent modal garnet, in contrast to Butler's (1965 a, Table 2, Nos. 9a, 10, 16, 18, 21 and 22) Type 3 micaschists which have an average of 4.6% modal garnet. The composition of Butler's average Type 1 and 2 micaschists lies on a different part of the Na^{+} - Ca^{2+} - K^{+} diagram (Fig. 14) from any of the Ardgour "pelites", though there is a slight resemblance to P132 and P192, which differ in having a $^{Na}/Ca$ ratio resembling that of the granitic gneiss. In the proportions of (Fe^{2+} + Fe^{3+}), Mg, and/

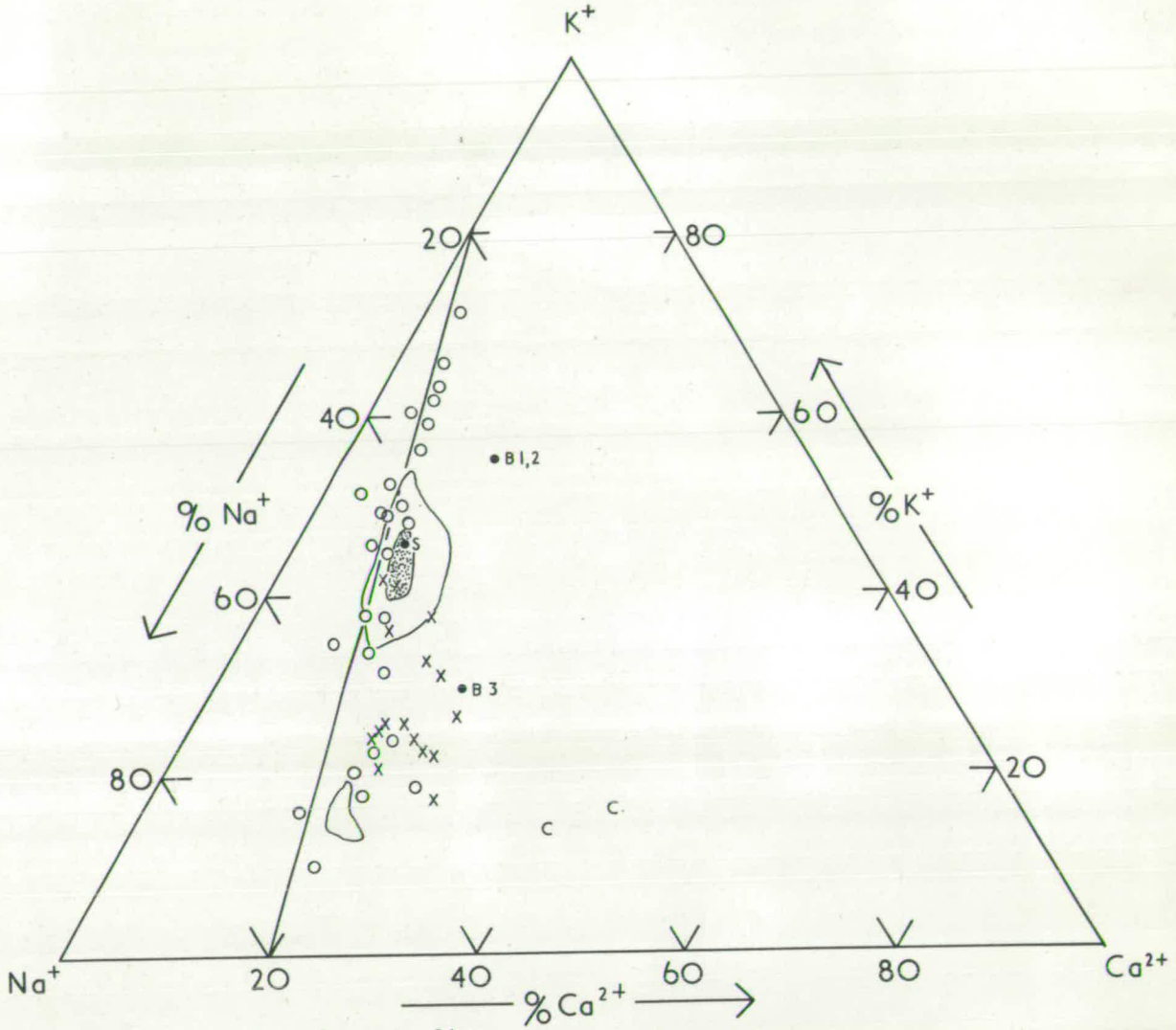
FIGURES 14 - 16.

Plots of certain chemical quantities showing the relation of the compositions of the metasediments to that of the granitic gneiss.

Figure 14 - K^+, Na^+ and Ca^{2+} (1-cation percentages).

Figure 15 - $(Na^+ + K^+)$, Mg^{2+}, and $(Fe^{2+} + Fe^{3+})$ (1-cation percentages).

Figure 16 - Normative quartz, albite and orthoclase (Eskola molecular norm using C.I.P.W. minerals).



K^+, Na^+, and Ca^{2+} are expressed as 1-cation percentages.

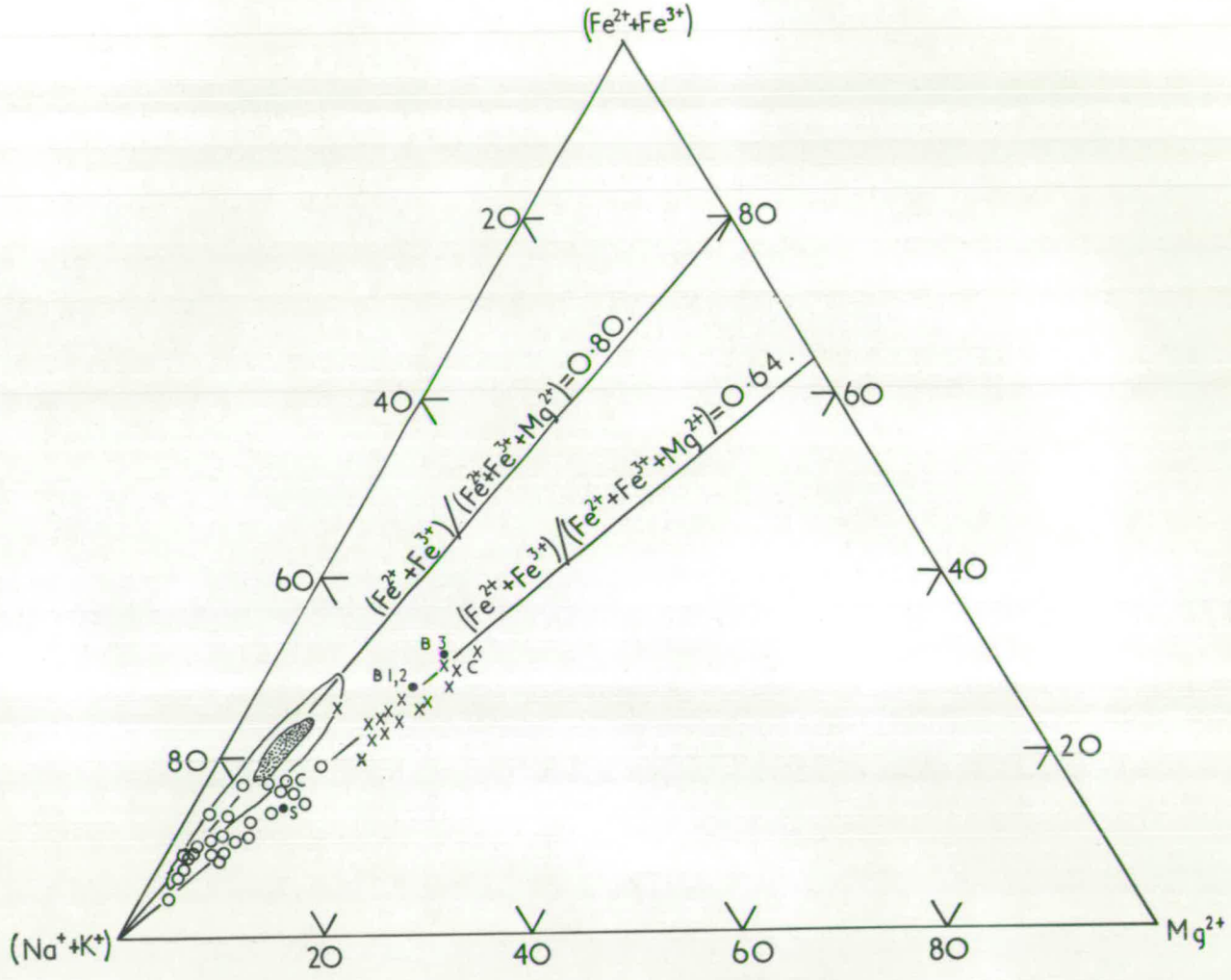
x – "Pelite".
o – Psammite.
c – Calcareous psammite.

• B1,2 – Average of 13 Type 1 & 2 micaschists (Butler, B., 1965a).
• B 3 – Average of 6 Type 3 micaschists (Butler, B., 1965a).
• s – Average of 13 previously analysed Moine psammites (Butler, B., 1965a, and Dalziel 1963a).

o – Area containing granitic gneiss composition: stippled area contains two-thirds of granitic gneiss analyses.

The straight line joins the average compositions of the analysed potash felspars and plagioclases from psammites and granitic gneisses.

FIG. 14.



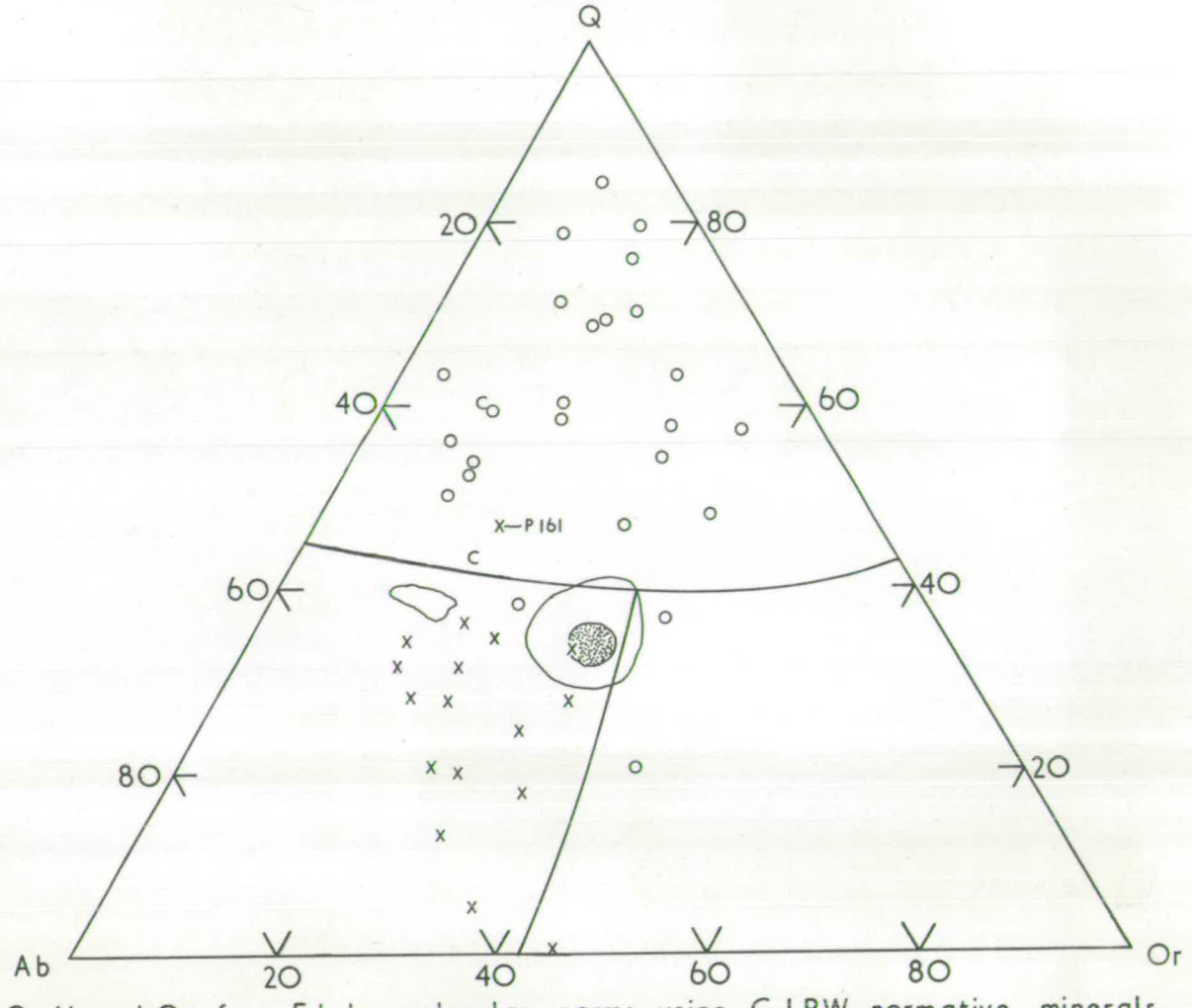
$(Fe^{2+}+Fe^{3+})$, $(Na^{+}+K^{+})$, and Mg^{2+} are expressed as 1-cation percentages.

x – "Pelite".
o – Psammite.
c – Calcareous psammite.

• B 1,2 – Average of 13 Type 1 & 2 micaschists (Butler, B., 1965a).
• B 3 – Average of 6 Type 3 micaschists (Butler, B., 1965a).
• S – Average of 13 previously analysed Moine psammites (Butler, B., 1965a, and Dalziel, 1963a).

– Area containing granitic gneiss composition: stippled area contains two-thirds of granitic gneiss analyses.

FIG. 15.



Q, Ab, and Or from Eskola molecular norms using C.I.P.W. normative minerals (see Table XIII, Appendix).

x – "Pelite."
o – Psammite.
c – Calcareous psammite.

—Area containing granitic gneiss composition: stippled area contains more than half of granitic gneiss analyses.

—Cotectic lines at 2000 bars p_{H_2O} for Ab/An=4.1 (molecular), converted to molecular percentages from von Platen (1965).

FIG. 16.

and (Na + K), the type 1 and 2 micaschists are the same as the type 3 micaschists and the Ardgour "pelites" (Figure 15).

The composition represented by Butler's Type 3 micaschists and by most of the Ardgour "pelites" is a perfectly possible one for a sedimentary rock. It is similar to that of Pettijohn's (1957) average greywacké (quoted in Table 24) but contains more K and Al, and Mercy (1963, p. 206) considered that his analyses of Moine "pelitic" and psammitic gneisses were chemically equivalent to arkosic sediments. This does not prove that the Ardgour "pelitic" rocks were, in fact, derived isochemically from their parent sediments, but the similarity in composition to Butler's Type 3 micaschists, of which those classified by him as Type 3b probably come from the same stratigraphic horizon as the Ardgour "pelites", makes it very likely that the bulk composition of the Ardgour "pelites" is the same as that of the sediments from which they were formed. This, however, does not entitle any similar conclusion to be drawn about the "pelitic" rocks of similar composition to that of the Ardgour "pelites" which are involved in migmatite complexes such as those of Sutherland, especially since the existence of unmigmatised metasediments of this composition so far north in the Moines has not yet been proved.

The number of previous analyses of Moine psammites is even less than of "pelitic" rocks. They are all quoted by Butler, B. (1965 a,/

(1965 a, Table 3, A, N, S, W, X, G', O', Q', U', V', W' and X'), the last six coming from Mercy (1963). The average of these analyses together with one from Dalziel (1963 a, Appendix II, No. 5) is given in Table 25 where it is compared with the average of the Ardgour psammites (i) belonging to the Beinn an Tuim striped group and Druim na Saille pelitic group and (ii) belonging to the Glen Garvan psammitic group (see Tables IX and X for actual analyses). The average of all previous analyses of Moine psammites is very similar to that of the psammites from the Glen Garvan psammitic group analysed in the present study. The average composition of the psammites from the Beinn an Tuim striped group and the Druim na Saille pelitic group is slightly richer in Si and poorer in the remaining elements, but the only important difference lies in the much higher $^{Na}/_{K}$ ratio of these rocks, in which they resemble the Ardgour "pelitic" rocks (Table 26). Both groups of Ardgour psammites, however, show a greater degree of variability in $^{Na}/_{K}$ ratio than do the previous analyses of Moine psammites. In the previously analysed Moine psammites, $^{Na}/_{K}$ (calculated from 1-cation percentages) ranges from 0.31 to 1.42, while in the Glen Garvan group psammites it varies from 0.35 to 4.10, and in the Druim na Saille and Beinn and Tuim group psammites $^{Na}/_{K}$ values from 0.58 to 7.28 occur. As shown in Figures 14 and 15, this variation in $^{Na}/_{K}$ ratio in the Ardgour psammites is/

TABLE 24.

Comparison of Ardgour "pelitic" rocks with other Moine "pelites" and with Pettijohn's (1957) average greywacke.

(a) 1-cation percentages.

	A	B	C	D
Si^{4+}	59.73	57.93	60.92	63.31
Ti^{4+}	0.68	0.73	0.73	0.37
Al^{3+}	18.66	20.26	19.18	17.07
Fe^{3+}	0.67	0.63	1.22	1.10
Fe^{2+}	3.86	4.74	3.51	3.19
Mn^{2+}	0.09	0.08	0.06	0.08
Mg^{2+}	2.82	3.00	2.60	3.21
Ca^{2+}	2.61	3.00	1.66	3.25
Na^{+}	6.92	5.72	3.51	5.88
K^{+}	3.80	3.68	6.46	2.37
P^{5+}	0.17	0.21	0.16	0.16

A - Average of 15 new analyses of Ardgour "pelitic" rock.

B - Average of 6 "Type 3" garnetiferous mica-schists from Ardnamurchan (Butler 1965a).

C - Average of 13 "Type land 2" mica-schists from Ardnamurchan (Butler 1965a)

D - Average greywacke (Pettijohn 1957).

(b) Eskola Molecular Norms, recalculated to high-grade metamorphic minerals.

	A	B	C	D
Quartz	23.33	24.51	28.56	32.17
Potash Felspar	4.65	-	19.77	-
Plagioclase	45.42	42.00	20.70	44.45
Muscovite	2.18	5.23	9.12	-
Garnet	-	1.23	-	2.51
Biotite	23.69	26.38	19.90	19.81
Hornblende	-	-	-	-
Magnetite	0.32	0.16	1.25	1.06
Ilmenite	-0.02	-0.05	0.30	-0.40
Apatite	0.43	0.53	0.40	0.40
% An in plagioclase.	26	32	34	34

TABLE 25.

Comparison of Ardgour psammitic rocks with previous analyses of Moine psammites.

(a) 1-cation percentages.

	E	F	G
Si^{4+}	81.90	77.34	75.32
Ti^{4+}	0.22	0.27	0.31
Al^{3+}	9.16	11.28	11.81
Fe^{3+}	0.28	0.25	0.56
Fe^{2+}	0.70	0.97	1.15
Mn^{2+}	0.02	0.03	0.04
Mg^{2+}	0.47	0.53	0.89
Ca^{2+}	0.94	1.16	1.09
Na^{+}	4.02	4.17	4.29
K^{+}	2.24	3.97	4.48
P^{5+}	0.04	0.03	0.05

E - Average of 10 new analyses of Moine psammites, one from the Druim na Saille Pelitic group and 9 from the Beinn an Tuim Striped group.

F - Average of 16 new analyses of Moine psammites from the Glen Garvan psammitic group.

G - Average of 13 previously analysed Moine psammites. (Butler 1965a, Table 3 Nos.A,N,S,W, X,G′,O′,Q′,U′,V′, W′,X′ and Dalziel 1963a, Appendix II, no.5).

(b) Eskola Molecular norms, recalculated to high grade metamorphic minerals.

	E	F	G
Quartz	61.43	50.79	47.11
Potash felspar	10.40	20.70	22.81
Plagioclase	22.42	22.26	21.94
Muscovite	1.66	0.99	0.92
Garnet	-	-	-
Biotite	3.42	4.66	6.14
Hornblende	-	-	
Magnetite	0.32	0.24	0.66
Ilmenite	0.24	0.27	0.26
Apatite	0.10	0.08	0.13
% An in plagioclase	20	25	23

TABLE 26.

Comparison of certain chemical ratios in the metasediments and in the granitic gneiss.

	E	F	A	H
Na/K	1.796	1.052	1.821	1.041
$(Fe^{2+}+Fe^{3+})/(Fe^{2+}+Fe^{3+}+Mg)$	0.680	0.702	0.621	0.783
K/Rb	315	290	196	216
K/Ba	29.4	28.6	33.1	46.1
Ca/Sr	28.0	30.5	51.8	68.2
(Ca+Na+K)/Sr	163	194	205	463

Na/K and $(Fe^{2+}+Fe^{3+})/(Fe^{2+}+Fe^{3+}+Mg)$ calculated from 1-cation percentages.

K/Rb, K/Ba, Ca/Sr and (Ca+Na+K)/Sr given as ratios by weight.

E - Average of 10 new analyses of Moine psammites, one from the Druim na Saille pelitic group and 9 from the Beinn an Tuim Striped Group.

F - Average of 16 new analyses of Moine psammites from the Glen Garvan psammitic group.

A - Average of 15 new analyses of Ardgour "pelitic" rock.

H - Average granitic gneiss (see Table 18).

is not accompanied by any large changes in the $^{Na}/_{Ca}$ or $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratios, which are appreciably constant.

The calcareous rocks analysed (Table XI) are not, unfortunately, typical calc-silicates. The sample S172 consists of alternating fine bands of calcareous and semipelitic material, while L116, though homogeneous, is intermediate between a calc-silicate and a normal psammite. The presence of muscovite in the modified norm of the rock (Table XV) is accounted for by the fact that garnet is the chief ferromagnesian mineral. The petrography of these two samples is described briefly on p.27. Chemically, these two rocks are distinguished from the other analysed Ardgour metasediments by their richness in Ca and Mn and, inthe case of L116, poverty in Ti. Three analyses of Moine calc-silicates are given by Kennedy (1949); their high Mn contents are striking, but they do not, in general, have a lower Ti content than most Moine psammites.

2. COMPARISON WITH THE GRANITIC GNEISS

The main chemical features which distinguish the Ardgour metasediments from the granitic gneiss can be conveniently displayed in triangular diagrams. In Figure 14 the relative proportions of Ca^{2+}, Na^{+} and K^{+} (1-cation percentages) in the rocks are shown. The psammites and granitic gneisses lie very close to/

to a line joining $K_{0.83}Na_{0.17}$ to $Na_{0.80}Ca_{0.20}$, which corresponds to the average compositions of the potash felspar and plagioclase in these rocks (see Chapter Six). The "pelites", however, tend to plot to the calcic side of this line; this is reflected in their rather more calcic plagioclase (Table 3) and in the fact that the chief K-bearing mineral in these rocks is biotite, which is much poorer in Na than potash felspar.

As can be seen from the diagram, the psammites show a much greater range in Na/K ratio than the granitic gneisses; the tendency noted above for the psammites from the Glen Garvan psammitic group to be less sodic than those from the Beinn an Tuim striped group is distinguishable, but not very well marked. The two calcareous psammites clearly differ from the other metasediments and from the granitic gneiss, but have a Na/K ratio similar to that of many of the "pelitic" rocks.

The plot of $(Fe^{2+} + Fe^{3+} + Mn)$ against Mg and (Na + K) (from 1-cation percentages) is shown in Figure 15. The granitic gneisses have a $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio close to 0.80 (see also Table 26), while for the "pelites" and calcareous psammites this ratio has a value close to 0.64, except in the case of P161 which in many features is transitional to the granitic gneiss. The $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio of the psammites is intermediate between that of the "pelites" and that of the granitic gneisses, but there is a greater variation in/

TABLE 27.

Exploration of the possibility of isochemical formation of the granitic gneiss from a mixture of psammitic and pelitic material.

	E	C	K	H	L	P161
Si^{4+}	81.90	60.92	71.41	68.12	70.82	70.03
Ti^{4+}	0.22	0.73	0.48	0.41	0.45	0.57
Al^{3+}	9.16	19.18	14.17	14.67	13.91	14.07
Fe^{3+}	0.28	1.22	0.75	0.70	0.48	0.40
Fe^{2+}	0.70	3.51	2.10	2.26	2.28	2.83
Mn^{2+}	0.02	0.06	0.04	0.06	0.06	0.09
Mg^{2+}	0.47	2.60	1.54	0.82	1.64	1.14
Ca^{2+}	0.94	1.66	1.30	1.50	1.78	2.20
Na^{+}	4.02	3.51	3.76	5.78	5.47	5.78
K^{+}	2.24	6.46	4.35	5.55	3.02	2.79
P^{5+}	0.04	0.16	0.10	0.13	0.10	0.10
Rb	48	265	156	176	92	91
Ba	514	1075	794	825	656	610
Sr	232	282	257	155	294	270
Zr	288	215	251	438	296	470
Na/K	1.796	0.543	0.864	1.041	1.811	2.072
$(Fe^{2+}+Fe^{3+})$ / $(Fe^{2+}+Fe^{3+}+Mg)$	0.680	0.648	0.652	0.783	0.632	0.744
K/Rb	315	168	191	216	226	211
K/Ba	29.4	41.3	37.5	46.1	31.6	31.4
Ca/Sr	28.0	41.3	35.3	68.2	42.4	57.4
(Ca+Na+K)/Sr	163	249	210	463	188	215
Eskola molecular norms, recalculated to high-grade metamorphic minerals.						
Quartz	61.43	28.56	45.00	31.64	42.38	40.48
Potash Felspar	10.40	19.77	15.08	27.94	7.52	7.08
Plagioclase	22.42	20.70	21.56	29.05	33.92	37.73
Muscovite	1.66	9.12	5.39	-	1.92	0.34
Biotite	3.42	19.90	11.66	9.02	13.56	13.55
Hornblende	-	-	-	1.00	-	-
Magnetite	0.32	1.25	0.78	0.73	0.32	0.21
Ilmenite	0.24	0.30	0.27	0.27	0.11	0.35
Biotite	0.10	0.40	0.25	0.33	0.26	0.25
% An in plag.	20	34	27	20	23	27

Major elements as 1-cation percentages, trace elements as ppm by weight.

Na/K and $(Fe^{2+}+Fe^{3+})/(Fe^{2+}+Fe^{3+}+Mg)$ calculated from 1-cation percentages.
K/Rb, K/Ba, Ca/Sr and (Ca+Na+K)/Sr ratios given as weight proportions.

E - Average of 10 new analyses of Moine psammites, one from the-Druim na Saille pelitic group and 9 from the Beinn an Tuim striped group.

C - Average of 13 "Type 1 and 2" micoschists from Ardnamurchan (Butler, B.1965a) Trace elements, however, are average of P132 & P192 (Table VIII).

K - Mixture of 1 part E + 1 part C.

H - Average granitic gneiss (Table 14).

L - Mixture of 1 part E + 1 part A (Table 24).

P161 - sample of felspathic migmatite believed to be of pelitic origin.

in this ratio from sample to sample in the psammites than in the other rocks. Much of this variation, however, may be due to experimental error, because the amount of Mg in the psammites is often very low. The proportion of alkalis to iron and magnesia in the granitic gneisses is intermediate between that in the psammites and that in the "pelites".

The relative Q - Ab - Or contents of the metasediments (Eskola molecular norms based on the C.I.P.W. normative sequence, see Table XIII) are depicted in Figure 16. No concentration of the metasediment analyses about the ternary minimum composition can be seen. The diagram shows that if they were partially melted until two of the felsic minerals had disappeared all of the psammites except S42 and possibly two others would leave a residue of quartz, while all of the "pelitic" rocks except P161 would leave an anorthositic residue of plagioclase.

As shown by comparison of the norms in Table XV and the averaged values in Tables 25 and 27, the granitic gneisses also differ from the psammites and "pelites" in the level of Al saturation. Almost all of the "pelitic" and psammitic rocks have normative muscovite according to the mineral formulae used, the only significant exceptions being S77, P132 and P192, all of which are also unusual in having plagioclases more sodic than that found in the granitic gneiss. The granitic gneisses are, however, almost exactly Al-saturated with respect to the

assemblage/

assemblage quartz-potash felspar-plagioclase-biotite, though the southern granitic gneisses are mostly slightly undersaturated and the northern ones (those of Groups VI and VII especially) tend to be slightly oversaturated, as evidenced by the frequent occurrence of modal muscovite and sometimes garnet in these rocks.

The trace element compositions of the metasediments and of the granitic gneisses are compared, for Rb, Ba and Sr, in Figures 17-20. In Figure 17, the relationship between K and Rb is shown, and the $^{K}/_{Rb}$ ratio in the "pelites" is seen to be very similar to that in the granitic gneisses, whereas in the psammites it is very much higher (see also Table 26). However, the psammites show a definite relationship between the amounts of K and Rb in the rocks, but this relationship is very poorly developed in the "pelites" and in the granitic gneisses it is reflected only in a slight elongation of the cluster which the granitic gneiss analyses form.

The relationship between K and Ba is shown in Fig. 18. For both the psammites and "pelites" a distinct positive correlation between the two elements is noticeable and, as shown in Table 26, the $^{K}/_{Ba}$ ratio is very similar for the two types of rock. In the case of the granitic gneisses, however, there is, except for the samples G13, G221, G274 and G283, very little correlation between the Ba and K contents, and the analyses are clustered about the mean Ba and K values in an apparently random manner. The $^{K}/_{Ba}$ ratio is definitely higher in the granitic gneisses/

FIGURES 17 - 20.

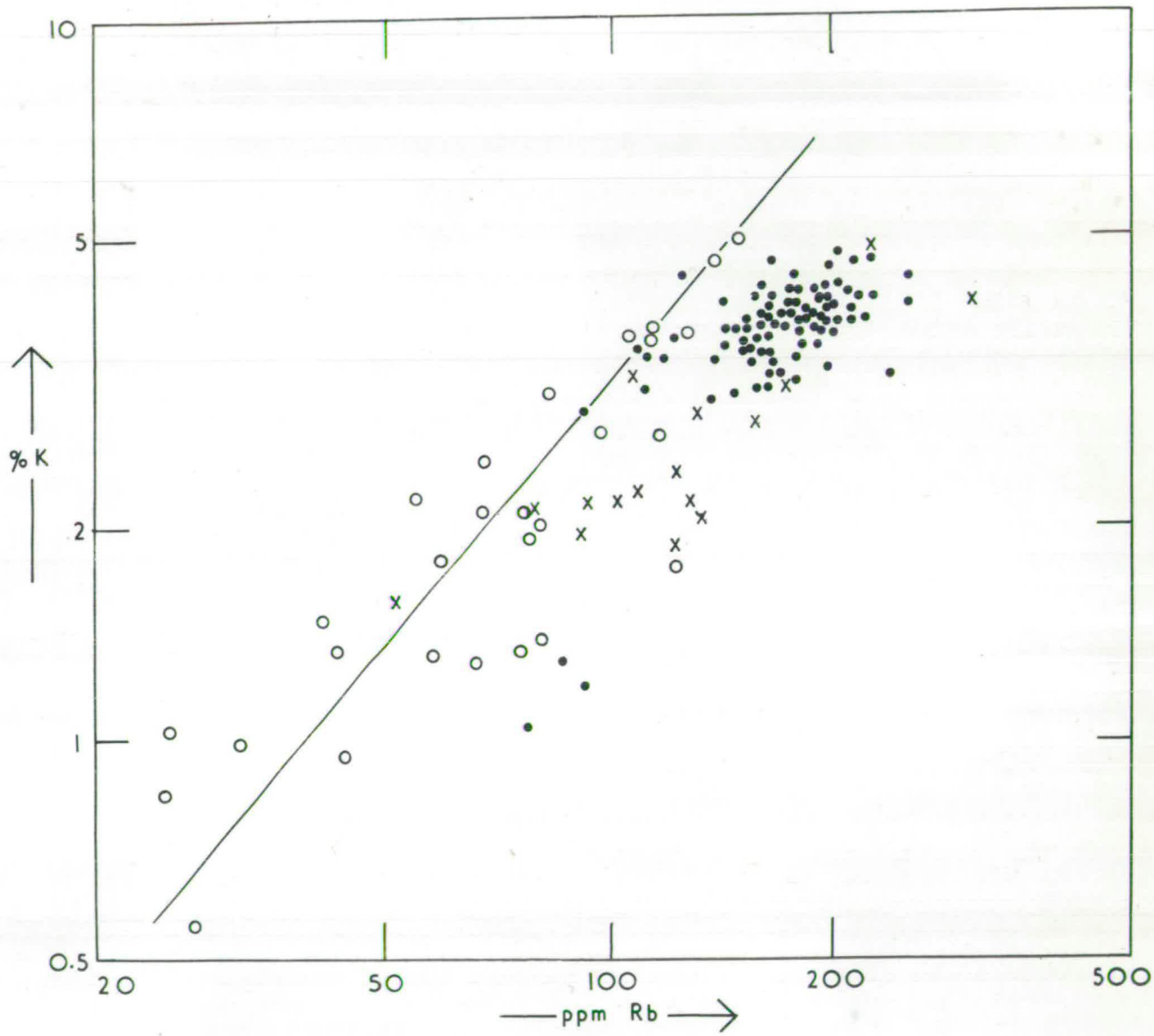
The relationship between certain trace and major elements in the Ardgour rocks.

Figure 17 - Rubidium and potassium.

Figure 18 - Barium and potassium.

Figure 19 - Strontium and calcium.

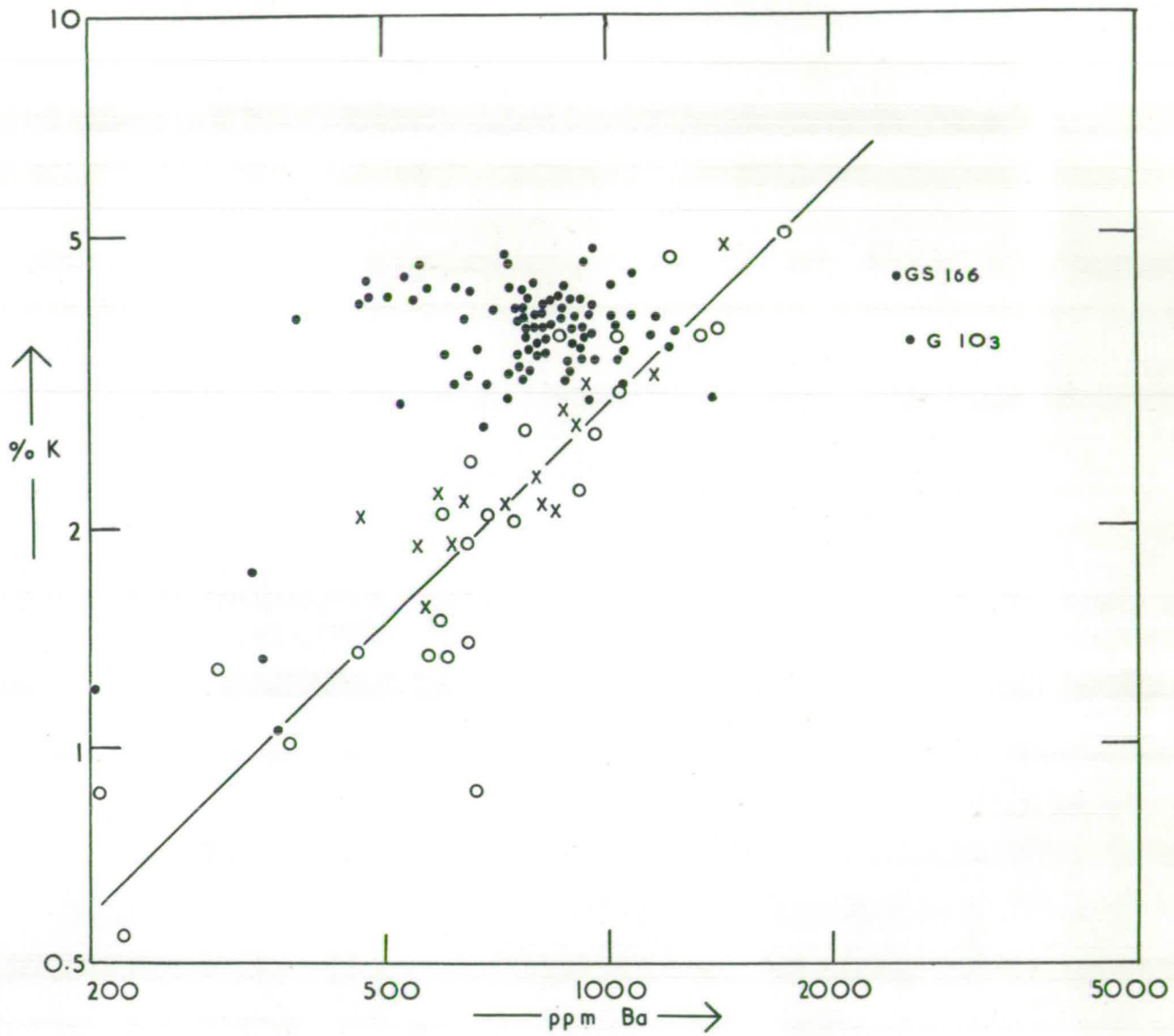
Figure 20 - Strontium and sodium.



• Ardgour granitic gneiss.
x Ardgour pelite.
o Ardgour psammite.

The straight line illstrates the approximate constancy of K/Rb in the psammites.

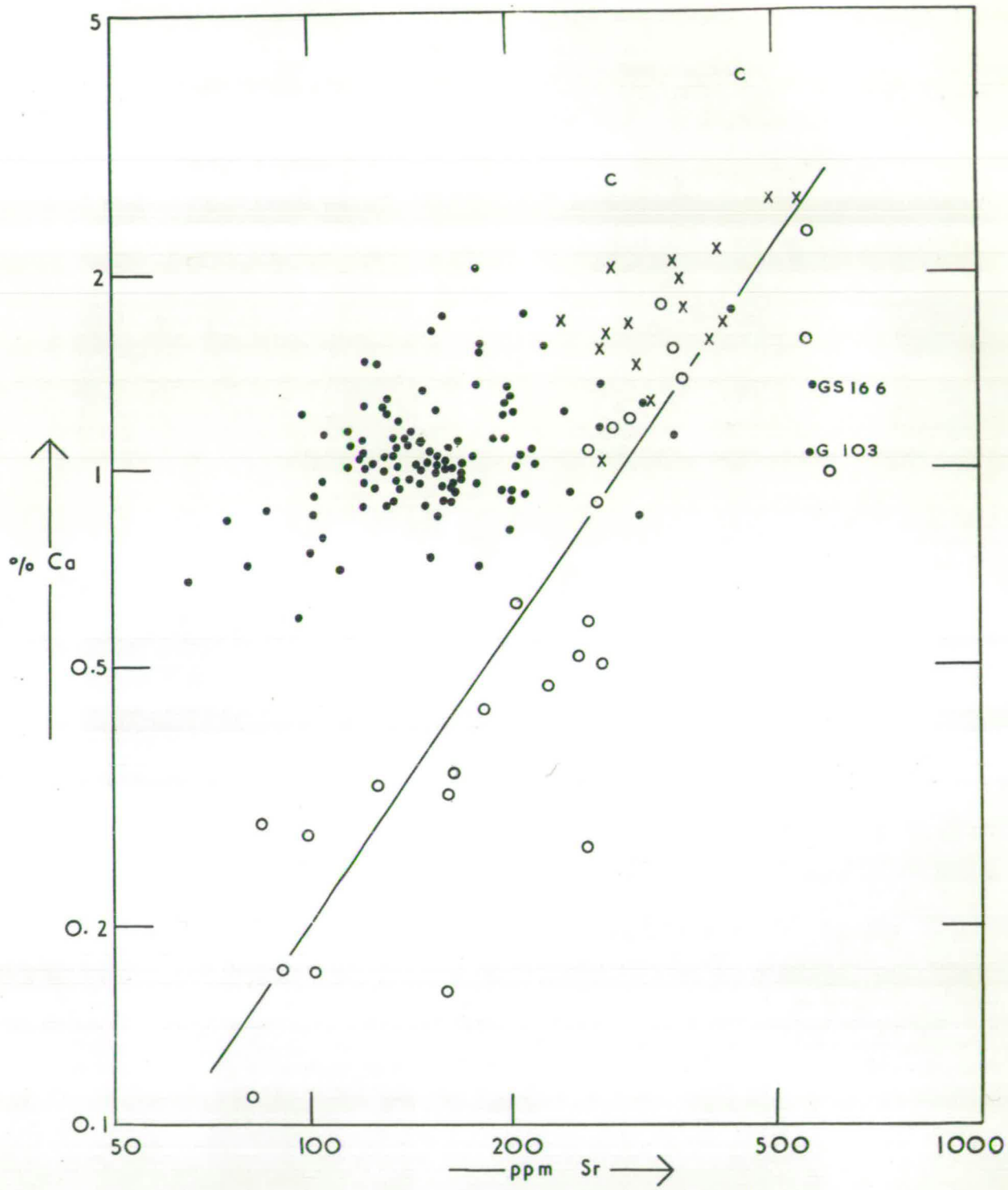
FIG. 17.



• Ardgour granitic gneiss.
x Ardgour pelite.
o Ardgour psammit e.

The straight line illustrates the approximate constancy of K/Ba in the psammites.

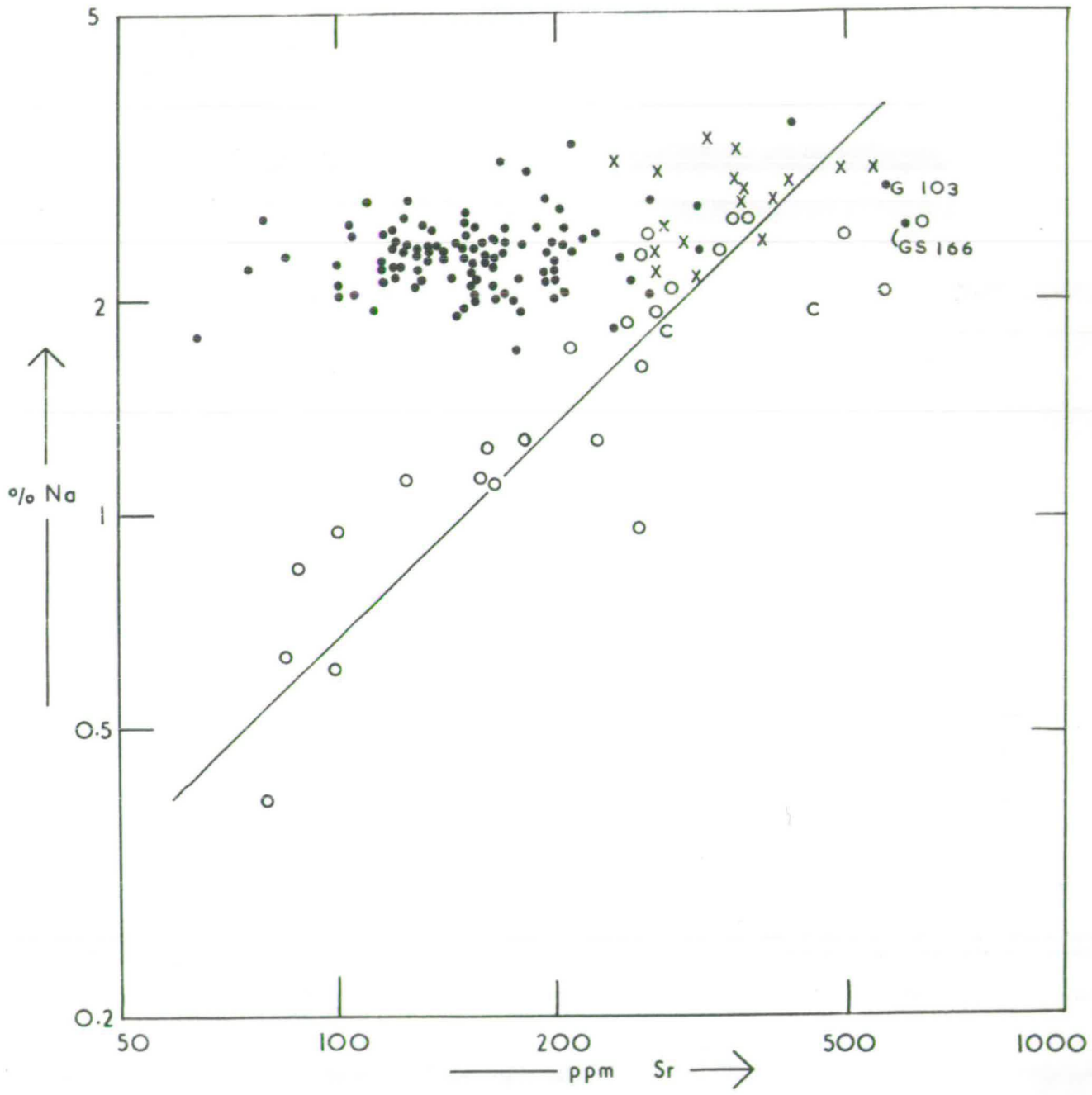
FIG. 18.



• Ardgour granitic gneiss.
x Ardgour pelite.
o Ardgour psammite.
c Calcareous psammite.

The straight line illustrates the relationship between Ca and Sr in the psammites.

FIG. 19.



• Ardgour granitic gneiss.
x Ardgour pelite.
o Ardgour psammite.
c Calcareous psammite.

The straight line illustrates the relationship between Na and Sr in the psammites.

FIG. 20.

gneisses than in the psammites and "pelites".

The behaviour of Sr is similar to that of Ba in that the granitic gneisses are definitely impoverished in Sr as against Ca (Fig. 19) or even Na (Fig. 20) compared with the psammites and "pelites". In the psammites there is a positive correlation between Sr and Ca (Fig. 19), but just as good a correlation is obtained (Fig. 20) if Sr is plotted against Na. The granitic gneisses show a considerable spread whether Sr is plotted against Ca or Na, and the correlation between Sr and Ca or Na is very low. In both diagrams (Figures 19 and 20), the "pelites" occupy a position between that of the psammites and the granitic gneisses, and show less correlation between Sr and Ca or Na than the psammites but more than the granitic gneisses.

The bulk composition of the granitic gneiss thus differs in a number of respects from that of the surrounding rocks. The granitic gneiss was supposed by Dalziel (1963a, pp. 206-208) to have formed in some way from the Beinn an Tuim striped group, and it was therefore though desirable to compare the average composition of the rocks of the Beinn an Tuimm striped group with that of the granitic gneiss in order to see what chemical changes would be required to form the granitic gneiss from the rocks of the striped group. Direct comparison of the average compositions of the two masses of rock was not possible due to the great inhomogeneity of the striped group and to the relative poverty of exposure in the region where the striped

group/

group outcrops, which resulted in only the more quartzose psammites being collected for analysis. Thus to obtain an average composition for the Beinn an Tuim striped group a considerable amount of pelitic material must be added to the average striped group psammite (whose composition is given as E in Tables 25 and 27). Butler, B. (1965 a) showed that "pelitic" rocks forming parts of the striped schist lithology (Butler, 1965 a, Table 2, nos. 13 and 17) are similar in composition to those forming bands among the quartzites and arkoses of the major psammitic groups of the Moines (Butler, B. 1965 a, Table 2, Nos. 1, 4, 5, 7, 8, 9, 11, 12, 14, 15, 19 and 20) and differ from those forming the bulk of the "pelitic" groups of the Moines. Hence the granitic gneiss composition is compared with a mixture of average striped group psammites and Butler's type 1 and 2 micaschists (Butler, B. 1965a, Table 5,A) in the arbitrary proportion of 1:1 (Table 27K) which gives approximately the same amount of ferromagnesian constituents as in the granitic gneisses. To estimate the average amounts of Rb, Ba, Sr and Zr in the striped group, one part of average striped group psammite has been combined with one part of the average of P132 and P192, these being the Ardgour "pelites" which most resemble Butler's type 1 and 2 micaschists. The granitic gneiss composition (Table 27 H) is also compared with a mixture of equal proportions of striped group psammites and Ardgour "pelitic" rock (Table 27 L).

As/

As shown in Table 27, whatever the composition of the pelitic bands within the Beinn an Tuim striped group, the formation of a rock of the composition of the Ardgour granitic gneiss from the striped group necessitates the addition of Na or K or both, and the subtraction of Mg relative to Fe, Sr relative to Ca or (Ca + Na + K), and probably of Ba relative to K. The chemical changes required also involve a change from a slight excess of Al to a slight deficit of Al with respect to the assemblage quartz-plagioclase-potash felspar-biotite. The implications of these facts with respect to the origin of the granitic gneiss will be discussed in Chapter Seven.

The similarity of the average of the Ardgour "pelite" and the striped group psammite compositions to that of sample P161 is striking (Table 27). The only significant differences are in Ca/Sr and $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ in which P161 resembles the granitic gneisses. Its massive gneissose texture and its close proximity to the granitic gneiss outcrop combine with the chemical evidence to suggest that the sample P161 is a pelitic or semipelitic rock which has undergone partial conversion to a granitic gneiss.

VI. THE MINERALS OF THE ROCKS

Nineteen rocks were chosen for mineral separation. These were: ten granitic gneisses, including G103, which has remarkably high Ba and Sr contents and is very poor in dark minerals, and GB, which is much poorer in potash felspar than most of the granitic gneisses, four psammites, four pelites, and a calcareous psammite.

Biotite and plagioclase were separated from all of these rocks, potash felspar from all except P125, P223, P291 and S172, muscovite from G241, P291 and SK211, and hornblende from S172. The biotites and hornblende were analysed in full for major elements, and also for Rb, Ba, and Sr. The muscovites and felspars were analysed for Ca, Na, K, Rb, Ba and Sr, while $(OH)^-$ and F were also determined in the muscovites.

The biotite and hornblende analyses are given as weight percentages in Table XVII and as mineral formulae based on 24 (O + OH + F) anions in Table XVIII. This calculation was conducted according to the method of Foster (1960, pp. 13-15), because the values of X and Y are no longer dependent upon the H_2O results, which are not very accurate at such high levels of concentration. It was assumed that the P present in the separated fractions represented contamination by apatite, and the Ca results were corrected to allow for this. Table XIX contains the potash felspar and muscovite analyses, both as weight/

weight per cent of oxide, and as molecular percentages of Orthoclase, Albite and Anorthite (or Muscovite, Paragonite and Margarite in the case of the muscovites). The plagioclase separations contained varying amounts of quartz, but in all cases but one, the percentage of plagioclase exceeded 40. The analyses of the (plagioclase + quartz) mixtures are given in Table XX, together with the results converted to 100% plagioclase (assuming that the quartz is pure SiO_2 and the felspars have the correct stoichometric composition), and expressed both as weight of oxide per cent, and as molecular percentages of Or, Ab and An. Tables XVII - XX will be found in the Appendix.

The biotite analyses will be discussed below and compared with other analyses of metamorphic and igneous biotites, and then the analyses of coexisting minerals from the same rock will be discussed and compared with the findings of other workers in this field.

1. THE CHEMICAL COMPOSITIONS OF THE BIOTITES.

As shown in Table XVIII, the biotite analyses satisfy the formula requirements fairly well, except for the biotite from G262. This biotite, which is very low in K and excessively high in OH, has undergone some form of hydrothermal alteration akin to chloritisation, and though it remains brown in thin section, some form of alteration can be seen to be taking place; this/

this analysis will be disregarded. The other biotites, except for those from G62 and G103, show a deficiency in the (OH + F) group, the amount present being often as low as $\frac{3}{4}$ of the theoretical quantity. This deficiency, particularly noticeable in the biotites from the metasediments, is also found in many previous analyses of biotites (for example, many of the analyses quoted by Foster (1960) and Deer, Howie, and Zussman (1962) show large deviations from the theoretical (OH + F) content). Such deviations may be due to (a) errors in analysis, including the expulsion of combined water at 110°C, (b) partial breakdown of biotite at high grades of metamorphism, or (c) weathering and hydrothermal activity. Of these causes, only the last is likely to cause OH results to be higher than the theoretical value. No significance can be attached to the lowness of the total (OH + F) occupancy in the present case, but it is interesting to note that 9 of the 13 biotites analysed by Lambert (1959) also have between 2.88 and 3.70 (OH + F) atoms per unit cell of 24 (O + OH + F).

The X (interlayer cation) sites are also under-occupied, the average occupancy being 92%. This is by no means unusual, and, though slight contamination by chlorite may be responsible for the very low value for the biotite from G103, the vacancy of some of the X sites must be regarded as an original feature of these biotites. Al^{iv} is fairly high, but not unusually so for the more iron-rich biotites and the Ti and Fe^{3+} contents are

quite/

quite normal, while the amount of Al^{vi} is, if anything, lower than in most biotites from metamorphic and granitic igneous rocks.

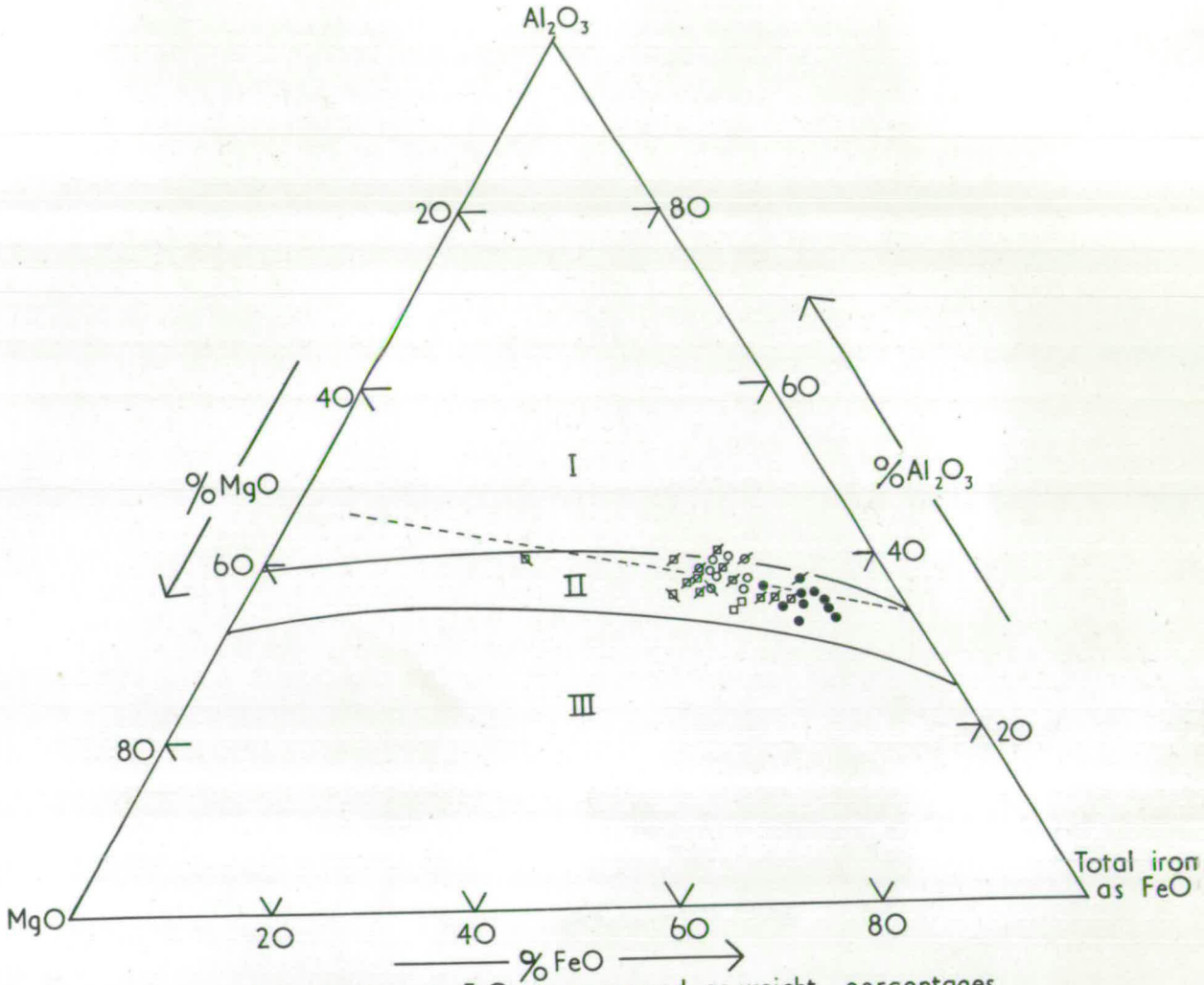
The biotite compositions may conveniently be plotted on a diagram of the type devised by Nockolds (1947), with the apices representing Al_2O_3, MgO and total Fe as FeO (Fig. 21). Nockolds showed that biotites coexisting with aluminous minerals such as muscovite and topaz were more aluminous, for any given Fe/Mg ratio, than biotites coexisting with no other "dark" mineral, which, in turn, were more aluminous than biotites coexisting with Al-poor ferromagnesian minerals such as pyroxenes, amphiboles, or olivine. As shown in Fig. 21, the Ardgour biotites conform only roughly to this rule. The biotites from G241 and SK211, rocks which contain muscovite, are only slightly more aluminous than those from the other rocks, while that from P291, which also contains muscovite, is exactly similar in Al content to the ordinary biotites. Fig. 21 does, however, bring out the higher Fe/Mg ratio and generally lower Al content of the granitic gneiss biotites, showing that there is very little overlap between the granitic gneiss and metasediment biotites. For comparison, the compositions of the biotite used in recalculating the norms of the Ardgour rocks (Table XIV, Appendix) and the biotites analysed by Lambert (1959) are also included in this diagram. The normative biotite is slightly more aluminous than the granitic gneiss/

FIGURE 21.

Plot of Al_2O_3, MgO and FeO for biotites from Ardgour and Knoydart.

FIGURE 22.

Plot of (Fe_2O_3 + TiO_2), MgO and (FeO + MnO) for biotites from Ardgour and Knoydart.



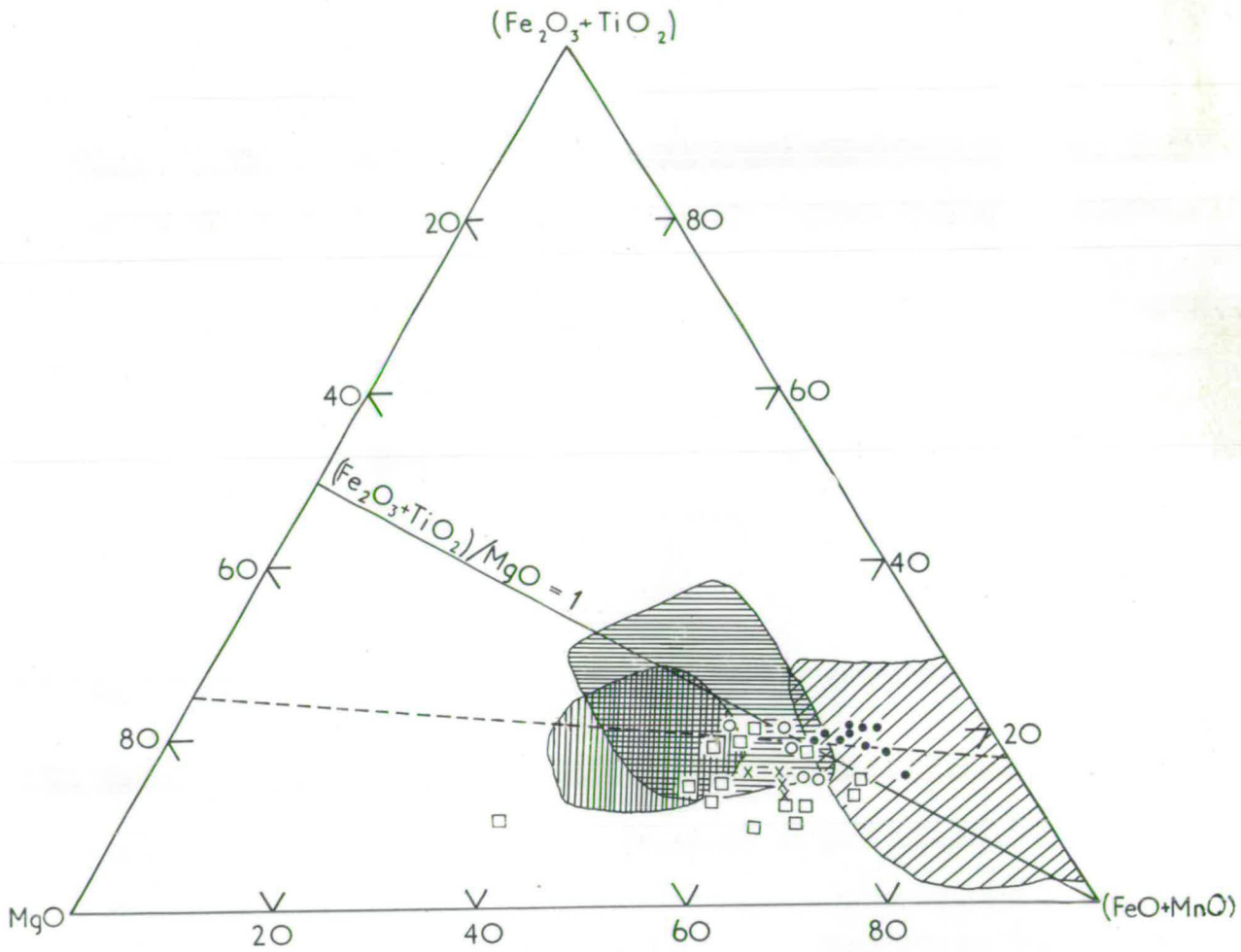
Al_2O_3, MgO, and FeO are expressed as weight percentages.

•,ø – Biotites from Ardgour granitic gneiss.
o,ø,ø – Biotites from Ardgour metasediments.
□,ø,ø – Biotites from Knoydart schists (Lambert 1959).
•,o,□ – Biotites coexisting with neither muscovite nor hornblende.
ø,ø,ø – Biotites coexisting with muscovite.
ø,ø – Biotites coexisting with hornblende.

I Field of biotites coexisting with muscovite or topaz.
II Field of biotites coexisting with no other "dark" mineral.
III Field of biotites coexisting with amphibole, pyroxene, or olivine.
} from Nockolds (1947).

The dashed line shows the biotite composition used in recalculating the norms of the analysed rocks (see Tables XIV–XV, Appendix).

FIG. 21.



$(Fe_2O_3+TiO_2)$, MgO, and (FeO+MnO) are expressed as weight percentages.

- • Biotite from Ardgour granitic gneiss.
- x Biotite from Ardgour pelite.
- o Biotite from Ardgour psammite.
- □ Biotite from Knoydart schist (Lambert 1959).
- Field of biotites from granite pegmatites.
- Field of biotites from granites, quartz-monzonites, and granodiorites.
- Field of biotites from gneisses and schists.

} from Heinrich (1946)

The dashed line shows the biotite composition used in recalculating the norms of the analysed rocks (see Tables XIV–XV Appendix).

FIG. 22.

gneiss biotites but slightly less aluminous than the metasediment biotites. This may help to account for some of the small amount of normative hornblende in the recalculated norm of the average granitic gneiss composition (given in Table 17, p. 66). Lambert's Knoydart biotites are for the most part similar to those from the Ardgour metasediments, but the biotites coexisting with neither muscovite nor hornblende are less aluminous than those from the Ardgour metasediments, and the band of compositions representing biotites coexisting with neither muscovite nor hornblende is very narrow. It should be remembered that Lambert's biotites come from an area of lower metamorphic grade than the Ardgour biotites. The correlation found by Lambert between the grade of metamorphism and the Mg content of the biotite was believed by him to be spurious and caused by changes in the bulk composition of the rocks; the trend is not continued in the Ardgour metasediment biotites.

An attempt to correlate the compositions of biotites with the rock types from which they came was made by Heinrich (1946), who plotted FeO, MgO, and (Fe_2O_3 + TiO_2) against each other. The attempt was only partially successful, however, with considerable overlap between the different fields. As shown in Fig. 22, the biotites from the Ardgour granitic gneiss plot in Heinrich's field of "biotites from granite pegmatites", while those from the metasediments plot in the field of "biotites from granodiorites, quartz monzonites, and granites". The diagram

does/

does, however, bring out the fact that the metasediment biotites are relatively poor in Fe^{3+} and Ti and have a $(Fe_2O_3 + TiO_2)/MgO$ ratio of less than 1, while for the granitic gneiss biotites this ratio exceeds unity. The normative biotite composition is also drawn on Fig. 22, and the fact that many pelites show negative quantities of normative ilmenite and magnetite (Table XV) is accounted for by the small amounts of Fe^{3+} and Ti in the biotites from these rocks. Lambert's (1959) biotites are generally poorer in Fe^{3+} and Ti than the Ardgour metasediment biotites but the amount of Mg is, on the average, only slightly higher. This does not seem to be due to the bulk composition of the rocks but is probably due to the difference in metamorphic grade. Butler (1965 b) has noted that Al^{iv} increases with metamorphic grade, and, to ensure electrical neutrality (Al^{vi} + Fe^{3+} + Ti) must also increase; the Knoydart biotites, which mostly have 2.4 - 2.6 Al^{iv} per 24 (O + OH + F), as against 2.6-2.9 Al^{iv} in the Ardgour biotites, have rather more Al^{vi} and must therefore contain less (Fe^{3+} + Ti) than the Ardgour biotites.

The principal changes in biotite composition in the series Knoydart pelitic and psammitic rocks - Ardgour metasediments - Ardgour granitic gneiss may be summarised as follows:-

(i) increase in $Fe^{2+}/(Fe^{2+} + Mg)$ ratio. In respect of this trend the Knoydart biotites overlap those from the Ardgour metasediments/

metasediments considerably.

(ii) decrease in Si and consequent increase in Al^{iv}. In this respect the Knoydart biotites are clarly separated from the other two groups, which show some overlap.

(iii) decrease in Al^{vi}, which is very marked. There appears to be a decrease in Al^{vi} within the Knoydart biotites with increasing grade (Lambert 1959 Table II). This trend continues in the Ardgour metasediments and culminates in the biotites of the Ardgour granitic gneiss, which contain the smallest amount of Al^{vi}. In biotites coexisting with muscovite Al^{vi} is generally slightly greater in amount than in biotites from muscovite-free rocks.

(iv) increase in $(Fe^{3+} + Ti)$. Ti is considerably higher in the Ardgour biotites than in those from Knoydart, but there is little difference between the Ardgour metasediment and granitic gneiss biotites. On the other hand, the Knoydart and Ardgour metasediment biotites have similar Fe^{3+} contents, the Knoydart biotites being, if anything, slightly richer in Fe^{3+}, while the Ardgour granitic gneiss biotites have considerably more Fe^{3+} than any of the other biotites considered.

Other respects in which the granitic gneiss biotites differ from the Ardgour metasediment biotites are: (a) the granitic gneiss biotites are, on the average, richer in F by a factor of about 1.4, (b) they are very slightly richer in Li (factor about 1.1). The proportions of Ca, Na, and K are similar in all of the/

the biotites analysed by Lambert or included in the present study, and there seems to be no relation between metamorphic grade and the proportions of Ca, Na, and K or the percentage of the X (12-coordinated) sites filled.

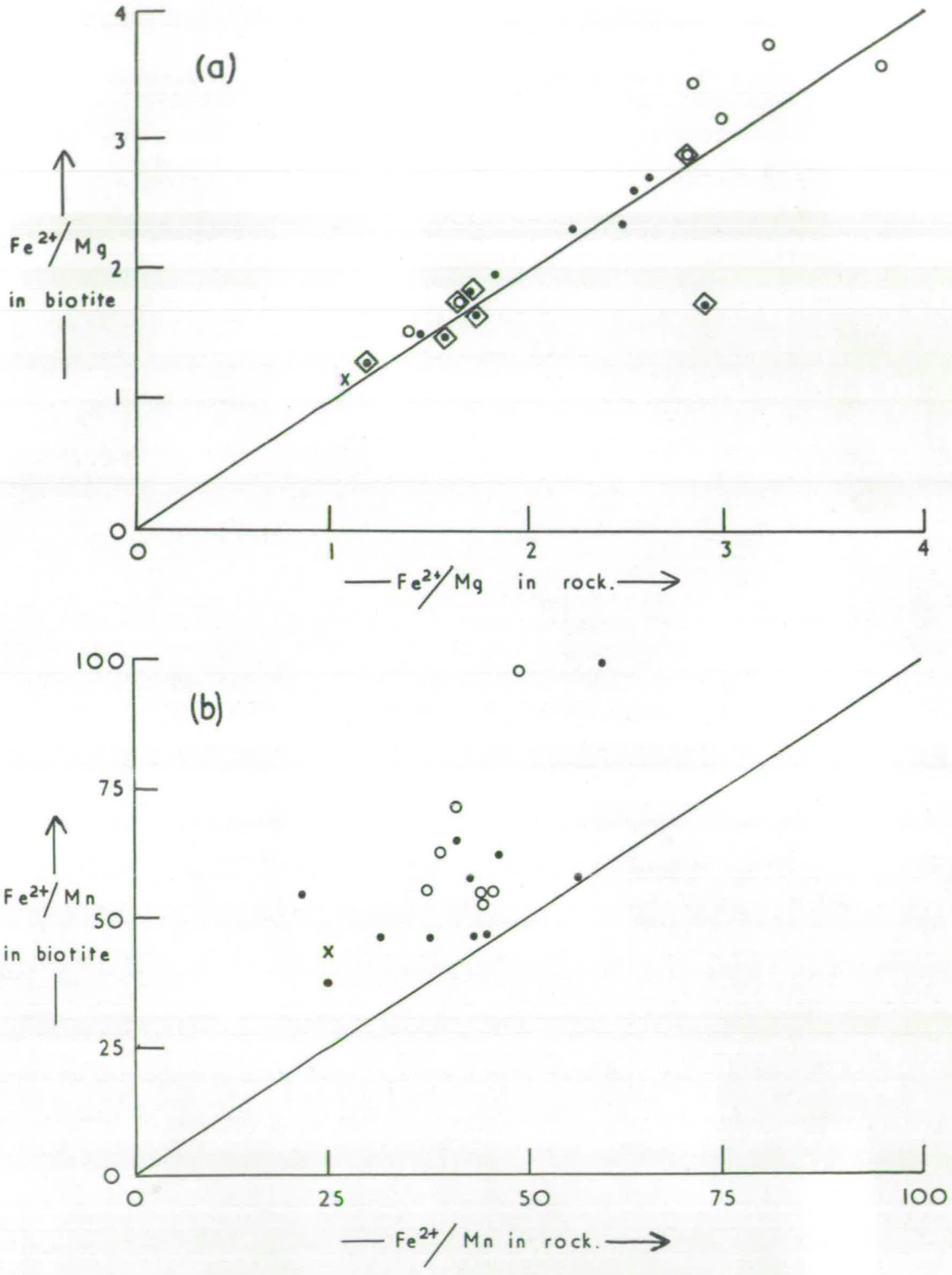
In all of the Ardgour rocks whose biotites were analysed, except P125, the amount of garnet if present is less than 2%. However, as shown in Fig. 23 b the biotites have consistently higher Fe^{2+}/Mn ratios than the rocks from which they come, and a similar relationship exists between the Fe^{2+}/Mn ratios of the rocks and the biotites which they contain, whether the rocks also contain garnet or not. This indicates that some mineral other than garnet or hornblende must take up manganese preferentially to biotite. Muscovite is unlikely to do so, because not only will it admit relatively small quantities of Mg and Fe, but it has been found (Butler, B. 1965 b) to have a lower $Fe^{2+}/(Fe^{2+} + Mg)$ ratio than coexisting biotite. It is possible, however, that the ore mineral of the rock may contain substantial quantities of Mn and have a lower Fe^{2+}/Mn ratio than the rock from which it comes.

Comparison of the Fe^{2+}/Mg ratios of the biotites with those of the rocks from which they come (Fig. 23 a) shows that despite the presence of an ore mineral and sometimes garnet in these rocks, the Fe^{2+}/Mg ratio of the biotites is usually slightly higher than that of the rocks from which they come. This effect is not attributable to the preferential concentration of/

FIGURE 23.

Comparison of Fe^{2+}/Mg and Fe^{2+}/Mn ratios of biotites with the same ratios in the rocks from which they come.

(a) Fe^{2+}/Mg; (b) Fe^{2+}/Mn; both ratios expressed molecularly.



• Rock containing neither garnet nor hornblende.
o Rock containing garnet.
x Rock containing hornblende.
◇ Rock containing muscovite (Fig. 23a only).

FIG. 23.

of Mg relative to Fe^{2+} in muscovite since only G241, P291 and SK211 have more than 1 - 2% of muscovite. The agreement between Fe^{2+}/Mg rock and Fe^{2+}/Mg biotite is usually fairly close, however, and the only case where there is a large difference in the ratios is S77. In this case Fe^{2+}/Mg rock exceeds Fe^{2+}/Mg biotite, and this can be explained if the rock has an abnormally high ore/biotite ratio. Whether the ratio ore/biotite is unusually high in this rock cannot be stated positively, especially since the chemical analysis is anomalous in yielding a norm with more hornblende than biotite, though biotite and ore are the only dark minerals in the rock.

2. ELEMENT DISTRIBUTIONS BETWEEN COEXISTING MINERALS.

(a) General considerations: The distribution of an element between two phases both of which can accept it depends on the energies of the bonds which an ion of the element in question forms when it becomes incorporated into the competing crystal lattice sites. The bond energies vary for ions of different elements, and Goldschmidt (1937), Ahrens (1953) and Ringwood (1955) have shown the importance of ionic radius, ionic charge, and the relative proportion of ionic and covalent bonding (expressed quantitatively as ionisation potential or electronegativity) in determining the crystal lattice sites into which ions of an element are likely to be accepted.

Even/

Even the bond energies for the same element, however, are dependent upon many factors which may affect the distribution of the element between coexisting phases. The chief of these are:

(i) Temperature.

(ii) Pressure.

(iii) Variations in the crystal structure of the minerals in question.

(iv) Variations in the amount of some other element in the minerals in questions.

Kretz (1961) showed that if p and T remain constant and the minerals studied behave as ideal solid solutions, the equilibrium value of the ratio $\frac{X_A^R}{1-X_A^R} \cdot \frac{1-X_B^R}{X_B^R}$,

where $X_A^R = \frac{\text{Number of lattice sites in mineral A filled by R}}{\text{Number of lattice sites in A capable of accommodating R}}$,

remained constant, and that at small values of X_A^R and X_B^R this reduced to X_A^R / X_B^R = constant. This last is a statement of Nernst's distribution law, and $X_A^R / X_B^R = K_{A-B}^R$ is known as the distribution coefficient.

Graphical plots of X_A^R against X_B^R can provide quite a good guide to the distribution of trace elements within isofacial equilibrium assemblages, and in fact the closeness of fit to Nernst's distribution law can often be a good indication of whether equilibrium has been attained. Even here, however, care should be taken to see that, if the mineral assemblage differs in the rocks studied, none of the minerals whose trace element content is/

is being investigated differ in composition in a systematic way depending on the assemblage; for example, if it is found that biotites coexisting with muscovite have consistently more Al^{vi} than biotites coexisting with hornblende, this may cause systematic differences in the distribution of, say, Ti between coexisting biotite and garnet. Similarly, if a mineral shows gross variation in composition due to variation in the bulk composition of the rock or magma from which it is derived, this may affect the distribution of some other elements; for example, the distribution of Sr between albite and hornblende and between anorthite and hornblende would not be expected to be the same even if the two assemblages were in equilibrium at the same p and T.

Finally, it may be difficult to decide the exact value of the expression (number of lattice sites in A capable of accommodating R) in certain cases; for example, whether Sc in clinopyroxenes is capable of entering only Ca sites, or only $(Mg + Fe^{2+})$ sites, or possibly both, is not yet properly understood.

When the variation of the distribution coefficient with p and T is considered, the likelihood of ideal behaviour of natural mineral solid solutions becomes less. Ramberg and de Vore (1951), and Barth (1951, 1956) considered the application of thermodynamics to the variation of the distribution coefficient with temperature and, in the case of Ramberg and de Vore, with pressure/

pressure also, while Kretz (1961) in addition considered the effect of variations in the content of other elements in the minerals studied.

Ramberg and de Vore (1951), working on the distribution of Mg and Fe^{2+} between coexisting olivine and orthopyroxene, derived the equation $$\frac{X_{hyp}^{Mg}}{1 - X_{hyp}^{Mg}} \cdot \frac{1 - X_{ol}^{Mg}}{X_{ol}^{Mg}} = e^{\frac{-\Delta F}{RT}}$$ where T = absolute temperature, R = gas constant and ΔF = free energy change of the reaction

$$Mg_2SiO_4 + 2FeSiO_3 \rightarrow Fe_2SiO_4 + 2MgSiO_3,$$

and also considered the effect of pressure on the equilibrium distribution, deriving the equation

$$K = K' . e^{\frac{\Delta V (p' - p)}{2RT}}$$

where K and K' are the distribution coefficients at pressures p and p' respectively and ΔV is the volume change of a reaction of the type

$$Mg_2SiO_4 + 2FeSiO_3 \rightarrow Fe_2SiO_4 + 2MgSiO_3.$$

Calculations based on known densities of orthopyroxenes and olivines showed, however, that for the above reaction, the theoretical distribution coefficient only varied by a factor of 0.87 for a pressure difference of 10 Kb. Analyses of natural olivine-orthopyroxene pairs showed a decrease in distribution coefficient with increasing $Fe^{2+}/(Fe^{2+} + Mg)$ of the system. This was ascribed to differences in the temperature of crystallisation. The wide range of Al contents in the orthopyroxenes was noted, but/

but even when only Al- poor orthopyroxenes were considered, the distribution coefficient did not show any greater approach to constancy.

Barth (1951) derived an equivalent formula to that of Ramberg and de Vore for the temperature dependence of the distribution coefficient, and (1956) tried to apply his results to the distribution of Na between coexisting natural felspar pairs. Using a plot of Log_eK against $1/T^oA$, three straight lines were drawn empirically, each representing a different structural state in potash felspars. Not only has the non-ideality of the system $NaAlSi_3O_8$ - $KAlSi_3O_8$ been demonstrated by the experimentally determined solvus curves of Bowen and Tuttle (1950), Yoder, Stewart and Smith (1957), and Orville (1963), but the expectation that the Si-Al substitution taking place in plagioclases with increasing Ca content would have no effect on the ease of entry of Na into the plagioclase lattice is unjustified.

Kretz, working (1960) on the distribution of Mg and Fe^{2+} between clinopyroxene, garnet, and amphibole in skarns, and (1961) on the distribution of these elements between garnet, clinopyroxene, orthopyroxene and olivine in various rocks, tried to relate differences in the distribution coefficient to pressure and temperature differences during crystallisation. His interpretation of his results as far as they affect peridotites and olivine nodules from basalts has been severely criticised by O'Hara and Mercy (1963), who demonstrated that changes of distribution/

tribution coefficient attributed by Kretz (1961) to pressure differences could be due to variations in the Al content of the pyroxene phases.

Hence it must be concluded that differences in the distribution of an element between a pair of minerals can only be used to estimate the conditions of crystallisation of the rocks studied if the minerals considered have been studied experimentally from this standpoint and occur as constituents of the same mineral assemblage. Even then, great care must be exercised, as was shown by Ramberg and de Vore (1951) when comparing natural with experimental olivine-orthopyroxene pairs.

(b) The coexisting biotite and hornblende from S172: The rock S172 consists of alternating calcareous and semipelitic bands (see p.27). Consideration of the hornblende analysis and its relation to that of the biotite from the same rock is thus rendered more difficult by doubts as to whether equilibrium has been attained. The biotite is confined to the semipelitic bands of the rock and the hornblende to the calcareous bands; though the plagioclase composition appears to be An_{70} in the calcareous and An_{50} in the semipelitic bands, both these compositions probably reflect a considerable change in Na/Ca ratio from the expected ratios in the original sediments. It is thus reasonable to suppose that even if equilibrium between biotite and hornblende has not been achieved, the minerals have undergone some change in composition resulting in a closer approach to equilibrium/

equilibrium distribution of the elements between the calcareous and semipelitic bands.

The compositions of coexisting biotite and hornblende from high-grade metamorphic rocks have been studied by Kretz (1959). He found that Ti, Mg, Fe (total) and Mn (the last at low concentrations only) had roughly constant distribution coefficients in 14 analysed mineral pairs. The average distribution coefficients found by Kretz are listed in Table 28 where they are compared with the distribution coefficients for the Ardgour biotite-hornblende pair. The distribution coefficients for Ti, Mg, and $(Fe^{2+} + Fe^{3+})$ in S172 are very similar to those found by Kretz. The distribution coefficient for Mn was found by Kretz to be constant only when $Mn/(Mn + Mg + Fe^{2+} + Fe^{3+} + Ti)$ was less than 0.009 in the hornblende, and to fall off at higher levels of Mn concentration. The Ardgour pair contain more Mn than any studied by Kretz and there is no reason to suppose that the distribution coefficient found could not represent the equilibrium distribution of Mn.

It is unfortunate that the minerals studied by Kretz (1959) were not analysed for Fe^{2+}, as this would have enabled proper mineral formulae to be calculated. It might then have been possible to decide whether to regard Ti, Fe^{3+} and Al^{vi} as a separate group from Mg, Mn and Fe^{2+} for the purposes of calculating distribution coefficients, or whether to regard all elements occupying 6-coordinated sites in the mica and amphibole lattices/

TABLE 28.

ELEMENT distribution between Hornblende and Biotite.

	From S172		S172	Avge dist.coeff.
	Biotite	Hornblende	Dist.coeff Horn/Biot	Horn/Biot fd.By Kretz (1959)
$Mn/(Mg+Fe^{2+}+Fe^{3+}+Ti+Mn)$	0.0111	0.0248	2.23	3.2 *
$Ti/(Mg+Fe^{2+}+Fe^{3+}+Ti+Mn)$	0.0785	0.0327	0.417	0.50
$Mg/(Mg+Fe^{2+}+Fe^{3+}+Ti+Mn)$	0.381	0.388	1.02	1.00
$(Fe^{2+}+Fe^{3+})/(Mg+Fe^{2+}+Fe^{3+}+Ti+Mn)$	0.530	0.554	1.04	1.12
$Fe^{2+}/(Mg+Fe^{2+}+Fe^{3+}+Ti+Mn)$	0.445	0.383	0.861	-
$Fe^{3+}/(Mg+Fe^{2+}+Fe^{3+}+Ti+Mn)$	0.0847	0.171	2.02	-
Li/Total Y group cations	0.0185	0.0073	0.395	-
F/(OH+F)	0.0481	0.0708	1.45	-
Rb/X group cations	0.00193	0.000050	0.0259	-
Ba/X group cations	0.00958	0.000099	0.0103	-
Sr/X group cations	0.000125	.000219	1.75	-
Rb/K	0.00212	0.000556	0.262	-
Ba/K	0.0106	0.00110	0.104	-
Sr/Ca	0.00239	0.000289	0.121	-
Sr/Na	0.00305	0.00141	0.462	-

* Wide divergence from this ratio was found when $X_{Mn}^{Horn} > 0.009$ and $X_{Mn}^{Biot} > 0.003$.

All ratios in terms of cations (or anions).

lattices as potentially equivalent.

The distribution ratios of the various elements between biotite and hornblende might be expected to vary with metamorphic grade (that is, p and T conditions.) The metamorphic grade of the Grenville gneisses studied by Kretz (1959) differs slightly from that of the Ardgour rocks; the Grenville gneisses belong to the sillimanite-almandine-orthoclase subfacies of the almandine-amphibolite facies, whereas the Ardgour rocks probably belong to the sillimanite-almandine-muscovite subfacies of that facies (see pp.160-167 for a discussion of the metamorphic grade of the Ardgour rocks.) The Grenville gneisses therefore represent a slightly higher temperature or a lower p_{H_2O} than the Ardgour rocks, the total load pressure remaining similar. The similarity of the distribution coefficients $K^{Ti}_{Horn-Bi}$, $K^{Mg}_{Horn-Bi}$ and $K^{(Fe^{2+} + Fe^{3+})}_{Horn-Bi}$ between the Ardgour biotite-hornblende pair and those from Kretz (1959), which were presumed from textural evidence to be likely equilibrium assemblages, suggests strongly, although it does not prove, that the hornblende and biotite from S172 crystallised in equilibrium with each other. The difference in plagioclase compositions between the different bands in the rock need not indicate lack of equilibrium because the mineral composition of the bands differs so much, and the fact that the higher Mn content of the calcareous band parallels an original sedimentary composition difference does not necessarily mean that equilibrium between biotite and hornblende has not been attained/

attained as far as Mn is concerned.

If it is assumed that the biotite and hornblende from S172 are in equilibrium, it becomes worthwhile to work out distribution coefficients for various trace elements. These results are also shown in Table 28. As expected, the distribution of Li shows a marked preference for biotite, but although the amount of F in the biotite exceeds that in the amphibole, the amphibole has the higher F/OH ratio. The F/OH ratio is even lower in muscovite than in biotite, the $X^{F}_{Biot} / X^{F}_{Musc}$ ratios for the rocks G241, P291 and SK211 being 2.95, 4.86, and 2.85 respectively. As would be expected, Rb and Ba show a strong preference for biotite compared with hornblende, whether their concentration is compared with the total X group occupancy or with the K concentration alone. The amount of Sr in the hornblende is higher than in the biotite, as expected, but in proportion to the Ca content of the two minerals the amount of Sr in the hornblende is remarkably low.

(c) Felspar thermometry: The conditions mentioned above (p.115) as being necessary for differences in the distribution coefficient of an element between coexisting minerals to be used as a guide to the conditions of crystallisation of a series of rocks appear to be fulfilled to a greater extent in the felspars than in most other mineral pairs. Experimental solvi have been determined in the system $NaAlSi_3O_8$ - $KAlSi_3O_8$ at p_{H_2O} = 1000 bars (Bowen and Tuttle 1950), at p_{H_2O} = 5000 bars (Yoder, Stewart/

Stewart and Smith 1957) and at p_{vapour} = 2000 bars in the presence of KCl and NaCl vapour (Orville 1963). Unfortunately only Yoder, Stewart and Smith (1957) considered the effect of varying quantities of Ca on the solvus, and it is difficult to calculate distribution coefficients from their published diagram, but, since a pressure of p_{H_2O} = 5000 bars was considered to give the closest approximation to the likely conditions of formation of the Ardgour rocks, the Yoder, Stewart, and Smith (1957) solvus was used as a basis for drawing distribution curves for Ca, Na, and K between coexisting felspar pairs.

The distribution coefficient, C_R, of an element R was defined as $C_R = \frac{\text{molecular proportion of R in plagioclase}}{\text{molecular proportion of R in potash felspar}}$ ($= K^{R}_{plag-Kfels}$ in nomenclature of Kretz (1961)) while the symbol X was taken to mean $X^{Ca}_{plag} = \frac{\text{mols. An in plagioclase}}{\text{mols. (An+Ab) in plagioclase}}$ unless otherwise specified. Curves showing the distribution of Na between plagioclase and potash felspar were drawn directly from the Yoder, Stewart and Smith (1957, Figs. 38, 44, 45) solvus for X = 0, 20 and 40. A curves showing the variation of C_K for X = 0 was also drawn, but curves for C_{Ca} and C_K for X = 20, 40 and 100 could only be sketched in roughly because of the difficulty of reading the published diagrams and because of the difficulties experienced by the experimental workers in determining the position of the solvus in certain places. For this reason temperatures of crystallisation of the Ardgour rocks were determined/

FIGURE 24.

Curves showing the distribution of Ca, Na and K between coexisting felspars at different temperatures. Drawn for p_{H_2O} = 5000 bars.

$$C_R X = \frac{\text{Molecular proportion of R in plagioclase}}{\text{Molecular proportion of R in potash felspar}}$$ when the molecular percentage of anorthite in the plagioclase is X (see p.119).

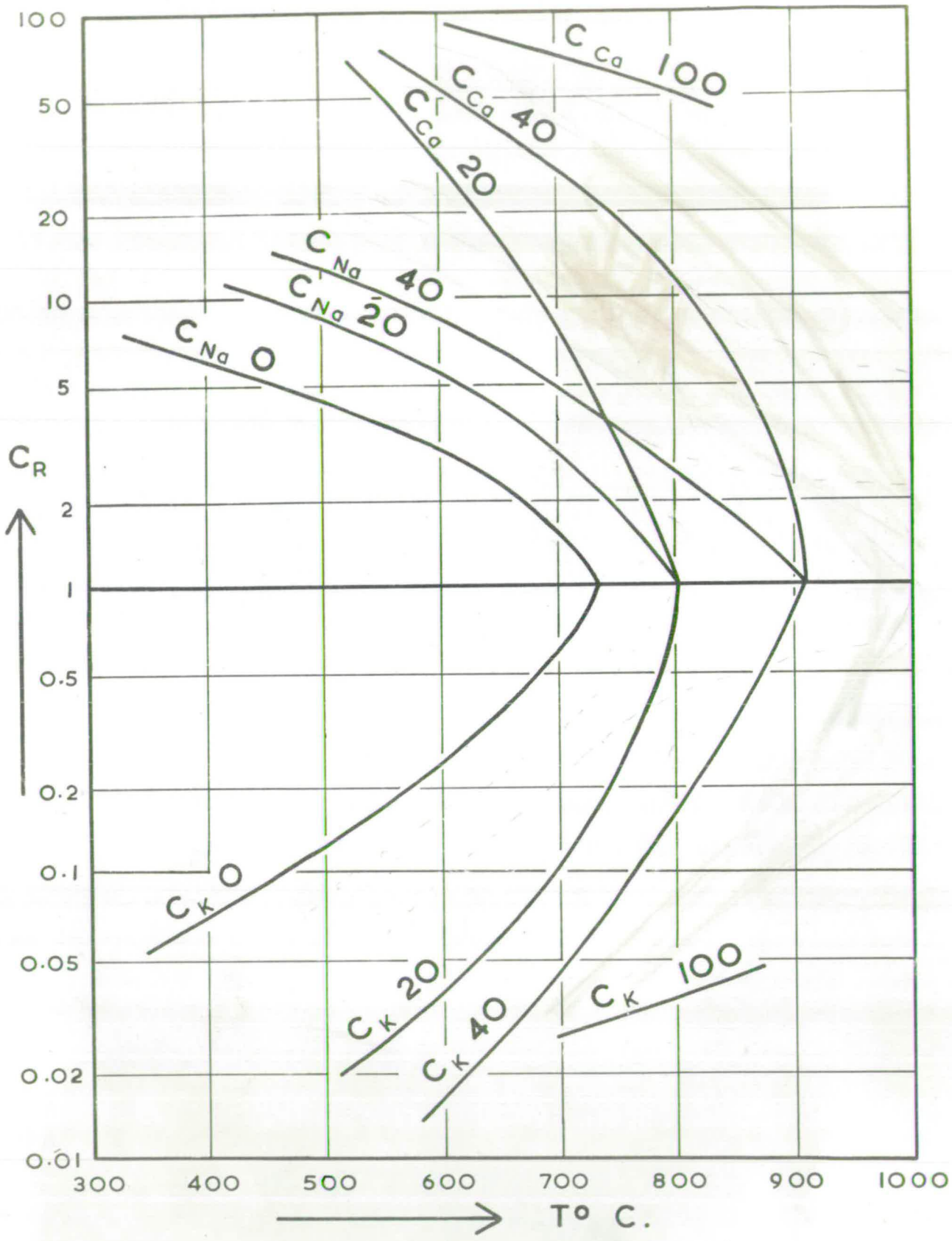

100
50
20
10
5
2
1
0.5
0.2
0.1
0.05
0.02
0.01
C_R
300
400
500
600
700
800
900
1000
T° C.
C_{Ca} 100
C_{Ca} 40
C_{Ca} 20
C_{Na} 40
C_{Na} 20
C_{Na} 0
C_K 0
C_K 20
C_K 40
C_K 100

FIG. 24.

determined using the distribution of Na only between the felspars, and the values of C_{Ca} and C_K for the Ardgour felspar pairs were plotted against T_{Na} (the temperature obtained from the Na distribution) in order to fix more accurately the positions of the C_{Ca} and C_K curves for X = 20. A similar technique using the analyses of coexisting felspars from two charnockites from Madras (Howie, 1955) was applied to the fixing of the C_{Ca} and C_K curves for X = 40.

The results for the Ardgour and Madras felspar pairs are listed in Table 29 and the curves as finally set up are shown in Fig. 24. The consistency as temperature estimators of C_{Na}, C_{Ca}, and C_K is shown by comparing the actual values of C_K and C_{Ca} for the Ardgour and Madras rocks with those predicted using the final curves for C_K and C_{Ca} (Table 29). As an independent check, the values of C_{Na}, C_{Ca}, and C_K for 8 felsparppairs from Langøy (Lofoten Is., Norway) (Heier 1960) were plotted on Fig. 24, and the resulting temperatures compared with each other (Table 30). The temperatures found from the three curves show a spread for any one rock of 35 - 110°C. Thus the precision of the method is not very good, and the temperatures derived from the Ca, Na, and K distributions are probably individually subject to errors of ± 50°C, while the mean of the three is probably subject to an error of ± 30°C. This is combined with an inaccuracy of unknown magnitude resulting from possible errors in the transcription from Yoder, Stewart and Smith (1957) of the values from which the C_{Na} curves for X = 20 and 40 were constructed.

The/

TABLE 29.

Values of C_{Ca}, C_{Na} and C_K for the analysed felspar pairs used to modify the curves for C_{Ca} and C_K in Fig.24.

Rock Type	Rock No.	X	C_{Na}	T_{Na}	C_{Ca} actual	C_{Ca} predicted	C_K actual	C_K predicted
Granitic Gneisses	G 12	21.9	5.3	620	35	27	0.045	0.040
	G 13	20.6	3.4	685	10	11	0.032	0.079
	G 62	20.2	4.8	635	24	21	0.043	0.051
	G 92	20.4	5.7	600	33	32	0.036	0.037
	G103	20.4	3.0	700	10	8.5	0.040	0.105
	G163	20.3	6.1	585	39	39	0.036	0.034
	G183	18.7	6.2	580	46	38	0.031	0.036
	G241	18.5	6.4	570	60	41	0.025	0.032
	G262	19.8	3.1	700	15	7.4	0.107	0.105
	G281	21.1	5.5	605	25	31	0.043	0.038
Psammites	S 42	17.7	6.3	570	34	40	0.033	0.033
	S 77	18.3	4.8	615	43	34	0.081	0.044
	S141	22.6	5.0	630	21	24	0.068	0.046
	SK211	14.7	6.4	545	29	42	0.028	0.029
"Pelite"	P192	15.7	6.2	555	26	41	0.026	0.031
Charnockites *	2270	33.7	4.6	685	24	21	0.076	0.052
	6436	32.7	3.8	720	31	16	0.023	0.072

X = Mol. % An in plagioclase.

T_{Na} = Temperature derived from the distribution of Na between coexisting felspars, using the curve for Na distribution based on experimentally determined solvi. (in °C.)

C_{Ca}, C_K predicted = Points of intersection of the modified C_{Ca} and C_K curves with the ordinate corresponding to T_{Na}.

* Taken from Howie (1955).

TABLE 30.

Values of C_{Ca}, C_{Na}, C_K, T_{Ca}, T_{Na}, and T_K for the felspar pairs analysed by Heier (1960) and used to check the accuracy and precision of the curves set up using experimental data and the results shown in Table 29.

Rock Type	Rock No.	X	C_{Ca}	C_{Na}	C_K	T_{Ca}	T_{Na}	T_K
Porphyroblastic	293/55	30.3	7.3	3.0	0.106	770	740	735
Monzonitic	251/55	25.0	5.5	3.4	0.066	760	710	685
Granulites								
Retrograde Metamorphosed	238/55	38.8	16	2.5	0.049	775	800	715
gneisses	224/55	28.8	13	3.6	0.047	720	705	670
Banded series	an.12	31.9	8.3	4.0	0.118	770	700	745
Charnockitic border series	25/55	29.9	19	3.5	0.113	695	715	735
Amphibolite facies	37/55	24.8	14	5.1	0.076	695	645	695
veined gneisses	An. 1	25.2	17	7.3	0.061	685	575	670

Rock No.	T average.	Temp by Barth Method.
293/55	750	590
251/55	720	585
238/55	765	605
224/55	700	560
an.12	740	510
25/55	715	545
37/55	680	465
an. 1	640	400

X = Mol.% An in plagioclase.

T_{Ca}, T_{Na}, T_K = Temperatures derived from the distribution of Ca, Na and K respectively between the two felspar phases using the graphs of Fig.24.

$T_{Average}$ = Average of T_{Ca}, T_{Na} and T_K.

The sources of error to which the results from the Ardgour granitic gneiss are subject are:

(i) Analytical error in Na, Ca and K determinations.

(ii) Errors in the experimental work of Yoder, Stewart and Smith (1957) and errors in drawing the Na distribution curves from the published diagrams.

(iii) Exsolution in the perthites may have reached the stage when the potash felspar could be selectively altered in composition during purification of the mineral separation prior to analysis.

(iv) The temperature at which equilibrium was last attained may not have been the highest reached by the rock.

(v) The high Sr and Ba contents of G103 may have affected the distribution of Na between the felspars, thus accounting for the higher temperature estimate obtained for this rock.

As shown in Table 29, the average temperature obtained for the granitic gneiss is 630°C(range 570°C to 700°C) while that for the metasediments is 575°C (range 545°C to 630°C). The temperatures found are quite compatible with the mineral assemblages present in the rocks. The reaction

$$KAl_3Si_3O_{10}(OH)_2 \rightarrow KAlSi_3O_8 + Al_2O_3 + H_2O \quad \ldots\ldots(1)$$

has been found (Velde 1964 p. 141) to take place at temperatures of 700 - 720°C for pressures from 3 to 8 Kb. The reaction

$$KAl_3Si_3O_{10}(OH)_2 + SiO_2 \rightarrow KAlSi_3O_8 + Al_2SiO_5 + H_2O \quad \ldots\ldots(2)$$

has/

has not been investigated experimentally, but by analogy with the breakdown curves for biotite, the curve for this reaction is probably parallel to that for reaction (1) above, lying 10 - 40°C below it. On the basis of experimental work on the breakdown of pyrophyllite, Turner and Verkoogen (1960, p. 552) consider that, if water vapour pressure remains equal or nearly equal to load pressure, the almandine-amphibolite facies of regional metamorphism cannot be attained at temperatures less than 550°C. Thus the mineral assemblages in the Ardgour rocks probably formed at temperatures between 550°C and 690°C, and this is confirmed by the results obtained from the distribution of Na between coexisting felspars.

The metasediments show lower temperatures than the granitic gneisses, and, though there is some overlap, this difference is believed to be real. It is unlikely that the difference could be related to the presence or absence of muscovite in the rocks, since samples G163 and G183, which lack primary muscovite, give temperatures not significantly different from G241 which contains muscovite. Whereas the temperatures of crystallisation of the metasediments were almost certainly too low to allow partial melting to have taken place, the granitic gneisses probably crystallised at temperatures close to those established experimentally (see p. 58) as the lowest at which partial melting could have occurred (660°C).

The felspar pairs from a South Norwegian migmatite complex studied/

studied by Barth (1956) can be compared with the Ardgour felspar pairs if the temperatures are recalculated using the Na distribution curves of Fig. 24. This is done in Table 31. Assuming an error of ± 50°C and a temperature of 660°C for the beginning of anatexis under the conditions prevailing during the crystallisation of these rocks, the new temperatures would indicate that the ordinary gneisses (1, 2) and the aplitic granite (17) did not undergo any partial melting, while the augen gneisses (4 - 8), a small pegmatite dyke (10), and most of the large pegmatite dykes crystallised from a melt, and the other rocks were too close to 660°C to enable a decision whether partial melting had occurred to be made. The low temperature for the aplitic granite (17) is anomalous, and the high temperatures for the augen gneisses (4 - 8) may be in error due to disequilibrium between potash felspar porphyroblasts and groundmass plagioclase. Otherwise, the new temperatures are at least as consistent with the temperature indices from calc-silicates, etc. (Barth 1956, p. 32), as the temperatures given by Barth's method, and allow for the formation of the granitic migmatites of South Norway by reaction with or intrusion by an anatectic melt.

The absolute values for the crystallisation temperatures of the felspar pairs of Heier (1960), listed in Table 30, show that the curves of Fig. 24 place the transition between granulite and amphibolite facies on Langøy at 680 - 700°C, whereas Barth's thermometer gives 500°C for this transition.

The/

TABLE 31.

Comparison of the present estimates of temperature of crystallisation of rocks from the compositions of coexisting felspar pairs with the results obtained by Barth (1956) for certain South Norwegian rocks. Analyses from Barth (1956).

Rock Type	Rock No.	X	C_{Na}	T_{Na} (°C)	Temperature given by Barth.
Ordinary gneisses	1	25	8.1	530	390
	2	23	7.5	560	400
Coarse gneiss	3	28	6.1	630	430
Augen gneisses	4	29	2.6	750	640
	5	20	2.4	725	680
	6	29	3.2	730	580
	7	25	2.6	740	640
	8	22	2.5	730	660
Small pegmatite dykes	9	17	3.6	665	550
	10	16	3.4	710	560
	11	20	3.0	700	590
Anatectic granites	12	30	6.1	625	430
	13	29	5.9	635	450
	14	24	5.4	630	470
	15	30	5.8	630	450
Aplite vein	16	27	6.4	610	430
Aplitic granite	17	25	7.5	560	400
Diapiric granites	18	12	3.8	625	550
	19	19	3.6	670	550
	20	18	3.8	670	550
Large pegmatite dykes	21	8	3.1	640	590
	22	39	3.0	770	600
	23	28	2.9	740	610
	24	22	2.8	720	620
	25	6	2.8	630	620

T_{Na} = Temperature obtained from distribution of Na between coexisting felspar phases, using the curves of Fig.24.

X = Mol. % An in plagioclase.

The temperature of transition from amphibolite to granulite facies is determined by the breakdown of biotite and hornblende. The breakdown temperature of biotite has been shown (Yoder and Eugster 1954, Eugster 1956) to depend on its Fe/(Fe + Mg) ratio, p_{H_2O} and p_{O_2}, and the same is probably true of hornblende (see Boyd 1955 for the stability range of magnesian pargasite). The granulite facies rocks studied by Heier included some with the correct proportions of Q, Or and Ab to have formed an anatectic melt at 660 - 670°C and p_{H_2O} = 5000 bars. However, no liquid phase developed in these rocks. Thus either the total pressure was considerably less than 5000 bars or p_{H_2O} was less than p_{load}. In either case it is difficult to estimate accurately the temperature corresponding to the boundary between the granulite and amphibolite faces; the good agreement with outside evidence obtained for the temperatures derived from Fig. 24 in the case of the Ardgour and South Norwegian rocks make it likely that the Langøy granulites crystallised at temperatures of 700 - 800°C, in accordance with the views of Turner and Verhoogen (1960, p. 557) on the temperatures attained in granulite facies metamorphism, but the fact that these rocks belong to the granulite facies cannot be used to corroborate the temperatures of 700 - 765°C obtained for them by the new method.

(d) <u>The distribution of Rb, Ba and Sr between coexisting micas and felspars</u>: The laws governing the entry of trace elements into crystal lattices have been discussed by Goldschmidt/

Goldschmidt (1937), Ahrens (1953), and Ringwood (1955). These workers were concerned principally in accounting for magmatic differentiation trends. In the recrystallisation of metamorphic rocks the reconstitution of minerals to form equilibrium assemblages often involves the partition of a trace element between competing lattice sites in two or more mineral species. The bulk composition of the rock will decide the absolute major/minor element ratios in the constituent minerals, but will have little or no effect on the equilibrium value of the distribution coefficient of an element between two coexisting minerals, provided that the mineral assemblage in the rock does not alter. The distribution coefficient will be the same as if the competing minerals had crystallised from a liquid of the same original bulk composition as the final rock and been allowed to reach equilibrium at the same pressure and temperature as those at which the metamorphic rock recrystallised.

In actual fact, metamorphic rocks, unless they show evidence of late-stage retrogressive metamorphism, usually show a closer approach to equilibrium than do igneous rocks, which often display textural features, such as mineral zoning, indicative of disequilibrium crystallisation. Ideally the equilibrium distribution of an element between a crystal phase and a liquid or between two crystal phases should be fixed for a given temperature and pressure, but in fact may be expected to vary if one of the competing phases varies in composition for any reason. For example, in the fractionation trend of the

Skaergaard/

Skaergaard intrusion (Wager and Mitchell 1952), the ratio Sr in plagioclase/Sr in liquid increases from 2.5 to 12 and then decreases to 5. Whether this is due to changes in the felspar composition or in that of the liquid or is a temperature effect is not known.

The trace elements considered here, Rb, Ba, and Sr are all known to have close geochemical affinities with K, Na and Ca. They all have low valency, a large ionic radius, and a low ionic potential, and therefore have no tendency to form an MO_4 complex in aqueous or silicate fluid phases. In addition they have very low ionisation potentials and electronegativities (see Table 32, data taken from Ahrens (1952 and 1964)) and hence their bonding when they enter silicate lattices is almost entirely ionic. Their distribution between competing minerals is therefore controlled by the size and charge of their stable cations. The contrast with Pb^{2+}, which forms more covalent bonds and behaves as if it were smaller than it is, is great; Heier (1960) found that in amphibolite and granulite facies rocks from Langøy (Lofoten Is., Norway) the ratio Pb plagioclase/ Pb potash felspar averaged 2, yet Pb^{2+} has an ionic radius of 1.20Å (Table 32), larger than that of Sr^{2+} whose radius is 1.12Å, but which was found to prefer the potash felspar phase in these rocks, Sr plagioclase/Sr potash felspar averaging about 0.75.

It has been previously assumed (Wager and Mitchell 1952) that/

TABLE 32.

Ionic radii and ionisation potentials for certain elements, taken from Ahrens (1952). Electronegativities (Little & Jones, 1960, quoted by Ahrens, 1964).

Element	Ion	r	I	F	Electronegativity
Cs	Cs^{+}	1.67	3.9	2.34	0.86
Rb	Rb^{+}	1.47	4.2	2.86	0.89
Tl	Tl^{+}	1.47	6.1	4.15	1.44
K	K^{+}	1.33	4.34	3.26	0.91
Na	Na^{+}	0.97	5.14	5.30	1.01
Ba	Ba^{2+}	1.34	10.0	7.46	0.97
Pb	Pb^{2+}	1.20	15.05	12.54	1.55
Sr	Sr^{2+}	1.12	11.03	9.85	0.99
Ca	Ca^{2+}	0.99	11.9	12.02	1.04

I = ionisation potential (eV) (first for univalent cations, second for divalent cations).

r = ionic radius (Å).

F = I/r .

that when a trace element enters a site in a mineral which is shared between atoms of two or more major elements, the trace element should be regarded as being camouflaged by only one of the major elements; for example, Ni^{2+} entering the site in olivines occupied by Mg^{2+} and Fe^{2+} is regarded (Ringwood 1955) as being camouflaged by Fe^{2+} rather than by Mg^{2+}. The reasons for this have not been stated explicitly, however, and in some cases the element with which a trace element is associated may vary in different minerals. This does not apply to Rb or Ba, which always follow K, but Sr has always been regarded as replacing Ca in apatite and plagioclase and K in potash felspar. This problem will be discussed when the distribution of Sr is considered individually (pp. 134 - 139).

The determinations of Rb, Ba, and Sr in the separated minerals were done by X-ray fluorescence spectroscopy (pp. 51 - 54). This method is probably more precise than optical spectrography for values over 400 ppm Ba, 100 ppm Sr or 50 ppm Rb, but is relatively insensitive for low concentrations of these elements. Hence the results for Rb and Ba in plagioclases and hornblende and Sr in biotites and muscovites are considerably less precise than those for Rb and Ba in micas and potash felspars and Sr in potash felspars and plagioclases, and only the distribution coefficients for Rb and Ba between biotite, muscovite and potash felspar, and for Sr between potash felspar and plagioclase can be determined accurately.

Before/

Before the three elements are discussed individually, the relation between the Rb, Ba, and Sr contents of the separated minerals and those of the rocks from which they come will be considered. In the granitic gneiss, the Rb, Ba, and Sr contents of the minerals separated from all of the rocks except G103 are amazingly similar (see Tables XVII, XIX and XX, Appendix). This suggests that the process responsible for the formation of the granitic gneiss was able to effect something approaching equilibrium distribution of Rb, Ba and Sr throughout the mass of the granitic gneiss. The anomalous Rb, Ba, and Sr contents of the whole rock G13 (Table IV, Appendix) are due entirely to variation in the proportions of potash felspar and plagioclase and not to any change in the Rb, Ba and Sr contents of the constituent minerals. That this is probably so for G221, G274 and G283 also (Tables VI, VII and VIII respectively) can be shown by calculating theoretical values of the Rb, Ba and Sr contents for these rocks from the average Rb, Sr, and Ba in the biotites, potash felspars, and plagioclases and the proportions of these minerals in the modified norms (Table XV, Appendix) of these rocks, and comparing them with the actual values found. This is done in Table 33, which shows that apart from Ba in G283 and Sr in G221 the agreement is very good. The significance of these two anomalies is not known.

The variation in the Rb, Ba and Sr contents of the minerals separated/

TABLE 33.

Comparison of the trace element contents of certain granitic gneisses with values predicted from the normative amounts of biotite and felspars they contain and the average trace element contents of these minerals in the granitic gneiss.

Rock	Normative Biotite	Normative K felspar	Normative plagio-clase	Actual Rb	Actual Ba	Actual Sr	Pre-dicted Rb	Pre-dicted Ba	Pre-dicted Sr
G13	11.00	7.64	48.02	122	330	170	97	354	174
G221	8.98	4.77	54.00	87	340	430	79	259	185
G274	8.75	4.20	51.41	92	200	210	75	237	175
G283	11.08	0.90	50.57	77	360	180	78	149	165

Ele-ment.	Average amount in Biotite	Average amount in K felspar	Average amount in plagioclase
Rb	565	300	25
Ba	637	3089	100
Sr	16	251	318

separated from the psammites is greater than in the case of the granitic gneiss, but the individual results rarely differ by a factor of more than 2 for any one element in any one mineral species. The actual amount of Rb is about 20% less than for the corresponding mineral in the granitic gneiss, and the amounts of Ba and Sr are about 30% and 50% higher respectively. The bulk of the considerable variation in the Rb, Ba, and Sr contents of the psammites (see Tables IX and X, Appendix) is due to variations in the relative amounts of quartz, potash felspar, and plagioclase in the rocks. This is due to the fact that these rocks are metamorphosed arkoses and quartzites, and the Sr, Ba and probably even Rb contents of the minerals reflect those of the detrital felspars which went to form the original sediment. The small amounts of mica in these rocks reflect the small amounts of clayey material in the original sediments.

In the case of the pelites, however, the situation is different. The relative proportions of quartz, plagioclase, potash felspar, biotite, muscovite and garnet (where present) do not correspond to the proportions of these minerals in the original sediments. Hence the whole rock Rb, Ba, and Sr contents are not related to the proportions of the chief Rb, Ba and Sr minerals at present in the rocks. The potash felspar, plagioclase and biotite from P192, which contains abundant potash felspar, have similar Rb, Ba and Sr contents to these minerals/

minerals in the psammites, but the biotite, plagioclase, and, where present, muscovite from P125, P223, P291 and from the calcareous psammite S172, which have little or no potash felspar, show different trace element patterns. The biotites and muscovites contain more than twice as much Ba (Table 35) as any others analysed in the present study, and the muscovite, though not the biotites, is appreciably enriched in Sr (Table 36). The plagioclases appear to be slightly enriched in Ba, though the amount is still very low (Table 35), and contain rather more Sr (Table 36) than any others of the analysed plagioclases except those from G103 and S141. The absence of potash felspar from these rocks does not increase the Rb contents of the biotites, because of the large quantities of biotite in these rocks and the strong preference for biotites as against potash felspar shown by Rb.

The distributions of the individual elements between co-existing phases will now be discussed.

(i) Rubidium: The electronegativity and ionisation potential of Rb are similar to those of K, and Rb^+ ions differ significantly from K^+ ions only in size (Table 32). For this reason, the larger Rb^+ ion would be expected to prefer the 12-coordinated site occupied by K^+ in micas to the 8 - 10 co-ordinated site in potash felspars, while plagioclase, which is low in K^+, would be expected to be low in Rb^+ also. The relative ease of entry of Rb into muscovite and biotite could not/

TABLE 34.

Contents of Rb in, and distributions of Rb between analysed coexisting minerals.

Rock	ppm Rb.					Rb biot / Rb musc.	Rb musc / Rb Kfels.	Rb biot / Rb Kfels.	Rb Kfels / Rb plag.	K/Rb musc / K/Rb biot.	K/Rb Kfels / K/Rb musc.	K/Rb Kfels / K/Rb biot.	K/Rb plag. / K/Rb Kfels.
	Biotite	Muscovite	Kfelspar	Plagioclase	Hornblende								
G 12	630	-	344	21	-	-	-	1.83	16.4	-	-	2.94	0.75
G 13	605	-	276	11	-	-	-	2.19	25.1	-	-	3.04	0.85
G 62	510	-	285	33	-	-	-	1.79	8.6	-	-	3.28	0.39
G 92	595	-	280	20	-	-	-	2.12	14.0	-	-	3.62	0.54
G103	n.d.	-	237	14	-	-	-	n.d.	16.9	-	-	n.d.	0.75
G163	600	-	360	20	-	-	-	1.67	18.0	-	-	3.00	0.67
G183	565	-	310	19	-	-	-	1.82	16.3	-	-	3.00	0.55
G241	560	349	261	8	-	1.61	1.34	2.14	32.6	1.84	1.82	3.34	0.87
G262	410	-	272	67	-	-	-	1.51	4.1	-	-	2.98	0.45
G281	610	-	313	23	-	-	-	1.95	13.6	-	-	3.25	0.63
P125	365	-	-	9	-	-	-	-	-	-	-	-	-
P192	460	-	213	7	-	-	-	2.16	30.4	-	-	4.67	0.54
P223	450	-	-	22	-	-	-	-	-	-	-	-	-
P291	335	210	-	17	-	1.60	-	-	-	1.82	-	-	-
S 42	415	-	210	6	-	-	-	1.98	35.0	-	-	3.44	1.28
S 77	645	-	327	56	-	-	-	1.97	5.8	-	-	5.10	0.30
S141	455	-	235	28	-	-	-	1.93	8.4	-	-	2.34	0.81
SK211	515	314	150	5	-	1.64	2.09	3.43	30.0	1.91	1.77	3.38	1.27
S172	340	-	-	21	11	-	-	-	-	-	-	-	-

not be predicted.

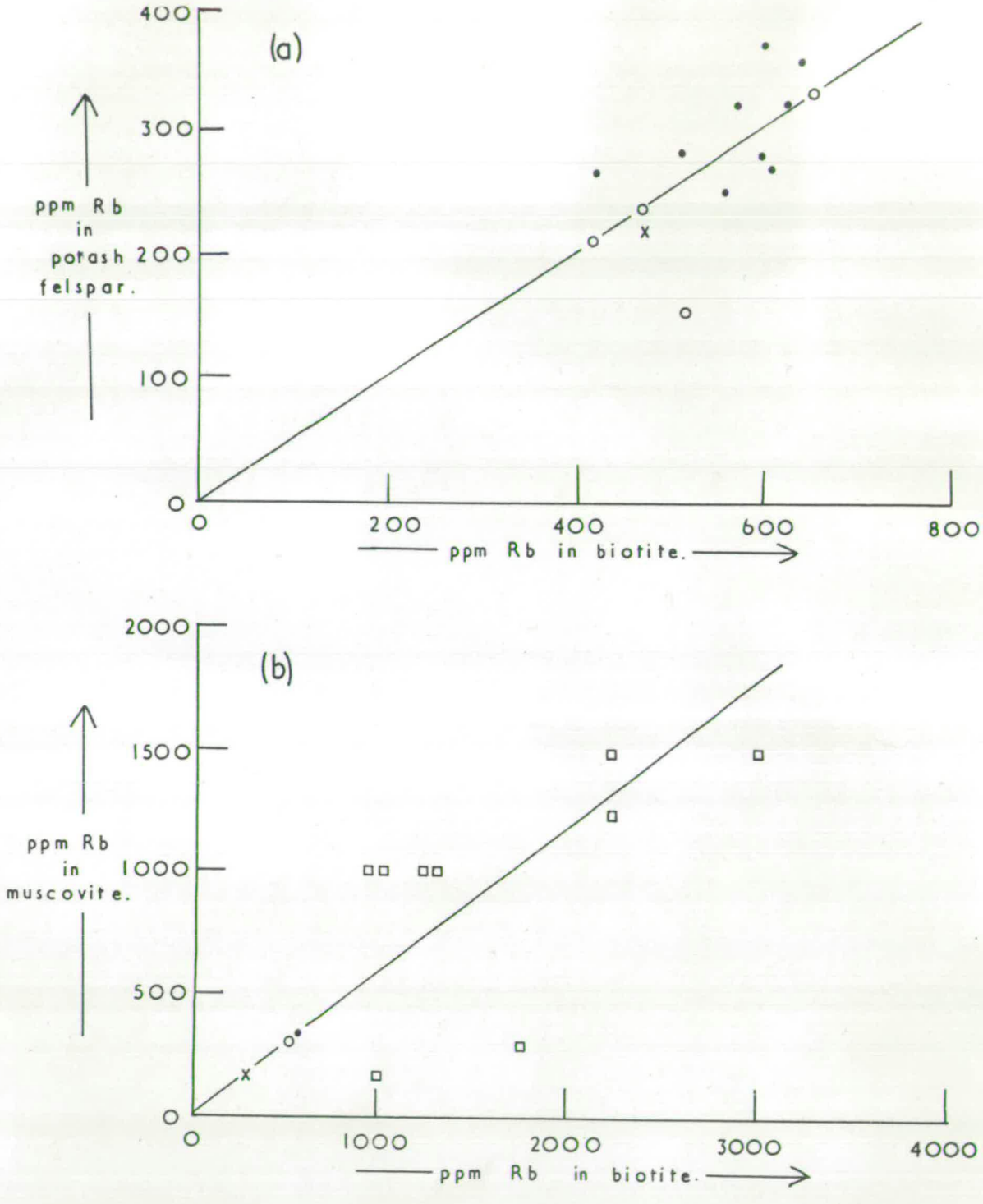
The results, as displayed in Table 34, show that, as predicted, rubidium shows a strong preference for biotite over potash felspar. In the major K-bearing minerals Rb shows a preference in the order biotite > muscovite > potash felspar, whether absolute values or K/Rb ratios are considered. Despite the much smaller absolute amounts of Rb in plagioclase than in coexisting potash felspar, there is, compared to K, relatively more Rb in the plagioclase. This result is difficult to explain unless it is due to the large amount of experimental error involved in determining the small quantities of K and Rb in the plagioclases. To see whether there was any relation between the distribution coefficients for Rb and the actual amount of Rb in the phases, the Rb content of the biotites was plotted first against that of the potash felspars (Fig. 25a), and secondly against that of the muscovites (Fig. 25b). Some biotite-muscovite pairs from Lambert (1959) are also included in Fig. 25b. Figure 25a shows rather a large spread, particularly among the granitic gneisses, but this does not necessarily mean that they are not all equilibrium assemblages. There seems to be a tendency for the potash felspars from rocks containing appreciable quantities of muscovite to be slightly depleted in Rb, but it is doubtful whether the present inadequate data are really significant in this respect. The three Ardgour muscovite-biotite pairs (Fig. 25b) show good linear relationship/

FIGURE 25.

Distribution of rubidium between coexisting potash felspar and biotite and between coexisting muscovite and biotite.

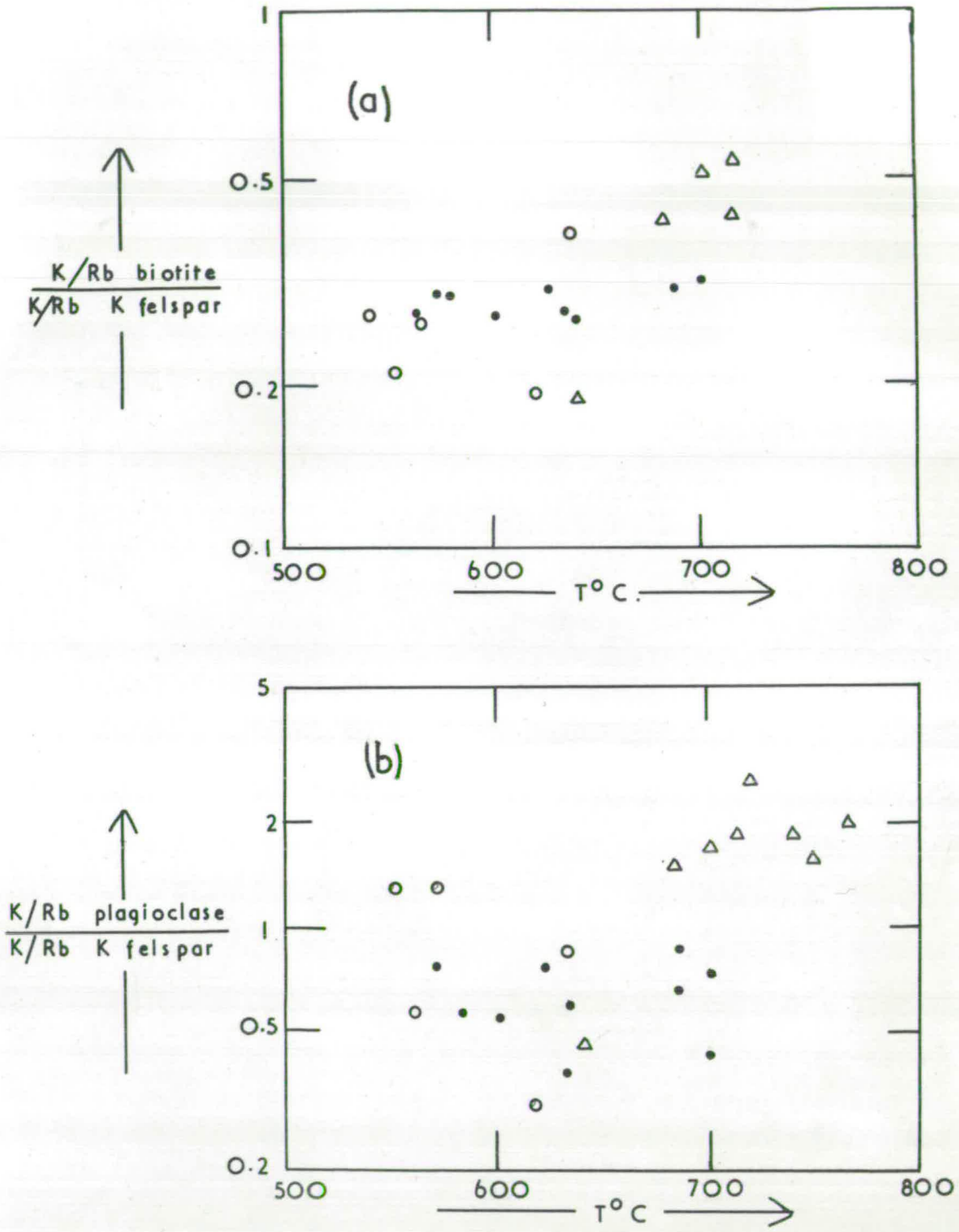
FIGURE 26.

Variation of $\frac{\text{K/Rb biotite}}{\text{K/Rb potash felspar}}$ and $\frac{\text{K/Rb plagioclase}}{\text{K/Rb potash felspar}}$ with temperatures of crystallisation derived from Figure 24, for rocks from Ardgour and Langøy.



• Ardgour granitic gneiss.
x Ardgour pelite.
o Ardgour psammite.
□ Knoydart schist (from Lambert 1959).

FIG. 25.



• Ardgour granitic gneiss.
o Ardgour metasediment.
△ Granulite and gneiss from Langøy (Heier 1960).

FIG. 26.

relationship which is not shared by the mica pairs from Knoydart (Lambert 1959); however, the lack of correlation between the Rb contents found for the Knoydart biotites and those for the coexisting muscovites may be due to the large degree of experimental error inherent in the particular optical spectrograph used by Lambert (1959).

The ratios K/Rb biotite/K/Rb potash felspar and K/Rb plagioclase/K/Rb potash felspar were plotted against the temperatures of crystallisation found for the Ardgour rocks from the felspar compositions (Table 29). For comparison, the data of Heier (1960) were plotted in the same fashion (Fig. 26). Though there appears to be a slight correlation with temperature, this may be spurious, particularly in the case of Fig. 26 (b), where the fact that Heier's (1960) plagioclases are more calcic than those from Ardgour (compare Tables 29 and 30) may explain why they are relatively poorer in Rb.

(ii) Barium: The Ba^{2+} ion is the same size (1.33Å) as the K^{+} ion, and both elements have low ionisation potentials and electronegativities (Table 32), so that any difference between the distributions of Ba^{2+} and K^{+} can be attributed to the difference in charge. This results in decreased ease of entry into the 12-coordinated sites in micas and an increased preference for the 8-10 coordinated site in the potash felspar lattice. The results are displayed in Table 35, and show that/

TABLE 35.

Contents of Ba in, and distributions of Ba between analysed coexisting minerals.

Rock	ppm Ba					Ba biot / Ba musc.	Ba musc / Ba Kfels.	Ba biot / Ba Kfels.	Ba Kfels / Ba plag.	K/Ba musc / K/Ba biot.	K/Ba Kfels / K/Ba musc.	K/Ba Kfels / K/Ba biot.	K/Ba Plag. / K/Ba Kfels.
	Biotite	Muscovite	Kfelspar	plagioclase	Hornblende								
G 12	705	-	2800	126	-	-	-	0.252	22.2	-	-	0.40	1.03
G 13	770	-	3000	62	-	-	-	0.233	48.4	-	-	0.35	1.64
G 62	625	-	3100	138	-	-	-	0.202	22.5	-	-	0.37	1.01
G 92	575	-	3150	113	-	-	-	0.182	27.9	-	-	0.31	1.06
G103	n.d.	-	6750	214	-	-	-	n.d.	31.5	-	-	n.d.	1,35
G163	635	-	2600	77	-	-	-	0.244	33.8	-	-	0.44	1.26
G183	580	-	3100	47	-	-	-	0.187	66.0	-	-	0.31	2.19
G241	710	1065	4200	67	-	0.67	0.254	0.169	62.7	0.77	0.34	0.26	1.64
G262	495	-	2800	159	-	-	-	0.177	17.6	-	-	0.35	1,97
G281	635	-	3050	115	-	-	-	0.208	26.5	-	-	0.35	1.22
P125	2300	-	-	240	-	-	-	-	-	-	-	-	-
P192	760	-	4350	99	-	-	-	0.175	43.9	-	-	0.25	1.24
P223	1740	-	-	158	-	-	-	-	-	-	-	-	-
P291	1670	3450	-	68	-	0.48	-	-	-	0.55	-	-	-
S 42	760	-	4250	141	-	-	-	0.179	30.1	-	-	0.31	1.09
S 77	350	-	2700	360	-	-	-	0.130	75.0	-	-	0.14	0.96
S141	810	-	4700	169	-	-	-	0.172	27.8	-	-	0.50	1.12
SK211	745	1295	4700	27	-	0.57	0.276	0.159	17.4	0.67	0.36	0.25	0.51
S172	2700	-	-	149	35	-	-	-	-	-	-	-	-

that Ba^{2+} exercises, as expected, a strong preference for the potash felspar phase over biotite while muscovite is, as in the case of Rb^{+}, intermediate between the two. As with Rb^{+}, the potash felspar is much richer in Ba^{2+} than the coexisting plagioclase, but this time the K/Ba ratio is higher in the plagioclase than in the coexisting potash felspar. This, together with the Rb^{+} result suggests that, though experimental error may be responsible for this unexpected behaviour, it may be due to distortion of the plagioclase lattice by K^{+} ions resulting in the appearance of a small proportion of large sites which will accept Rb^{+} slightly more, and Ba^{2+} slightly less, readily than will the K^{+} sites in potash felspar. The barium contents of the biotites and potash felspars were plotted against each other (Fig. 27a), but the results show rather a large spread. The barium contents of the biotites and muscovites were compared in Fig. 27b which shows that although the Ardgour mica pairs show similar distribution coefficients for Ba, those analysed by Lambert (1959) are much more variable.

The ratios K/Ba biotite/K/Ba potash felspar and K/Ba plagioclase/K/Ba potash felspar were plotted against the temperatures found for the crystallisation of the Ardgour (Table 29) and Heier's Langøy (Table 30) rocks (Fig. 28). The results show even less correlation between the variation in the ratios studied and the temperature than in the case of Rb./

FIGURE 27.

Distribution of barium between coexisting potash felspar and biotite, and between coexisting muscovite and biotite.

FIGURE 28.

Variation of $\frac{\text{K/Ba biotite}}{\text{K/Ba potash felspar}}$ and $\frac{\text{K/Ba plagioclase}}{\text{K/Ba potash felspar}}$

with the temperatures of crystallisation derived from Figure 24, for rocks from Ardgour and Langøy.

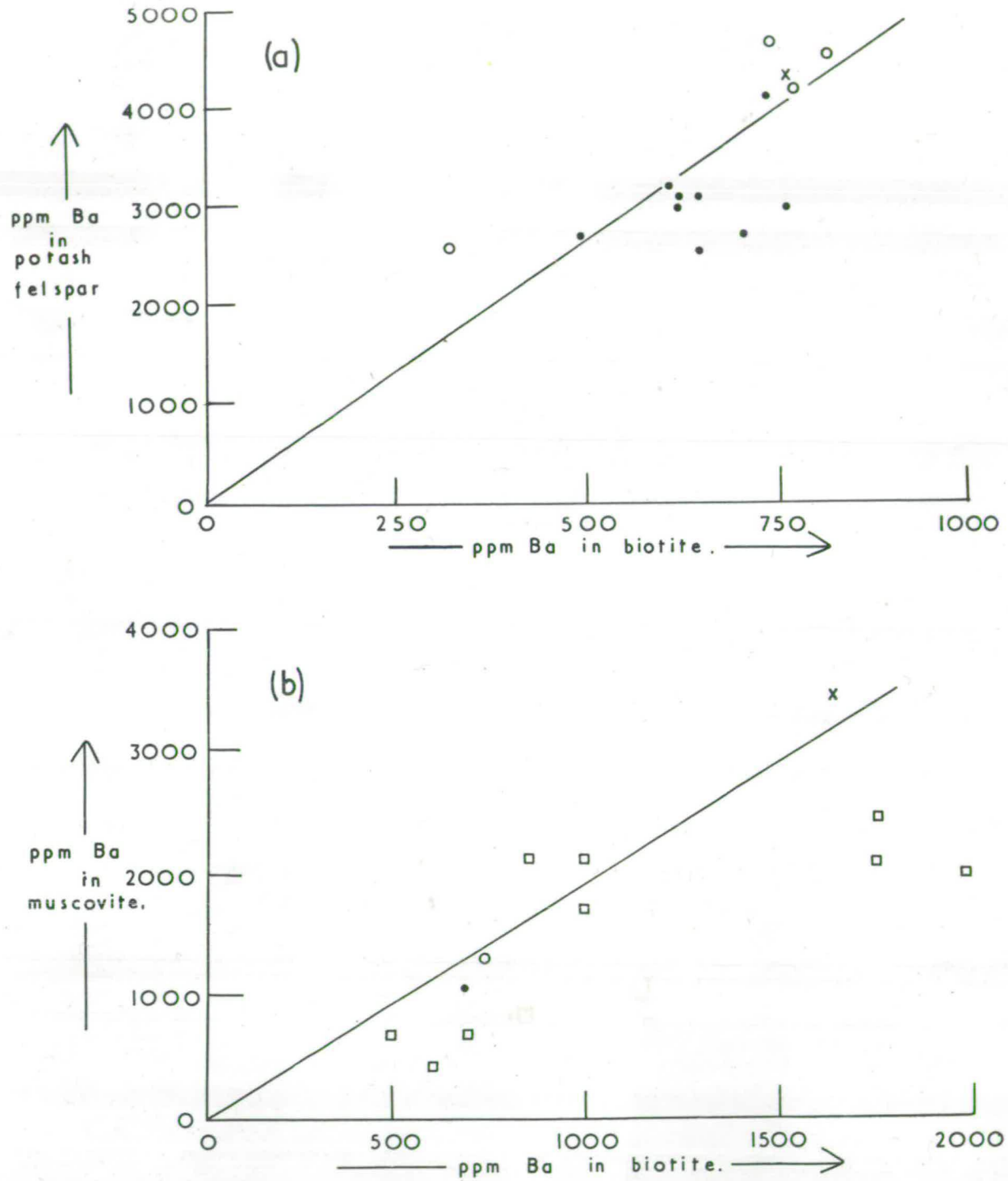

(a)
5000
4000
3000
2000
1000
0
ppm Ba in potash felspar
0
250
500
750
1000
ppm Ba in biotite.
(b)
4000
3000
2000
1000
0
ppm Ba in muscovite.
500
1000
1500
2000
ppm Ba in biotite.
• Ardgour granitic gneiss.
x Ardgour pelite.
o Ardgour psammite.
□ Knoydart schist (from Lambert 1959).

FIG. 27.

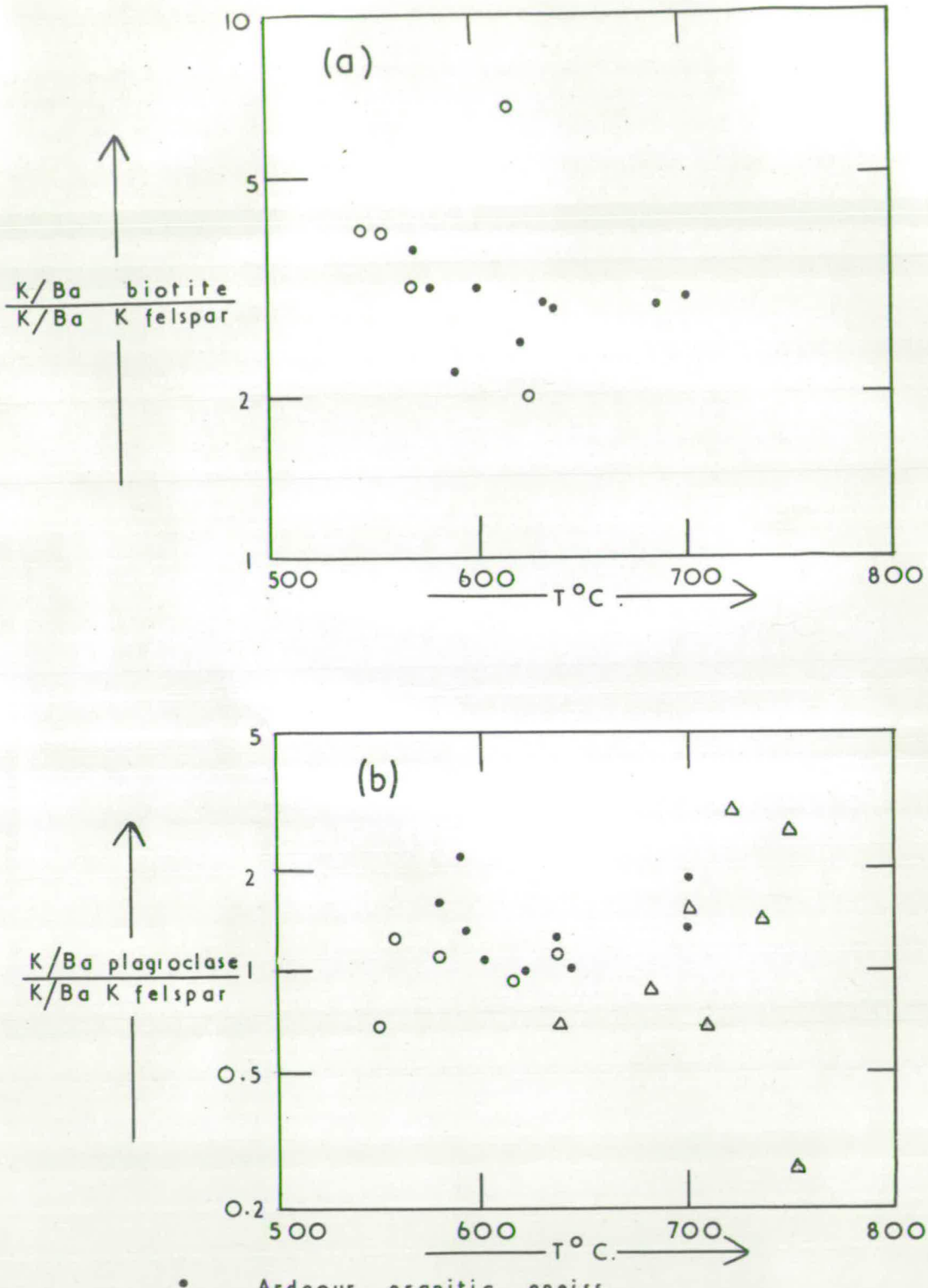

(a)
10
5
2
1
K/Ba biotite
K/Ba K felspar
500
600
700
800
T °C
(b)
5
2
1
0.5
0.2
K/Ba plagioclase
K/Ba K felspar
500
600
700
800
T °C.
• Ardgour granitic gneiss.
o Ardgour metasediment.
△ Granulite and gneiss from Langøy (Heier 1960).

FIG. 28.

Rb.

(iii) Strontium: The Sr^{2+} ion (radius 1.12 Å) is considerably smaller than K^+ (1.33Å) and rather larger than Ca^{2+} (0.99Å) or Na^+ (0.97Å). Strontium has long been considered to follow calcium in igneous and metamorphic rocks, but some major differences between the behaviour of the two elements are well known. The principal ones are the small amounts of Sr which can be accepted into the lattices of clinopyroxenes and amphiboles of which Ca is a major component, and the relatively large quantities of Sr which can enter potash felspars, from which Ca is fairly rigorously excluded. The geochemical coherence of Ca and Sr was found to be low by Turekian and Kulp (1956), who showed that only in deep sea sediments and granitic rocks was there any significant positive correlation between the amounts of Ca and Sr in different rocks of the same type. For these reasons, it was considered worthwhile to explore the possibility that Sr might tend to follow Na rather than Ca in igneous and metamorphic rocks. That this suggestion is plausible is shown by the fact that the chief Sr-bearing minerals, plagioclase and potash felspar, both contain considerable amounts of Na; in a magmatic differentiation sequence such as that shown by the Skaergaard intrusion (Wager and Mitchell 1952), the ratio Sr in plagioclase/Sr in liquid is believed to have been highest when the plagioclase was a fairly sodic andesine. The only common Sr-bearing/

bearing mineral which is poor in Na is apatite, from which Na^+ is usually barred by its charge, since entry of Na^+ must be compensated for either by replacement of PO_4^{3-} by SO_4^{2-}, or by replacement of another Ca^{2+} by $(\text{rare earth})^{3+}$. Compared with Na^+, Sr^{2+} has a higher charge, which would encourage its preferential admittance into Na-bearing minerals, but a larger ionic radius (1.2Å compared with 0.97Å), which would be expected to have the opposite effect. These conflicting tendencies are demonstrated by the behaviour of Rb^+ (1.47Å) And Ba^{2+} (1.33Å) compared with K^+ (1.33Å). The opposite effects of doubling the charge and increasing the ionic radius by a factor of 1.10 seem to be approximately equal, since the average ratio (granitic gneisses only)

$$\frac{\text{K/Rb potash felspar}}{\text{K/Rb biotite}} \times \frac{\text{K/Ba potash felspar}}{\text{K/Ba biotite}} = 3.16 \times 0.35 = 1.11.$$

The ratio of the radius of the Sr^{2+} ion to that of the Na^+ ion is 1.15, and this, together with the fact that the electronegativities of Sr^{2+} and Na^+ are almost identical (Table 32), suggests that the distributions of Sr^{2+} and Na^+ between biotite and potash felspar are probably very similar; the actual average values found for the Ardgour granitic gneiss are, Sr biotite/Sr potash felspar = 0.065; Na biotite/Na potash felspar = 0.106. This evidence does not necessarily mean that Sr^{2+} and Na^+ behave similarly in their distribution between all the coexisting minerals of a rock, but encourages the supposition

that Sr and Na might show similar geochemical behaviour in silicate rocks.

The evidence regarding igneous differentiation series will be listed below. In the Skaergaard intrusion (Wager and Mitchell 1952) the Sr content of the liquid reached a maximum when 60% of the rock had crystallised, while the Ca content of the liquid underwent a steady slight decrease and the Na content showed a barely perceptible rise; in the solid phases the maximum Sr content occurs at the level corresponding to 70% crystallised, while the Ca content shows no maximum and Na, Ba, and K show maxima at 90, 99 and 100% crystallised. In calcalkali igneous rocks (Nockolds and Allen, 1953) Sr usually reaches a maximum at / ($\frac{1}{3}$ Si + K) - (Ca + Mg) / = 10 while Ca shows a steady decrease and Na a steady rise throughout. In alkaline igneous series (Nockolds and Allen 1954) Sr behaves in a fashion intermediate between Ca and Na, often showing a pronounced maximum at / ($\frac{1}{3}$ Si + K) - (Ca + Mg) / = 5 - 10, while Na has a maximum when this index is near 10 and Ba between 10 and 15, though Ca shows a steady decrease throughout; alternatively the curves may show no maxima, and in this case Sr usually parallels Na in showing a slight rise during differentiation. However, Sr, along with Ba and the rare earths, is concentrated in carbonatites. In tholeiitic differentiation series (Nockolds and Allen 1956) Sr often remains nearly constant and its/

TABLE 36.

Contents of Sr in, and distributions of Sr between analysed coexisting minerals.

Rock	ppm Sr					Sr biot. / Sr musc.	Sr musc. / Sr Kfels.	Sr biot. / Sr Kfels.	Sr Kfels. / Sr plag.	Ca/Sr plag. / Ca/Sr Kfels.	Na/Sr plag. / Na/Sr Kfels.
	Biotite	Muscovite	Kfelspar	Plagioclase	Hornblende						
G 12	10	-	190	257	-	-	-	0.053	0.74	28.2	4.00
G 13	16	-	260	290	-	-	-	0.062	0.90	9.3	3.25
G 62	22	-	328	408	-	-	-	0.067	0.81	20.8	4.00
G 92	16	-	218	276	-	-	-	0.073	0.79	29.2	4.63
G103	n.d.	-	715	839	-	-	-	n.d.	0.85	8.8	2.69
G163	8	-	252	331	-	-	-	0.032	0.76	32.7	4.83
G183	10	-	216	285	-	-	-	0.046	0.76	37.0	4.95
G241	14	28	296	356	-	0.50	0.094	0.047	0.83	49.5	5.68
G262	26	-	266	408	-	-	-	0.098	0.65	9.9	2.09
G281	24	-	234	267	-	-	-	0.103	0.87	22.7	5.13
P125	8	-	-	794	-	-	-	-	-	-	-
P192	14	-	456	594	-	-	-	0.031	0.77	19.6	4.95
P223	14	-	-	738	-	-	-	-	-	-	-
P291	26	128	-	1032	-	0.20	-	-	-	-	-
S 42	8	-	350	432	-	-	-	0.023	0.81	29.2	5.35
S 77	18	-	316	432	-	-	-	0.057	0.73	32.9	3.62
S141	20	-	598	752	-	-	-	0.033	0.79	17.1	4.17
SK211	14	34	506	635	-	0.41	0.067	0.028	0.80	22.1	5.29
S 172	22	-	-	887	48	-	-	-	-	-	-

its trend parallels that of Na.

In sedimentary carbonate rocks, the Sr/Ca ratio is lower than in most other rocks, and the difficulty of Sr in finding its way into calcite (though it probably finds entry into aragonite easier) marks another point of discordance between the geochemistry of Sr and Ca.

In metamorphic rocks, the bulk composition of the rocks is usually not liable to change except at the very highest grades. Therefore, the geochemical nature of an element can only be expressed through its distribution between the various coexisting phases in the metamorphic rock. In the Ardgour rocks, nearly all of the Sr is taken up in the felspars; even hornblende, when present, contains very little Sr. The Sr content of the biotites and its relation to that of the potash felspars is mentioned on p. 135; comparison with the distribution of Ca between these phases is impossible because of the erratic Ca contents obtained (even when corrected for the presence of apatite in some of the separations) for the biotites.

As shown in Fig. 29, and in contrast to 6 of the 8 felspar pairs analysed by Heier (1960), the plagioclases from the Ardgour rocks contain slightly more Sr than the coexisting potash felspars, and the distribution coefficient shows little variation. The analysed Ardgour felspar pairs were considered in conjunction with those given by Heier (1960, Tables III and IV in Appendix) in an attempt to relate the strontium/

FIGURE 29.

Distribution of strontium between coexisting plagioclase and potash felspar.

FIGURE 30.

Variation of $\frac{\text{Ca/Sr plagioclase}}{\text{Ca/Sr potash felspar}}$ and $\frac{\text{Na/Sr plagioclase}}{\text{Na/Sr potash felspar}}$

with temperatures of crystallisation derived from Figure 24, for rocks from Ardgour and Langøy.

FIGURE 31.

Variation of Sr plagioclase/Sr potash felspar with temperatures of crystallisation derived from Figure 24 and with the anorthite content of the plagioclase, from rocks from Ardgour and Langøy.

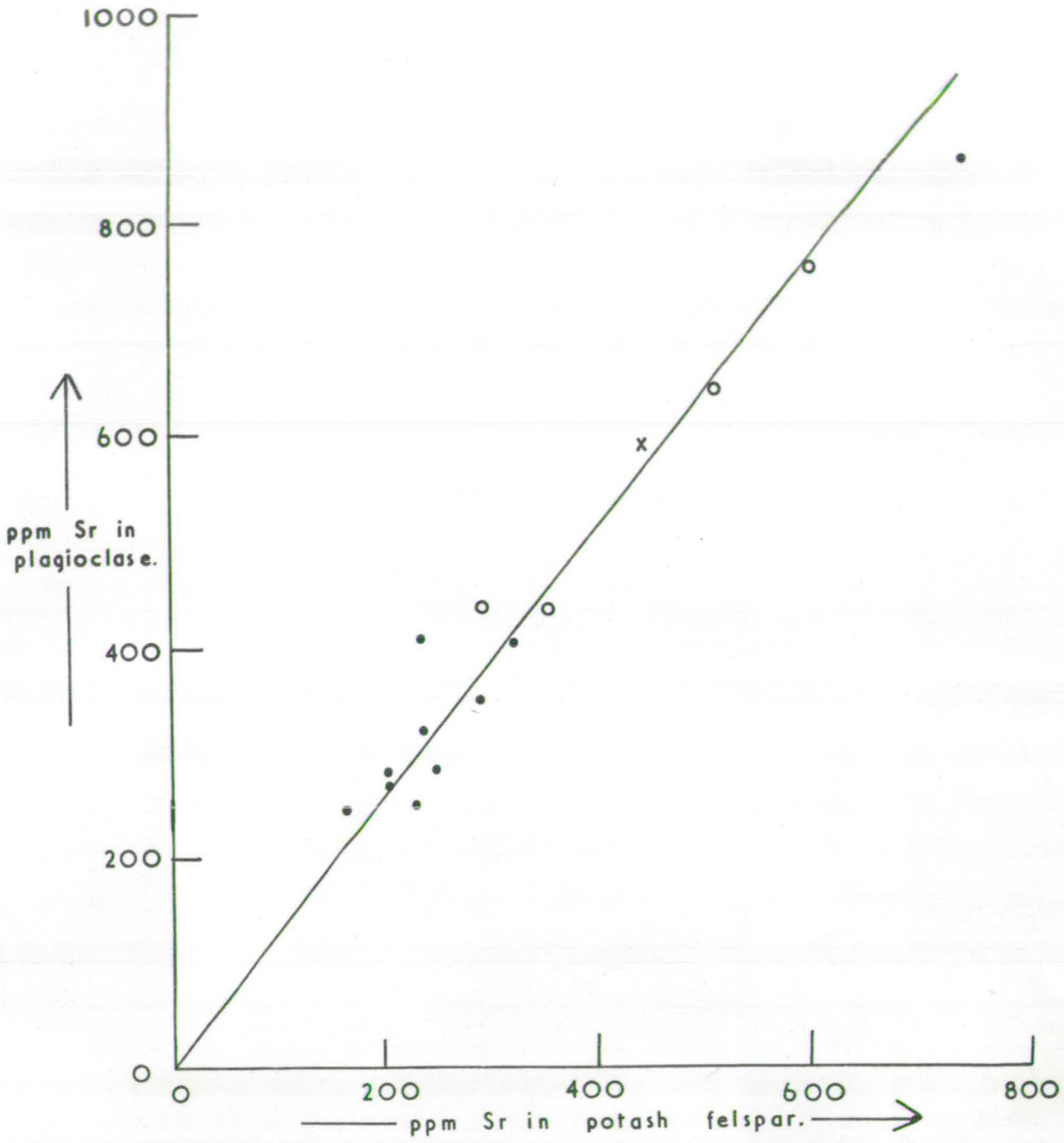
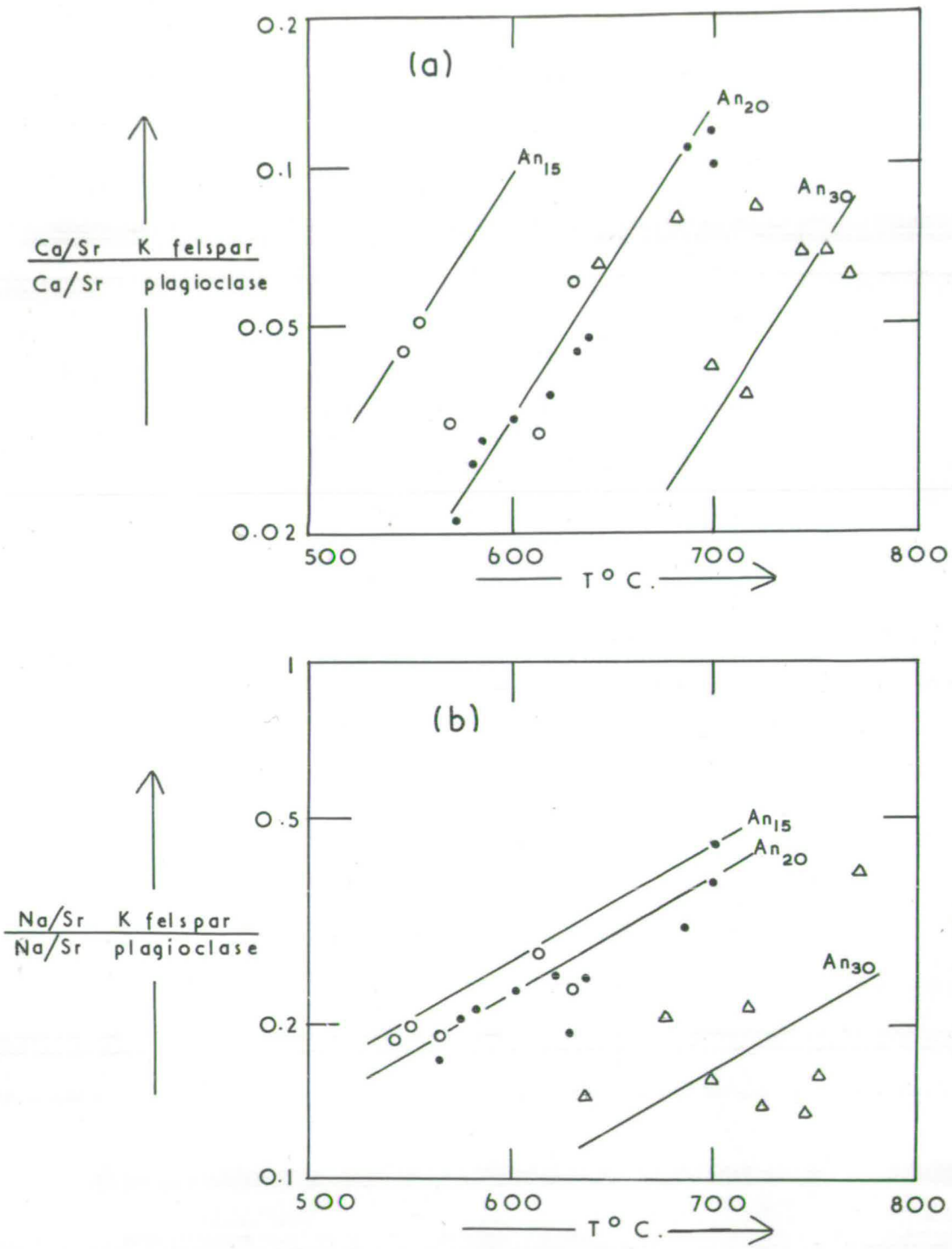

1000
800
600
400
200
0
ppm Sr in plagioclase.
0
200
400
600
800
ppm Sr in potash felspar.
• Ardgour granitic gneiss.
x Ardgour pelite.
o Ardgour psammite.

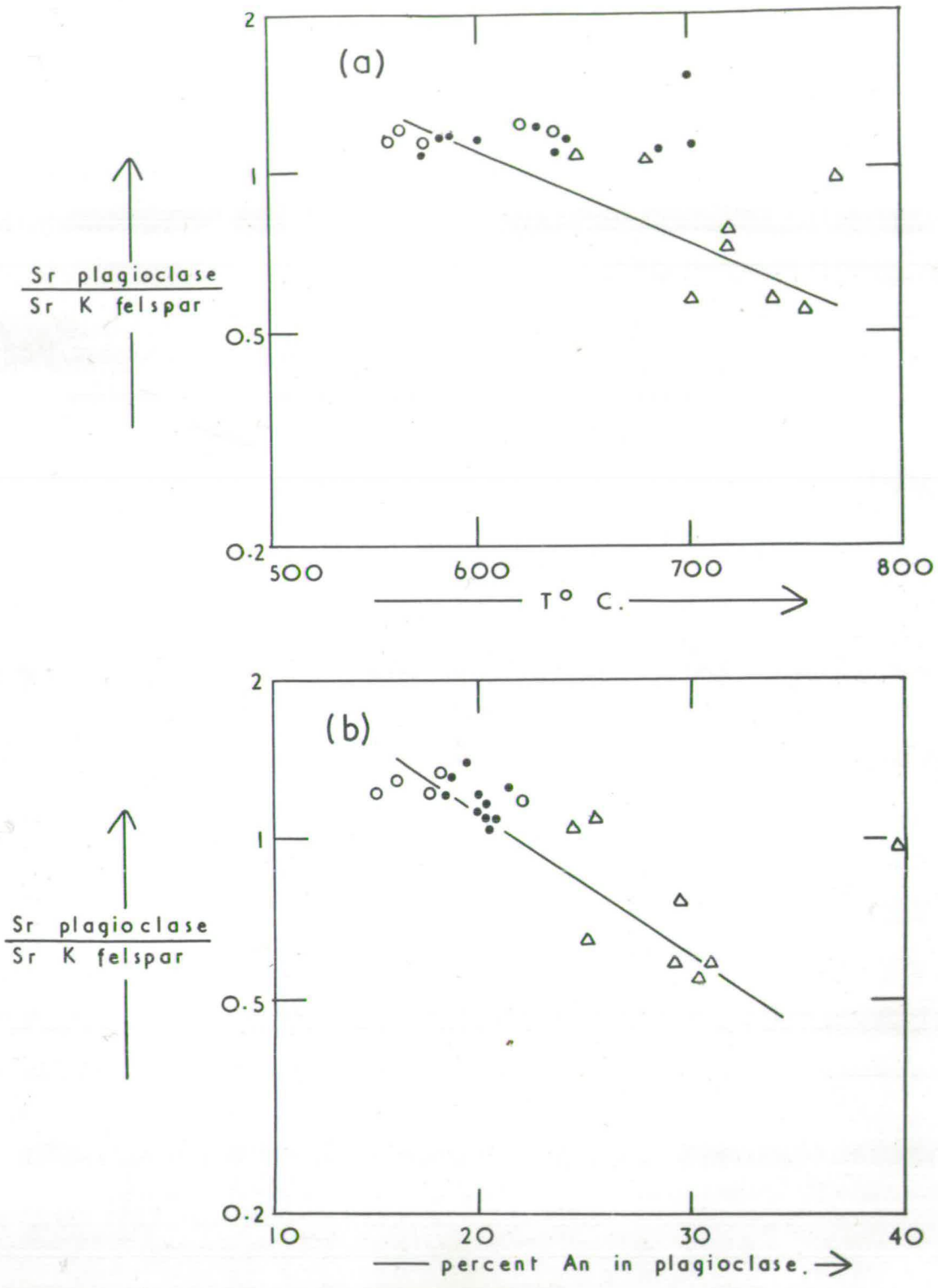
FIG. 29.



• Ardgour granitic gneiss.
o Ardgour metasediment.
△ Granulite and gneiss from Langøy (Heier 1960).

Lines illustrate the effect of plagioclase composition on the ratios plotted.

FIG. 30.



- Ardgour granitic gneiss.
- Ardgour metasediment.
- Granulite and gneiss from Langoy (Heier 1960).

FIG. 31.

strontium distribution to the temperature of crystallisation given by the distribution of Ca, Na, and K between the co-existing felspars (Tables 29, 30). The ratio Sr in plagioclase/Sr in potash felspar was plotted against the temperature of crystallisation (Fig. 31 (a)), and though some measure of correlation was observed, a better curve was obtained by plotting (Fig. 31 (d)) Sr in plagioclase/Sr in potash felspar against % An in plagioclase (Fig. 31 (b)). In order to allow for this factor, the ratios Na/Sr in potash felspar/Na/Sr in plagioclase and Ca/Sr in potash felspar/Ca/Sr in plagioclase were plotted against the assumed temperature of crystallisation (Fig. 30 (b), (a)). Both graphs gave straight lines for pairs with equal An contents in the plagioclase, and departure from the line was proportional to the difference in the An content of the plagioclase from that of the rocks lying on the line. These results indicate that the distribution of Sr between coexisting felspars may be able to yield an estimate of the temperature of crystallisation of a coexisting felspar pair given the An content of the plagioclase, but more data, including preferably some from experimental work, are ncessary before this can be done.

The results listed in Table 36 show that the distribution coefficient of Sr between plagioclase and potash felspar is closer to that of Na than to that of Ca between the felspar phases. Nevertheless, the accumulated evidence, though it suggests that Sr/

Sr in igneous and metamorphic rocks shows greater geochemical coherence with Na than with Ca, is insufficient to prove this conclusively. The key to the problem may well lie in the location of the Sr^{2+} ions in natural "peristerites", that is, plagioclases between An_5 and An_{23} which have unmixed into a submicroscopic intergrowth of An_{0-5} and An_{23-25}.

3. THE PETROGENETIC SIGNIFICANCE OF THE Rb, Ba and Sr CONTENTS OF THE POTASH FELSPARS.

Recent work on the trace element contents of potash felspars from various gneisses, granites, and pegmatites from southern Norway, summarised by Taylor and Heier (1960), has demonstrated that the K/Rb, Ba/Rb and Ba/Sr ratios of potash felspars may be related to the origin of the rocks from which they come. It was shown that only extreme fractionation could cause the ratio K/Rb to decrease to values significantly lower than that characteristic of most terrestrial rocks. Lesser degrees of fractionation could be detected by decreases in Ba/Rb and increases in Ba/Sr in the potash felspars.

The potash felspars from the diapiric Fevig granite showed an inward decrease from the margins in K/Rb from 400 to 250 and a decrease in Ba/Rb from 18 to 5; this was considered to be the result of relatively mild crystal fractionation. More extreme fractional crystallisation took place in the large pegmatites/

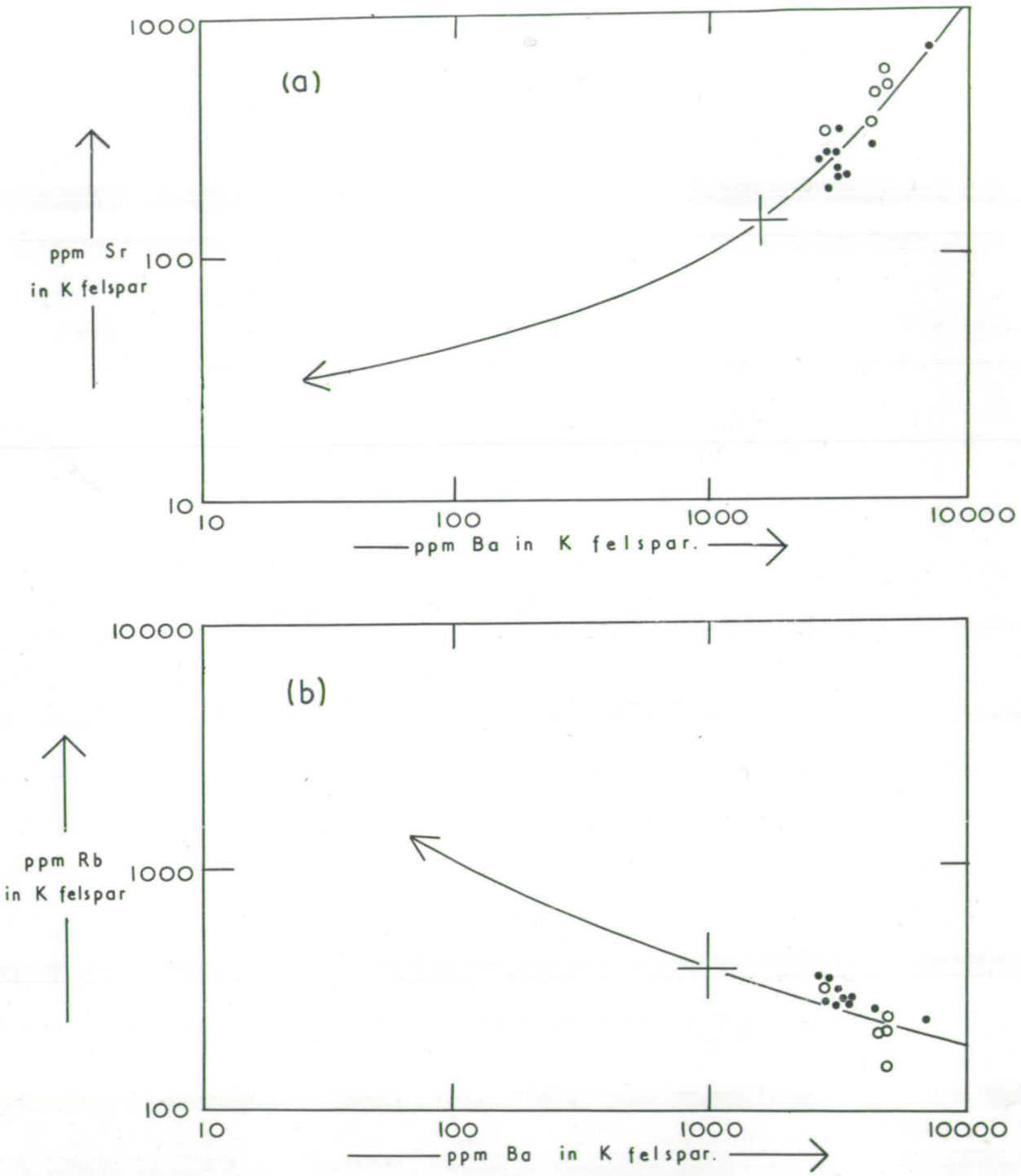
pegmatites (those over 1 m. in length and capable of being mined), where K/Rb ranged from 200 to 25, Ba/Rb from 1.1 to 0.5 and Ba/Sr from 3.5 to 8.6 (Taylor and Heier 1960).

By contrast, potash felspars from gneisses, augen-gneisses, anatectic granites, and small pegmatites were found to have "normal" K/Rb, Ba/Rb and Ba/Sr ratios (Taylor and Heier, 1960, p. 59), indicating that little or no fractionation had taken place. The opinion was expressed that a "granitic" liquid, whether "derived from residual melting or formed by anatexis, would be strongly differentiated", and transfer of material over distances not exceeding a few metres was considered adequate to account for the compositions of potash felspars of the augengneisses, anatectic granites and small pegmatites. However, Taylor (1965, pp. 191-198) considered that small granitic veins in migmatites, derived from the country rock by ultrametamorphic processes which perhaps involved localised melting, would not show enrichment in rare elements; granites produced by soaking of country rock in granitic melts would only show large-scale enrichment of the potash felspars in Rb, Cs and Tl if the granitic melt prior to soaking had undergone fractional crystallisation.

The variation in the Rb, Ba, and Sr content of potash felspars during fractional crystallisation of granitic liquids is shown in Taylor and Heier (1960, Fig. 4) and Heier and Taylor (1959 b, Fig. 3); the trend is reproduced in Fig. 32. The application/

FIGURE 32.

Variation of Rb, Ba and Sr in potash felspars from Ardgour granitic gneisses and metasediments compared with fractionation trends established by Heier and Taylor (1959). Curves drawn from Taylor (1965).



Differentiation beyond the point indicated by the cross is only achieved in large pegmatites (Taylor and Heier 1960).

FIG. 32.

application of these results to the potash felspars of rocks from other areas is subject to two restrictions:

(i) Variations in the proportions of K, Na, and Ca may change the relative ease of entry of Rb, Ba, and Sr into the potash felspar lattice. Granulite-facies potash felspars, which, having crystallised at higher temperatures than amphibolite-facies potash felspars, contain less K and more Na and Ca, often show a considerable Rb depletion (Heier 1960, Heier and Taylor 1959 a) relative to both K and Ba.

(ii) The availability of Rb, Sr and Ba may change the K/Rb, Ba/Rb and Ba/Sr ratios in the potash felspars. In isochemically metamorphosed sedimentary rocks, the amounts of Rb, Sr and Ba in the potash felspars are dependent upon the original Rb, Sr and Ba contents of the sediments and the amounts of micas and plagioclase which have crystallised (see pp.128-130). In a similar way, if a liquid has been involved, that liquid and the rocks formed by its crystallisation or by reaction with it may have anomalous trace element contents due to anomalous trace element composition of the material from which the liquid was formed. It is well known that K/Rb ratios are approximately constant in all magmatic provinces throughout the world, but provinces showing abnormally low or high contents of Ba and Sr may exist, and this would affect the absolute Ba/Rb and Ba/Sr values, though the same differentiation trends could be displayed in a fractionated sequence.

The/

The relationships between Ba and Rb and between Ba and Sr for the Ardgour potash felspars are shown in Fig. 32. For comparison, the trends of differentiation found in south Norwegian alkali felspars (Heier and Taylor 1959 a, and Taylor and Heier 1960) are also drawn. The Ardgour granitic gneiss has undergone only a very slight degree of fractionation compared with the surrounding metasediments (Fig. 32), and the actual Rb, Ba and Sr contents of the Ardgour potash felspars are very similar to those of the potash felspars from the South Norwegian gneisses, augen-gneisses and anatectic granites. Unfortunately, the potash felspars were not separated from any of the Ardgour pegmatites and aplites, but consideration of the few whole rock analyses of these rocks available (see Table XII, Appendix, Figs. 17 - 20, and pp. 158-159) suggests that they have not undergone any significant degree of fractionation compared with the Ardgour granitic gneiss. These relations would be similar to those found in southern Norway between the small pegmatites (less than 1 m. in length) and the surrounding gneisses and augen-gneisses (Taylor and Heier 1960).

The interpretation of these results depends on whether the granitic liquids produced by anatexis (that is, partial melting) of metasediments or basement rocks would be expected to be highly differentiated. The opinions of Taylor and Heier (1960) and Taylor (1965) differ on this point. In view of the results/

results obtained by Taylor and Heier (1960) for potash felspars from anatectic and diapiric granites, only a small amount of Rb enrichment and Ba and Sr depletion would be expected in a granitic partial melt of metasediments or basement rocks. Thus it seems likely that a granitic liquid was involved in the origin of the Ardgour granitic gneiss, and that this liquid did not undergo any significant amount of crystallisation differentiation.

VII. THE ORIGIN OF THE GRANITIC GNEISS

A considerable amount of geochemical detail regarding the Ardgour granitic gneiss has been presented in Chapters Four to Six. Any attempt to assess the probable origin of the granitic gneiss in the light of this evidence involves consideration of recent experimental and petrological studies. Discussion of the geochemical evidence concerning the origin of the granitic gneiss will therefore be preceded by a summary of recent experimental and petrochemical work.

The structural and petrographic evidence concerning the origin of the granitic gneiss, which was largely amassed by Harry (1953) and Dalziel (1963a, b), will then be discussed in the light of the geochemical evidence and an attempt will be made to find the mode of origin which is in best accord with all the available evidence. This mode of origin will then be briefly compared with those found by other workers for certain selected granitic rocks.

1. EXPERIMENTAL AND PETROCHEMICAL WORK ON THE ORIGIN OF GRANITIC ROCKS.

The modes of formation which have been postulated for granitic rocks can be divided into "magmatic" hypotheses, including all which involve a silicate liquid phase, hypotheses of granitisation by an aqueous vapour phase containing dissolved silicates/

silicates, and hypotheses of granitisation by ionic diffusion in the solid state. The first of these categories is much the largest, and includes ideas such as the soaking of metasediments in granitic liquids, the partial melting of either metasediments or "basement" material, the forceful intrusion of a granitic liquid or a (liquid & crystal) mush, and many others. The experimental and petrochemical work will be presented as it affects each of the three main categories in turn.

(a) The composition and behaviour of silicate liquids:

The "granite system" $NaAlSi_3O_8$ - $KAlSi_3O_8$ - SiO_2 - H_2O was thoroughly investigated by Tuttle and Bowen (1958). The presence of a minimum on the liquidus surface is important, and its composition controls both the end-product of crystallisation differentiation and the first liquids formed by partial melting of non-granitic material. The effect of the addition of varying amounts of $CaAl_2Si_2O_8$ to the "granite system" was studied by von Platen (1965) and Kleeman (1965), who showed that, at constant p_{H_2O}, the position of the liquidus minimum in the system $NaAlSi_3O_8$ - $KAlSi_3O_8$ - SiO_2 moved nearer the $KAlSi_3O_8$ - SiO_2 boundary as the ratio Ca/Na increased.

An interesting fact discovered by Bowen and Tuttle (1958, Figs. 26-27) is that as p_{H_2O} increases above about 2Kb, the temperature of the liquidus minimum decreases very slowly but the amount of water dissolved in the melt increases considerably, reaching/

reaching 9% at c. 4Kb.

Unfortunately, there is little experimental evidence with which to estimate the effect of the presence of ferromagnesian phases (biotite, hornblende, garnet) or minerals containing excess Al (garnet,muscovite) on the liquidus minimum compositions. The dry systems CaO - Al_2O_3 - SiO_2 (Greig 1927), Na_2O - Al_2O_3 - SiO_2 (Schairer and Bowen 1947) and K_2O - Al_2O_3 - SiO_2 (Schairer and Bowen 1947) have all been investigated at 1 atm. pressure. In no case does the mullite field cross the quartz-felspar join, and there are ternary eutectics corresponding to quartz + mullite + felspar which lie very close to the quartz-felspar join. In the system BaO - Al_2O_3 - SiO_2 (Thomas 1950), however, the point representing the assemblage quartz + celsian + mullite + liquid is a reaction point and the liquid lies on the BaO side of the quartz-felspar join. The systems MgO-Al_2O_3 - SiO_2 (Greig 1927) and FeO - Al_2O_3 - SiO_2 (Schairer and Yagi 1952) have likewise only been investigated under dry conditions. If the Ardgour biotites are split into normative orthoclase plus a ferromagnesian component and the latter is plotted in the system (Fe, Mg)O - Al_2O_3 - SiO_2, the line joining this point to SiO_2 does not cut the mullite field in the experimentally determined system. These results combine to suggest that, except possibly in the presence of very large quantities of Ba, partial melting of any material containing more Al than can be represented in the system $NaAlSi_3O_8$ - $CaAl_2Si_2O_8$ - $KAlSi_3O_8$ - (Ardgour biotite) - SiO_2 would give a liquid/

liquid containing excess Al, though the amount of excess Al would be small.

Tuttle and Bowen (1958, pp. 84-89) found that in the system $K_2O - Al_2O_3 - SiO_2 - H_2O$ at 1Kb, the liquidus temperature on the alkaline side of the quartz-potash felspar join fell steadily, while on the Al-rich side there was a eutectic separated from the falling curve by a thermal barrier at the quartz-potash felspar join. This, however, is the only study of such systems involving the H_2O component. The lack of more factual information in the system $SiO_2 - Al_2O_3 - FeO - CaO - Na_2O - K_2O - H_2O$, which contains the fewest components necessary for a really adequate study of granitic rocks, reduces estimates of the variation of liquidus temperatures in granitic compositions containing different amounts of Al to the level of informed guesswork. The suggested relations under conditions ($T = 600-700^{\circ}C$, $p = 3 - 11Kb$, $p_{H_2O} > 3Kb$, see p. 166)likely to have prevailed in Ardgour during the formation of the granitic gneiss are shown in Fig. 33, and are based partly on the dry systems mentioned above, and partly on the compositions of natural granitic rocks.

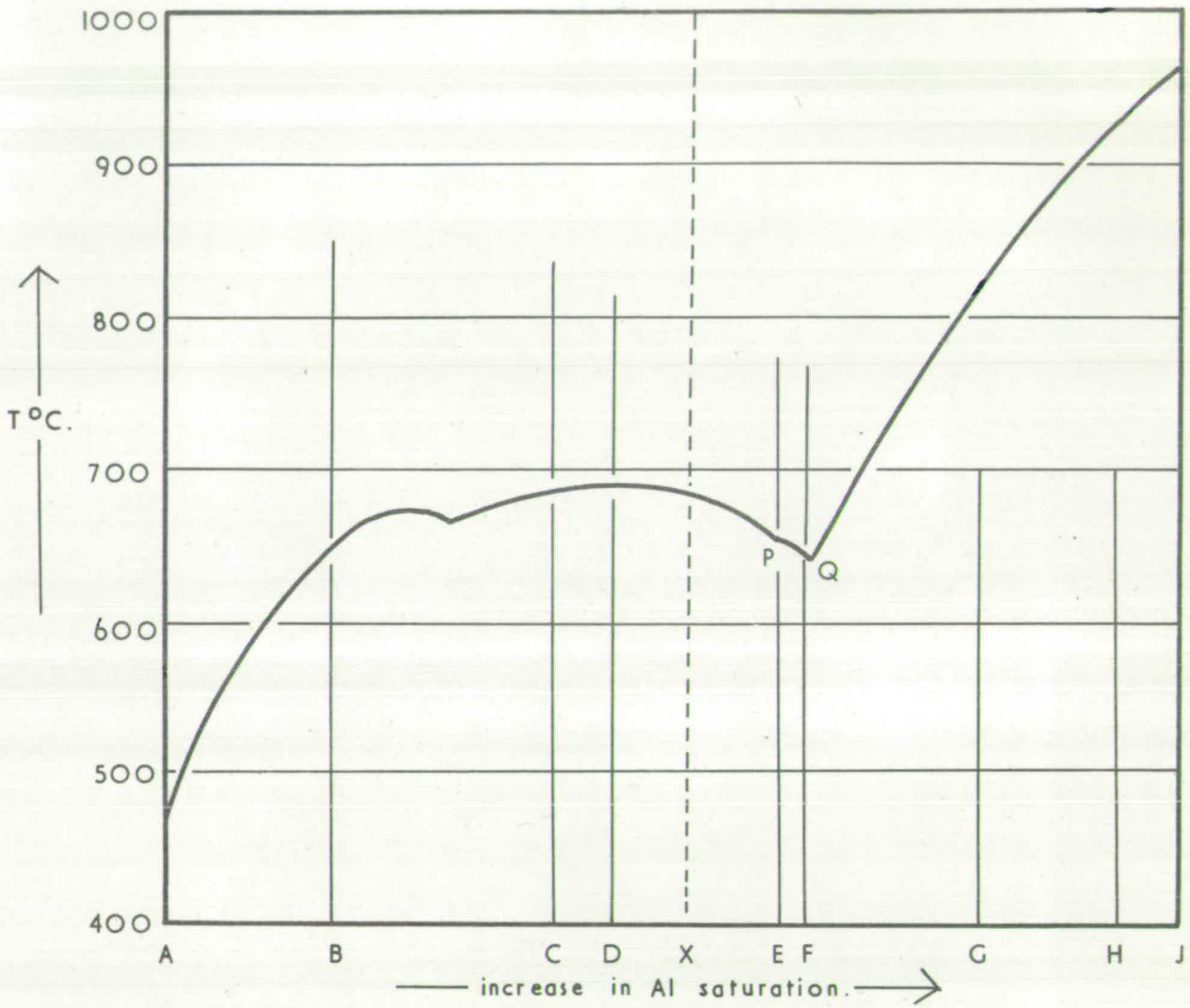
True granites and adamellites usually contain biotite alone, biotite + muscovite, or alkali amphiboles (riebeckite or arfvedsonite) $\pm$ acmite as their dark minerals. True granites and adamellites rich in hornblende are relatively scarce, as are cordierite - or almandine - bearing granites, which are confined almost/

FIGURE 33.

Hypothetical liquidus curve for granitic compositions with varying excesses or deficiencies of Al, for p_{H_2O} = about 5000 bars.

The minerals developed in the various composition ranges are –

A	Quartz + acmite + alkali silicates.
A – B	Quartz + plagioclase + potash felspar + riebeckite or acmite + alkali silicates.
B	Quartz + plagioclase + potash felspar + riebeckite or acmite.
B – C	Quartz + plagioclase + potash felspar + hornblende + riebeckite or acmite.
C – D	Quartz + plagioclase + potash felspar + hornblende.
D – E	Quartz + plagioclase + potash felspar + biotite + hornblende.
E – F	Quartz + plagioclase + potash felspar + biotite.
F – G	Quartz + plagioclase + potash felspar + muscovite + biotite (below 700°C).
G	Quartz + plagioclase + muscovite + biotite (below 700°C).
G – H	Quartz + plagioclase + garnet + muscovite + biotite (below 700°C).
H	Quartz + plagioclase + garnet + muscovite (below 700°C).
H – J	Quartz + plagioclase + garnet + muscovite + sillimanite (below 700°C).
F – J	Quartz + plagioclase + potash felspar + sillimanite + biotite and/or garnet (above 700°C).
X	Boundary between corundum normative and diopside normative compositions using the C.I.P.W. scheme.



P_{H_2O} = about 5Kb.

For explanation of lettering see facing page.

FIG. 33.

almost entirely to high-level plutons. Muscovite-bearing granites often form the end-products of high-level magmatic differentiation sequences, for example the G4 granite of the Rosses Ring Complex (Mercy 1960 b), but are sometimes found in deeper intrusion complexes. In the latter setting, they usually contain less than 5% of muscovite and the assumption commonly made that they have formed by assimilation by the magma of aluminous sedimentary material need not be correct. It is likely that a less aluminous magma, usually of biotite-hornblende - granodiorite composition often differentiates in such a way as to give a biotite - or biotite-muscovite-adamellite liquid without assimilation of aluminous material. On the basis of such evidence it is suggested that the point P (Fig. 33) may be a reaction point and Q a eutectic. If this is so, it follows that partial melting of aluminous sedimentary material could not give rise to a liquid less aluminous than Q.

The relation between the $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratios of granitic liquids and the solid phases with which they co-exist in equilibrium is not adequately known. The differentiation trends for calc-alkali magmatic series drawn by Nockolds and Allen (1953) show a rather slight iron enrichment throughout the various sequences and it seems likely that the $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio of a granitic partial melt would be only slightly higher than that of the material from which it was derived under the conditions of p_{O_2} usually met with in orogenic belts/

belts.

(b) The behaviour of an aqueous vapour phase: Little experimental work relevant to the possibility of granitisation by the agency of a water-rich vapour phase rather than a silicate melt has been published. The only relevant material is given in Tuttle and Bowen (1958, pp. 85-91), Mackenzie (1960, pp. 382-385) and Orville (1963, p. 203 and pp. 225-235).

Tuttle and Bowen (1958, pp. 85-89) showed that in the system $K_2O - Al_2O_3 - SiO_2 - H_2O$ at 1Kb. the quartz-felspar join is a thermal barrier, and that whereas there is a eutectic on the Al-rich side of the join, the liquidus temperature steadily falls in more and more alkaline compositions until it drops below the critical temperature of water. Thus alkaline silicate liquids merge gradually into aqueous vapours, but in aluminous compositions the two phases are distinct. Also (pp. 89-91) it was found that if a granitic liquid was left open to attack by a current of aqueous vapour, the order of preferential removal, starting with the most mobile constituents, was SiO_2, $KAlSi_3O_8$, $NaAlSi_3O_8$, $CaAl_2Si_2O_8$, ferromagnesian components, $Ca_3(PO_4)_2$. At no stage, however, did mullite or nepheline crystallise from the liquid. A similar experiment carried out by Mackenzie (1960) showed that, when a powder of granitic composition was exposed to the action of aqueous vapour, the preferential abstraction of SiO_2 and $KAlSi_3O_8$ was sufficient to cause nepheline to crystallise where the vapour had been most active, while, if crystalline/

crystalline albite was exposed to attack by aqueous vapour, nepheline was left behind. Other workers (quoted by Orville 1963, p. 203) showed that, by varying the pH of the vapour phase, muscovite (if the vapour was acid) or nepheline (if the vapour was alkaline) could be produced from alkali felspar.

Orville (1963) demonstrated that, in the presence of aqueous vapour containing KCl and NaCl, a temperature gradient from 630 to 600°C acting on equal amounts of alkali felspar at opposite ends of the gradient caused the amount and K/Na ratio of the felspar in the cooler part of the tube to increase. In a brief investigation of part of the system $KAlSi_3O_8$ - $NaAlSi_3O_8$ - $CaAl_2Si_2O_8$ - SiO_2 - KCl - NaCl - H_2O it was shown that, if the Na/(Na + K) ratio of the vapour remained constant, the Na/(Na + K) ratios of rocks through which the vapour passed would change in such a way as to produce a constant Na/Ca ratio within the rocks. From Orville's (1963, Fig. 14) data, a vapour with a K/(Na + K) ratio of 0.4 would be in equilibrium with a plagioclase An_{20} at 680°C and 2000 bars, but changes in pressure and temperature might change this considerably. Thus the proportions of the two felspars in different bands of varying original Na/Ca ratio might come to vary enormously, though the felspar compositions would become constant over wide areas. Throughout Orville's experiments, the solid phases remained within the system $KAlSi_3O_8$ - $NaAlSi_3O_8$ - $CaAl_2Si_2O_8$ - SiO_2 , and nepheline, muscovite, etc., were not recorded.

The/

The experimental work which has been done leads to the conclusion that whereas an aqueous vapour phase could produce many of the phenomena of metamorphic differentiation and is almost certainly involved in the fenitisation which often accompanies nepheline syenite intrusions, there is no factor in the mechanism of silicate transfer by aqueous vapours, which could account for the particular narrow range of bulk compositions characteristic of granitic rocks.

(c) Ionic diffusion in the solid state: The possibility that ionic diffusion in the solid state could effect granitisation is now generally discounted. The existence of continuous composition gradients over large distances is unprovable, and there seems no reason to suppose that such gradients would be exactly those required to promote granitisation rather than some other kind of metasomatism. In addition, the few experiments on solid diffusion which have been carried out suggest that such a process would be far too slow to give rise to most granitic bodies in a reasonable length of time; movement of material over distances of a few centimetres is all that can be expected of solid diffusion.

The evidence so far adduced for solid diffusion is capable of an alternative explanation. In some cases, for example the results quoted by Reynolds (1946), the evidence can be interpreted as demonstrating the reaction of a granitic magma with metasedimentary and meta-igneous material, while in others

outward/

outward diffusion or convection of a silicate-bearing vapour phase can provide a feasible interpretation of contact or metamorphic phenomena.

2. THE GEOCHEMICAL EVIDENCE CONCERNING THE ORIGIN OF THE GRANITIC GNEISS

The evidence quoted above (pp. 144-52) leads to the conclusion that all granitic bodies have been formed through the agency of a silicate melt, and are therefore, in the widest possible sense, "magmatic" in origin. This does not mean to imply, however, that the whole of any particular granitic body was ever liquid or that the "magma" was necessarily derived by crystallisation differentiation of a more basic liquid. Evidence for the existence of a silicate melt "magma" at some stage in the formation of the Ardgour granitic gneiss is given by several of its geochemical features, which can also be used to elucidate the composition and behaviour of the melt.

The average bulk composition of the granitic gneiss is very similar in respect of the normative (C.I.P.W.) Q-Ab-Or proportions to that likely to have the lowest melting-point at water vapour-pressures in excess of 3Kb (see pp. 55-59). This renders it probable that any liquid involved in the formation of the granitic gneiss was at a temperature very little if at all above that at which complete solidification would ensue. This is confirmed by the crystallisation temperatures found from the distribution/

distribution of Na between coexisting felspars (pp.118-124). The granitic gneiss samples gave crystallisation temperatures of 570-700°C (average 630°C), while the surrounding metasediments gave 545-630°C (average 575°C). This is consistent with temperature estimates derived from the mineral assemblages present (pp. 160-167) and strongly suggests, within the limits of experimental error, that the granitic gneiss reached the temperature, estimated at 660°C (Tuttle and Bowen 1958 p. 82, von Platen 1965), at which partial melting would occur under the conditions of pressure believed to have prevailed in Ardgour. The metasediments, however, probably remained in the solid state throughout.

The granitic gneiss has a lower K/Rb ratio, and higher K/Ba and Ca/Sr ratios than the surrounding metasediments (Figs. 17-20, p. 98), and this is thought to confirm that a relatively low temperature silicate melt was involved in the origin of the granitic gneiss.

In contrast to the Sutherland migmatites (Read 1931, Cheng 1943), the Ardgour granitic gneiss contains no obvious relics of any metasedimentary "host rock", but the modes of origin for the body suggested by Harry(1953) and Dalziel (1963a, pp. 196-208) make it essential to consider the possibility that the granitic gneiss was formed, at least in part, from metasedimentary material similar to that by which it is now surrounded.

The composition of the granitic gneiss differs from that of the/

the surrounding metasediments in that, if the norms are re-calculated according to the scheme of Table XIV (Appendix), the metasediment analyses nearly all show normative muscovite, whereas the average granitic gneiss contains small quantities of normative hornblende, though modal hornblende is almost entirely absent from the granitic gneiss. Thus the granitic gneisses lie close to P in Fig. 33, while the metasediments lie to the right of Q. Thus if the relations of Fig. 33 are correct (see p.147), the granitic gneiss cannot be produced by partial melting of metasedimentary material similar to that exposed in Ardgour, or by reaction of a partial melt produced in this way with more metasedimentary material of a similar type. A liquid capable of forming the granitic gneiss by reaction with or assimilation of metasedimentary material could, however, be derived by partial melting of many of the more acid rock types included in the Lewisian, where the assemblage quartz + andesine + biotite + hornblende $\pm$ small quantities of potash felspar is common; the formation of a melt rich in the $KAlSi_3O_8$ component from material whose only K-bearing mineral is biotite has been described by von Platen (1965). The source material for a partial melt of the correct Ca/Na ratio to form a granitic gneiss whose plagioclase, An_{20-25}, is of the same composition as that of most of the surrounding psammites and only slightly more sodic than that of the Ardgour pelites, would probably have to be considerably more calcic, and might contain a plagioclase near An_{35}/

An_{35}.

The $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio of the granitic gneiss is almost constant at 0.78, compared with that of the metasediments, which varies from 0.62 to 0.70 (Table 27, p. 96), being lowest in pelitic rocks. The silicate liquid ("magma"), involved in the origin of the Ardgour granitic gneiss must therefore have had an $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio of at least 0.78, and the greater the quantity of metasedimentary material which contributed to the origin of the granitic gneiss the higher it must have been, unless large quantities of Mg were removed by some unknown agent from the neighbourhood of the granitic gneiss. The material from which the liquid was derived must also have had a very high $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio (see p. 148); one of the few possible source rocks would be fairly acid Lewisian gneiss interspersed with many eulysite bands.

The evidence presented in the last two paragraphs, together with the similar proportions of normative (C.I.P.W.) quartz, orthoclase, and albite in the granitic gneiss and the liquid involved in its origin, promotes doubts as to whether any part of the granitic gneiss is derived from a metasedimentary host. These doubts are strengthened by the much greater homogeneity of the granitic gneiss compared with the metasediments (compare the standard deviations in Tables I-X, Appendix). However, the amounts of $(Fe^{2+} + Fe^{3+} + Mn + Mg)$, P, and Zr are higher than in adamellites/

adamellites and granites formed as the end product of crystallisation differentiation, and it is thought unlikely that the amounts of biotite, apatite, and zircon present in the granitic gneiss could have dissolved in a low-temperature anatectic melt. Whether these minerals are derived from a metasedimentary host or were brought up in suspension by a silicate melt would be difficult to establish geochemically. The fact that the average granitic gneiss contains more Zr than the average of the metasediment analyses need not mean that all of the zircon in the granitic gneiss could not be derived from a metasedimentary host, because the tendency of certain thin bands in the metasedimentary succession to be exceptionally rich in zircon makes estimation of the average Zr content of the metasediments almost impossible.

Difficulty was found in estimating the average composition of the Beinn an Tuim striped group, which was thought by Dalziel (1963a, p. 208) to have been replaced by the granitic gneiss in some way. Though the relative proportions of Na and K were difficult to ascertain due to doubts about the composition of pelitic bands within the group, it was found that, in addition to the geochemical features already mentioned, the proportion of (Na + K) was much lower and that of Si rather higher than in the granitic gneiss, though the amount of (Fe^{2+} + Fe^{3+} + Mg + Mn) was similar. The Beinn an Tuim striped group is certainly much closer in composition to the granitic gneiss than the Druim na Saille pelitic group or the Glen Garvan psammitic group, but the/

the differences from the granitic gneiss composition are still very considerable. The possibility that the Beinn an Tuim striped group acted as host rock to a partial melt magma formed at slightly lower depths, possibly from Lewisian material, is not rejected outright, but the evidence for the incorporation of a metasedimentary host into the granitic gneiss is considered inconclusive.

The amount of crystallisation differentiation possible for a low-temperature anatectic melt is small, but a small degree of differentiation can be demonstrated. The trend of compositional variation within the granitic gneiss is similar to that established in calc-alkali magmatic series by Nockolds and Allen (1953) except that Ca is constant and Na decreases slightly in a north-south direction instead of Ca decreasing slightly and Na remaining constant. However, the amount of crystallisation differentiation required to produce the Ardgour trend is very small compared with that needed in the case of the Strontian "granite" or the Older Granodiorite of Donegal (Mercy 1963), and the Ardgour trend could alternatively be explained as the result of the incorporation by a liquid of constant composition of varying quantities of Ardgour pelitic material. In addition, the low significance of the trend surface for Ca, Na and K within the granitic gneiss (Table 21 and pp. 75-77) emphasises the relatively small part played by crystallisation differentiation in the origin of the Ardgour granitic gneiss.

The/

The amounts of Rb, Ba, and Sr in the granitic gneiss show little variation, and, except for G103, the variation which they show is determined by variations in the proportions of the chief Rb-, Ba-, and Sr-bearing minerals, and there is only a slight correlation between Rb and K, Ba and K, and Sr and Ca or Na (Figs. 17-20) within the granitic gneiss. Compared with the metasediments, the Rb, Ba, and Sr contents of the potash felspars (Fig. 32 and pp. 139-143) show a relation consistent with the granitic gneiss having been derived from the metasediments by a process including slight crystallisation differentiation. This relation does not prove that all, or even part, of the granitic gneiss was derived from the surrounding metasediments, but strengthens the suggestion that a very slight degree of crystallisation differentiation was operative during its formation.

Previous workers (Harry 1953, pp. 294-295, Dalziel 1963a, Fig. 24, p. 125) have shown that the earliest generation of pegmatites and aplites are congenetic with the granitic gneiss, but the limited number of analyses of pegmatites and aplites from Ardgour means that no convincing geochemical deductions regarding such a varied group of rocks can be made. Two analyses of concordant pegmatites, including what was thought to be the equivalent amount of biotitic selvedge, were performed and may be compared with analyses of ordinary granitic gneiss from the same locality. The concordant pegmatites G222 and G262A (Table XII, Appendix) come from the same localities as G222A (Table VI) and G262 (Table VII/

VII) respectively. The composition of the concordant pegmatites is very similar to that of the surrounding granitic gneiss, and the differences in the amounts of Ti, Fe^{3+}, Fe^{2+}, Mn, Mg and P are probably due to the difficulty of estimating the correct ratio of the amounts of pegmatite and selvedge material. It is notable that Rb, Ba and Sr contents of the pegmatites are very similar to those of the neighbouring granitic gneiss, and show no signs of having undergone differentiation compared with the granitic gneiss.

One aplite sample was analysed (G274A, Table XII). In contrast to the very sodic aplite analysed by Dalziel (1963a, Appendix II, No. 7), G274A resembles the granitic gneisses in the relative proportions of normative quartz, albite and orthoclase (Fig. 7, p. 56), though it has excess Al (the recalculated norm has muscovite, see Table XV). Compared with the granitic gneiss it also has a lower $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio of 0.69, to the 0.78 of the granitic gneiss, and is very slightly depleted in Rb and enriched in Ba and Sr. This suggests that it could have crystallised from a liquid at a slightly higher temperature than that at which the granitic gneiss was formed, but conclusions based on one analysis are unlikely to be significant.

The geochemical evidence indicates that a relatively low-temperature (about 660°C) silicate melt, produced at a slightly greater depth than at present exposed, possibly by partial melting of acid Lewisian gneiss interspersed with eulysite bands, played a major part in the origin of the Ardgour granitic gneiss.

A/

A relatively small amount of crystallisation differentiation, which did not affect the concordant pegmatites more than the surrounding granitic gneiss, was involved. The reaction of the granitic gneiss with, or its assimilation of a metasedimentary host rock cannot be proved, but if there was such a host rock it was much more likely to have been the Beinn an Tuim striped group than either of the other groups of metasediments in the Ardgour district. The oligoclase gneisses were shown to have been isochemically metamorphosed (pp. 90-95), and do not represent a "basic front" produced during the formation of the granitic gneiss.

3. PETROGRAPHIC AND STRUCTURAL EVIDENCE CONCERNING THE ORIGIN OF THE GRANITIC GNEISS.

(a) The metamorphic grade of the Ardgour rocks: It has been established (Dalziel and Brown 1965) that the granitic gneiss crystallised prior to and possibly during the early part of the second (F2) fold episode, and that sillimanite crystallised in certain rocks in the area at the same period and remained stable throughout F2 and probably until F3. However, the exact significance of the mineral assemblages present has not been discussed adequately.

The assemblages present will be described in terms of the mineral facies classification of Turner and Verhoogen (1960, pp. 509-560). It should be remembered that a mineral facies or sub-facies is an assemblage of mineral assemblages and that the mineral assemblage/

blage characteristic of a given facies will only occur in a restricted range of bulk composition, for example the assemblage hornblende-alamandine-(andesine or labradorite) which characterises the almandine-amphibolite facies is only developed in rocks of basaltic composition. Also, a metamorphic subfacies is often defined by a change of phase relations in rocks of one composition, whereas an assemblage of a different composition remains stable in several subfacies, for example the assemblage hornblende-almandine-plagioclase is stable throughout the almandine amphibolite facies, though four subfacies can be distinguished within that facies on the basis of different assemblages in pelitic rocks.

(i) Pelitic and quartzofelspathic compositions. The commonest mineral assemblages, which constitute practically all of the granitic gneiss, pelites, and psammites, are:

(1) Quartz + plagioclase + potash felspar + biotite + muscovite.

(2) Quartz + plagioclase + muscovite + biotite + garnet.

In addition, the sillimanite-bearing psammites which occur in the Glen Garvan psammitic group display the assemblage

(3) Quartz + plagioclase + biotite + muscovite + garnet + + sillimanite, in which the presence of biotite is due to the instability of a garnet sufficiently magnesian to reflect the $Fe^{2+}/(Fe^{2+} + Mg)$ ratio of the rock.

Dalziel/

Dalziel (1963a, p. 137), however, reported the presence of staurolite in a few rocks; it grew after the sillimanite with which it occurs and before the F3 movements. Its occurrence may indicate that after the formation of the granitic gneiss the metamorphic grade decreased slightly.

Some of the granitic gneisses which belong to assemblage (1) above contain, in addition, a few garnets. These are considered on textural grounds (Dalziel 1963a, Appendix I, p. xxvii) to be in disequilibrium with the rest of the rock, and this is confirmed by the fact that coexisting primary muscovite shows no signs of breakdown. These garnets could be xenocrystal relics formed during incorporation of a garnetiferous metasedimentary host rock into the granitic gneiss, but this would be very difficult to prove. The signs of replacement by groundmass felspar are a pointer in this direction, however, because garnet would not be stable with the liquid responsible for the formation of the granitic gneiss (see p. 154).

The one recorded occurrence of kyanite in the area (Dalziel 1963a, p. 137) is very uncertain.

(ii) Basic compositions. The hornblende schists and sometimes the granitic gneiss close to the contact with a hornblende schist body show the assemblages:

(4) Quartz + plagioclase + potash felspar + biotite + hornblende.

(5)/

(5) Quartz + plagioclase + biotite + hornblende + garnet.

The amounts of K-bearing minerals in the hornblende schists are often very low.

(iii) Calcareous compositions. The calc-silicate rocks present the greatest problems. The usual assemblage is:

(6) Quartz + plagioclase (An_{71} or more sodic plagioclases only) + hornblende + diopside + potash felspar + epidote (if plagioclase is An_{65-71}).

However, in addition to these minerals, garnets are often present in the calc-silicates. On the basis of colour alone, Kennedy (1949), Dalziel (1963a) and Brown (1964) decided that they were almandines. Kennedy (1949) suggested that they might contain considerable amounts of grossular in solid solution with almandine, but natural almandines from amphibolite facies rocks never have more than 12% CaO, that is 32% grossular (unpublished work quoted by Atherton, 1965, p. 288), while even at 950°C and 10 Kb (Yoder and Chinner 1960), grossular will accept only 6 to 7% of (almandine + pyrope). Consideration of the likely phase relationship shown in Fig. 34 shows that almandine could not possibly coexist with diopside, and could coexist with hornblende only in assemblage (5) above. A grossular garnet would, however, be stable in the assemblage

(7) Quartz + potash felspar + plagioclase + diopside/

diopside + grossular ± epidote.

Phase rule considerations suggest that there must then be two calcareous assemblages, (6) and (7) above, unless the amount of K in the rocks is sufficiently small to be entirely accommodated within the hornblende structure, when the assemblage

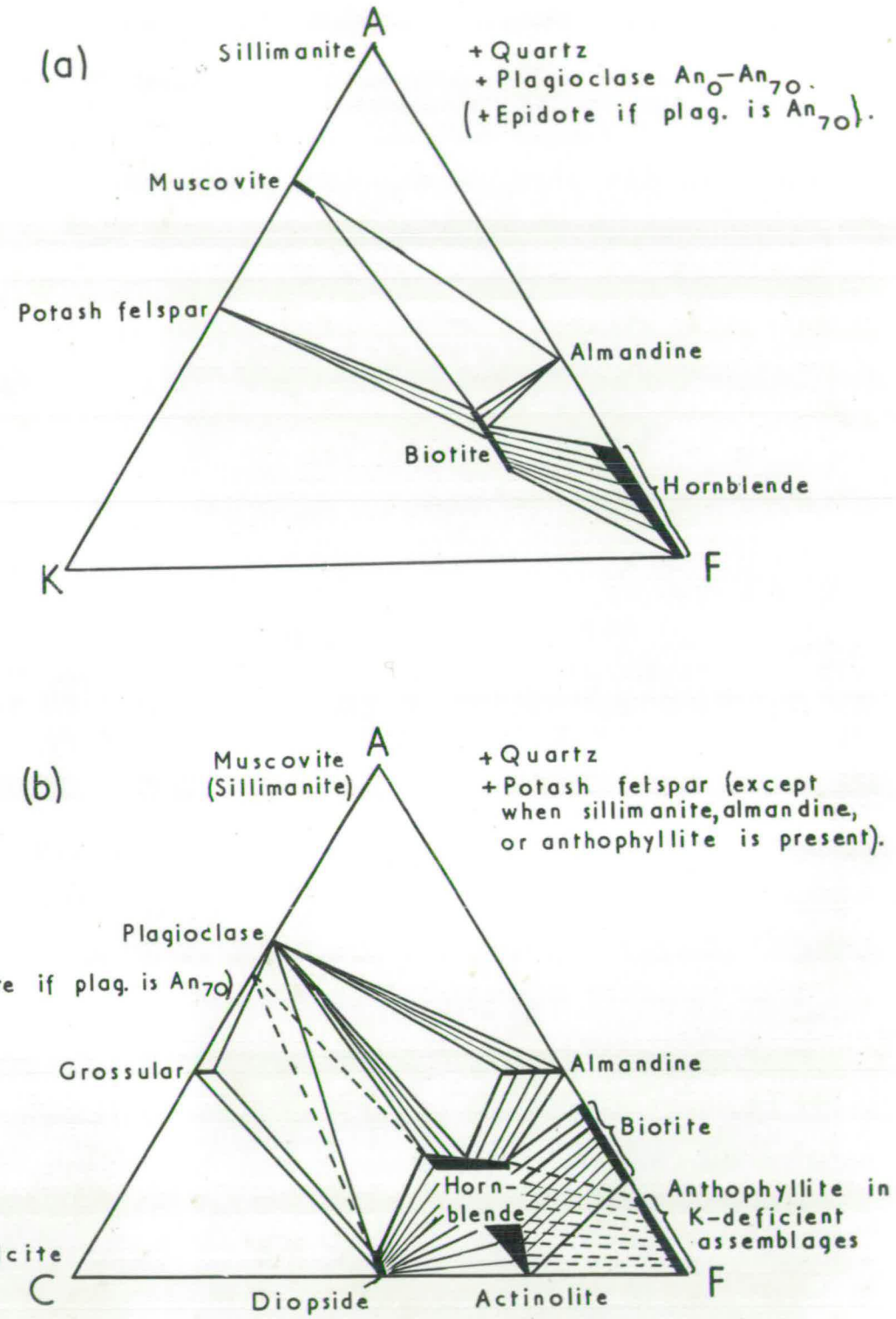
(8) Quartz + plagioclase + hornblende + diopside + grossular ± epidote might be stable, the hornblende acting as the K-bearing phase.

The presence of assemblage (5) above places the Ardgour rocks in the almandine-amphibolite facies (Turner and Verhoogen 1960, p. 544) and the other assemblages, whether or not they contain hornblende or almandine, are consistent with this classification. On the basis of assemblage (3) the Ardgour rocks can be assigned to the sillimanite-almandine-muscovite subfacies of the almandine-amphibolite facies (Turner and Verhoogen 1960, pp. 548-549). Fig. 34 gives ACF and AKF diagrams for this subfacies.

Thus, as far as the mineral assemblages developed in the rocks are concerned, the conditions of metamorphism were the same as in the migmatite areas of Central Sutherland (Read 1931, pp. 119-129) and Glen Clova and Glen Esk (Barrow 1893, pp. 343-346). However, in certain parts of the Dalradian the assemblage potash felspar + sillimanite occurs (Francis 1956, pp. 357-359) in the highest grades of metamorphism. This assemblage belongs to the sillimanite-almandine-orthoclase subfacies of the almandine-amphibolite facies, but as Turner and Verhoogen (1960, p 552) point out, the mineral assemblages of the "sillimanite zone" can be

FIGURE 34.

AKF and ACF diagrams for the sillimanite - almandine - muscovite subfacies of the almandine - amphibolite facies of regional metamorphism.



$A = Al_2O_3$, $K = K_2O$, $C = CaO$, $F = (FeO + MgO)$.

FIG. 34.

exceedingly complex and rocks containing potash felspar, muscovite, and sillimanite can be found, in which the different phases have shown different stability relationships at different times during metamorphism. None the less, the general scarcity of sillimanite and prevalence of muscovite in the Ardgour rocks makes it likely that they remained within the sillimanite-almandine-muscovite subfacies throughout, thereby restricting sillimanite to the rocks with the greatest amounts of (C.I.P.W.) normative corundum.

The lower temperature limit for the sillimanite-almandine-muscovite subfacies is defined by the breakdown of staurolite, which occurs at about 600°C (Francis 1956, Fig. 1), and the upper limit is defined by the breakdown of muscovite. The reaction

$$KAl_3Si_3O_{10}(OH)_2 \longrightarrow KAlSi_3O_8 + Al_2O_3 + H_2O$$

takes place at 680°C at 3Kb water vapour pressure and at 730°C at 11 Kb water vapour pressure (Velde 1964), but the breakdown which occurs in most rocks follows the reaction

$$KAl_3Si_3O_{10}(OH)_2 + SiO_2 \longrightarrow KAlSi_3O_8 + Al_2SiO_5 + H_2O,$$

which occurs at a temperature 15°C lower than that studied by Velde at the same water vapour-pressure (Yoder and Eugster 1955, pp.263-265).

The upper pressure limit is given by the sillimanite-kyanite transition, which occurs at 11 Kb total pressure at 600°C (Pitcher 1965), and the lower by the upper limit of the hornblende-hornfels facies/

facies, given by Turner and Verhoogen (1960, p. 519) as 3Kb water vapour pressure. Hence the Ardgour rocks crystallised at 600 - 700°C, P_{total} = 3 - 11 Kb, P_{H_2O} = at least 3Kb.

The coexistence of a plagioclase An_{65-71}, with an epidote-clinozoisite mineral in the calc-silicate rocks is noteworthy. Turner and Verhoogen (1960) say that anorthite is stable in the sillimanite-almandine-muscovite and sillimanite-almandine-orthoclase subfacies, and Kennedy (1949) correlates the kyanite and sillimanite zones of regional metamorphism with the stability of anorthite and the absence of an epidote mineral, the assemblage clinozoisite + acid plagioclase (andesine or oligoclase) being characteristic of lower grades of metamorphism. However, most workers have found that in a series of metamorphic rocks of increasing grade, the composition of the plagioclase coexisting with an epidote-clinozoisite mineral varies continuously from An_{25} upwards, though there is a sharp break between An_2 and An_{25}. The temperature and pressure conditions under which plagioclase of any specific composition coexists with an epidote-clinozoisite mineral are not known, however, and it is not safe to assume that the clinozoisite in the Ardgour rocks is necessarily all secondary in origin. Textural evidence (Dalziel 1963a, pp. 144-145) does not give a clear answer as to whether the clinozoisite crystallised at the same time as the other minerals in the calc-silicates, or represents an earlier or later metamorphism.

Thus the metamorphic conditions in Ardgour just before and during/

during the F2 episode were close to those where melting of granitic material occurs (Tuttle and Bowen 1958, Fig. 26, p. 56). However, the metamorphic grade remained unchanged right through F2 and until just before F3, and was constant over a much wider area than the immediate environs of the granitic gneiss; for example, the rocks of the Kinlochmoidart area (Brown 1964) reached the same grade as did those of the Ardgour district. All of the country from Kinlochmoidart to Ardgour lies within the "zone of injection" as drawn by Kennedy (1949, Fig. 5), but it is believed that the quartz-oligoclase lenses which give the pelites their "injected" appearance have formed by metamorphic differentiation over relatively short distances (less than 10 cm.) by the agency of a vapour phase within a system closed to all components except H_2O (see pp. 149-151). The occurrence of cross-cutting aplites and pegmatites of granitic composition suggests that a granitic liquid might have been able to form at slightly greater depths than are at present exposed over the whole of the "zone of injection".

(b) Zircon studies: A full study of the size-frequency and elongation-frequency distributions and shapes of zircons from a number of rocks was made by Dalziel (1963b). The rocks studied were: (i) Ardgour granitic gneiss, (ii) augen-gneiss facies of the Ardgour granitic gneiss, (iii) Ardgour metasediements, (iv) The intrusive Strontian "granite", (v) Augen gneisses from Carn Chuinneag and Inchbae.

The/

The zircons of the granitic gneiss were found to differ from those of the surrounding metasediments in containing more outgrown and overgrown specimens and a higher proportion of more or less euhedral grains with rounded corners; the granitic gneiss also had a higher proportion of zircons with elongations between 3 and 5, believed to result from the partial recrystallisation of the $ZrSiO_4$ in the rock. These features distinguishing the granitic gneiss zircons from those of the metasediments were slightly more marked in the zircons from the augen-gneiss facies of the granitic gneiss, which otherwise resembled those from the ordinary granitic gneiss very closely. The zircons from the Strontian "granite" and the Carn Chuinneag and Inchbae augen gneisses were nearly all euhedral and sharply terminated, and the Carn Chuinneag and Inchbae rocks contained many broken zircons which had not recrystallised during the regional metamorphism of the area. The Ardgour granitic gneiss zircons were considered to be intermediate between those from the metasediments and those from the Strontian "granite" and the Carn Chuinneag and Inchbae augen-gneisses, but to show a greater resemblance to the former than to the latter.

For this reason, Dalziel concluded that the granitic gneiss had never been through a magmatic stage (1963b, p. 360), but had formed from the metasediments in some way which involved recrystallisation of the outer parts of some of the zircon grains. However, the first part of this conclusion may not be valid in the case/

case of a relatively low-temperature (650 - 700°C) anatectic melt in which $ZrSiO_4$ would probably be almost insoluble; such a magma might only have been able to promote a slight degree of recrystallisation in any zircons with which it came into contact. The similarity of the granitic gneiss zircons to those of the metasediments suggests that they are derived from the latter, and that if the magma involved in the formation of the granitic gneiss was derived by partial melting of Lewisian material (see pp. 154-155), the zircons do not represent crystals carried upwards in suspension by the magma. Thus the shapes of the granitic gneiss zircons and the relatively large amount of zircon in the granitic gneiss (p. 156) are strong evidence for the involvement of a metasedimentary host rock in the origin of the Ardgour granitic gneiss.

(c) Structural evidence concerning the origin of the granitic gneiss:

The structural evidence for the dating of the granitic gneiss was presented by Dalziel and Johnson (1963). The foliation of the granitic gneiss is a modified S-S1 surface (bedding-plane schistosity developed during the first F1 movements) which underwent coarsening either before or early in the F2 episode (Dalziel and Johnson 1963, p. 250) and has in some places been folded by F2 folds though in others it has been brought into alignment with F2 axial planes. No F1 folds can now be distinguished in the granitic gneiss. Hence the granitic gneiss formed just prior to and during the early part of the F2 episode and was probably in an extremely plastic state during its formation (Dalziel 1963a, p. 170).

The/

The granitic gneiss nowhere has intrusive contacts against the metasediments, and was therefore considered to have remained essentially solid throughout its formation, though the presence of small quantities of silicate liquid was not ruled out.

The granitic gneiss contains more than one generation of pegmatites and aplites, of which the earliest are generally concordant with the foliation of the granitic gneiss and have been deformed by the F2 movements; this earliest generation of pegmatites and aplites is probably intimately concerned with the origin of the granitic gneiss, and this is confirmed by the chemical composition of the analysed concordant pegmatites (see p. 159). In the case of the aplites, the foliation of the granitic gneiss is sometimes truncated against them, and they form pods elongated parallel to the foliation of the surrounding granitic gneiss.

The highly localised development of the granitic gneiss close to the Loch Quoich line (see Fig. 1, p. 1) is remarkable, especially since the Beinn an Tuim striped group outcrops quite close to the granitic gneiss and at the same apparent stratigraphic horizon, and, except in a few thin bands near the granitic gneiss boundary on Beinn an Tuim, shows no signs of transition towards the texture or composition of granitic gneiss. Dalziel (1963 a, pp. 200-202) considered that the granitic gneiss probably formed on the limb of a major (?isoclinal) F1 fold which was refolded by the F2 movements, but no major F1 structures can now be recognised in Ardgour/

Ardgour due to the F2 and F3 movements. The restriction of the granitic gneiss to a narrow belt running NNE-SSW cannot be related to the F2 Meall nan Damh fold along which it is brought up if it formed by invasion of the Beinn an Tuim striped group by granitic material, but the latter part of its development may have been influenced by the movements which produced the Meall nan Damh fold.

The preservation of a modified S-S1 foliation by the granitic gneiss, and the fact that the granitic gneiss reacted to deformation during the F2 and F3 folding in much the same way as the surrounding metasediments were considered by Dalziel (1963a, pp. 198-199) to be the most important arguments against any magmatic theory for its origin, including one involving partial melting at the present level of exposure. A major influx of granitic liquid along foliation planes in the metasediments could have occurred in post-F1 and pre-F2 times, however, without unduly disturbing the S-S1 foliation of the metasediments, and if almost the whole mass of material had become crystalline before the F2 movements began, the resultant interbanded granitic and metasedimentary material would have reacted to the F2 and F3 movements as a solid and preserved an S-S1 foliation. Thus the arguments put forward by Dalzielcannot be said to rule out the possibility of a magmatic origin for the granitic gneiss, though they restrict any magmatic episode to the post-F1, pre-F2 period. A metamorphic aureole like that surrounding the Carn Chuinneag and Inchbae augen-gneisses (Harker 1954b) would not be expected if granitic liquids were intruded/

intruded into hot country-rock which was undergoing high-grade regional metamorphism.

4. A POSSIBLE MECHANISM FOR THE FORMATION OF THE GRANITIC GNEISS.

Any theory as to the origin of the Ardgour granitic gneiss must be consistent with the geochemical and structural evidence and explain the role of the early concordant pegmatites and aplites. It must also explain the relatively small number of inclusions of metasedimentary material within the granitic gneiss outcrop, and should try to account for its localised development along the "Loch Quoich line".

The geochemical evidence shows (p. 153) that a granitic liquid was almost certainly involved in the origin of the granitic gneiss and that it was not derived by partial melting of metasediments like those which surround it (pp.153-54), but there is no direct geochemical proof that a metasedimentary host rock contributed substantial amounts of material to the granitic gneiss (p. 157). However, the petrographic evidence indicates that the zircons (p.169) and the garnets (pp. 162) of the granitic gneiss are probably derived from a metasedimentary rock which most likely belonged to the Beinn an Tuim striped group and was therefore semipelitic in average composition. The structural evidence does not permit the existence of substantial quantities of liquid material at any period except for the interlude between the F1 and F2 movements. The geochemical/

geochemical and structural evidence are difficult to reconcile, and the greatest problem appears to be to account for the vast amount of homogenisation which would be necessary for the incorporation of striped group material into the granitic gneiss while preserving a modified S-S1 foliation in the granitic gneiss. Such homogenisation would almost certainly involve melting of nearly all of the quartz and felspars in the metasedimentary material or mechanical breakdown of metasedimentary bands by convection or eddies in surrounding granitic liquid, of which the former seems the more likely alternative.

The most likely origin for the granitic gneiss is considered to be the intrusion of relatively thin sill-like bodies of low-temperature potassic granitic liquid containing perhaps 9% of water along foliation planes in the upper portion of the Beinn an Tuim striped group. This implies that the liquid must have been a few degrees above the liquidus minimum temperature and must have lost heat to the surrounding metasediments very slowly during intrusion. If produced by filter-pressing of partly liquid material, the magma may have contained relatively few crystals in suspension. Successive intrusion of this type could preserve a modified S-S1 foliation in the rocks which would be capable of being folded or realigned by the F2 movements. The water vapour expelled from the granitic liquid on crystallisation could be responsible for the constancy of mineral composition in the granitic gneiss, but probably not for the homogenisation of striped group material/

material incorporated into the granitic gneiss. The compositional banding of the latter shows that on a small scale homogenisation was by no means complete, and what homogenisation has occurred may be due to the deformation effected by later fold movements.

The concordant pegmatites are difficult to explain, but they may represent either the last generation of granitic intrusive liquid or they may be a metamorphic differentiation product of ordinary granitic gneiss.

The trend of compositional variation in the granitic gneiss may indicate that the granitic partial melt originated in the northern part of the area and underwent some degree of crystallisation differentiation on moving south, or alternatively that a liquid of constant composition was intruded but that the metasediments incorporated into the granitic gneiss contained more psammitic material in the south and more pelitic material in the north. The hypothesis that the trend was caused by reworking due to partial melting during the F2 epidsode is inconsistent with the structural evidence.

The augen-gneiss facies of the granitic gneiss does not differ in chemical composition from the normal facies, but appears to have formed at the levels which were structurally lowest in post-F1, pre-F2 times. The zircon studies of Dalziel (1963b) may indicate that it formed at a slightly higher temperature than the ordinary granitic gneiss, which would result in the crystallisation of a greater proportion of potash felspar due to the more sodic composition/

composition of the potash felspar. However, this is not confirmed by the compositions of the analysed felspars (Table 29) p.120, where the augen gneiss, G62 gives a temperature similar to that of the other granitic gneisses), though there may be some doubt as to the period of crystallisation recorded by the felspar compositions, and the similarity of the compositions of the ordinary granitic gneiss and the augen gneiss suggests that they formed from liquids which crystallised at similar temperatures.

The restriction of the granitic gneiss to a relatively narrow belt of country (pp. 170-171) can readily be explained by assuming that only along this line was granitic liquid intruded into the Beinn an Tuim striped group, either because only here was the striped group underlain by material which could yield large quantities of granitic liquid, or because the temperature was unusually high along this line, due to the blanketing effect of the large thicknesses of psammitic rocks immediately to the east. Since the nature of any major F1 structures in Ardgour is unknown, the former alternative is preferred.

5. COMPARISON WITH OTHER GRANITIC ROCKS OF DISPUTED ORIGIN.

The intimate relation of rocks of granitic composition with high-grade regional metamorphic rocks in certain areas is well-known. Summaries of opinion on the origin of migmatites (that is, rocks in which an originally sedimentary and an apparently igneous portion are inextricably mixed together) are given by Turner (1948

pp./

pp. 304-317) and Turner and Verhoogen (1960, pp. 370-375 and 382-388).

The origin postulated above (p. 173) for the Ardgour granitic gneiss places it near the base of Eskola's (1933, p. 23) "zone of intrusions and of injection and potash metasomatism", while the granitic magma was derived by partial melting in his "zone of differential anatexis". The chief difference between it and an unmobilised migmatite is the homogenisation which has taken place during the incorporation of the metasedimentary host into the granitic gneiss; whether any trend of composition variation such as that found in the granitic gneiss also occurs in the granitic phase of unmobilised migmatites is unknown. The granitic gneiss differs from most intrusive granitic bodies in its potassic nature and high $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio, as well as the coarse compositional banding and small-scale inhomogeneity, though the presence of some sort of foliation in intrusive granitic bodies is not uncommon.

For comparison, the migmatites of Central and Northern Sutherland (Read 1931 and Cheng 1943), the augen-gneiss of Carn Chuinneag and Inchbae (Harker 1954a, b), and the Older Granodiorite of Donegal (Pitcher 1952, Mercy 1960a, Whitten 1960) will be discussed to bring out the factors believed by previous workers to be most relevant to the origin of the granitic rocks which they studied.

The Sutherland migmatites are developed in a belt forming most of the eastern part of the county from Bettyhill to Rogart. The/

The most fully studied parts of this immense area are the Loch Choire Complex (Read 1931, pp. 88-150) and the Bettyhill district (Cheng 1943). In these complexes, sheets and patches of granite showing an intrusive relationship to the other rocks occur, but they only form a very small part of the whole. The country rocks with which these small intrusions are associated show various stages in the transition from isochemically metamorphosed sedimentary and basic igneous rocks (which only near the "injection complex" reach sillimanite grade) to rocks with the mineralogy of biotite-muscovite granite, which still retain a totally concordant relationship with less completely altered rocks. Rocks of differing original composition appear to become transformed to a granitic composition at different rates, and quartzites, calc-silicates, and sometimes hornblende schists (derived from pre-metamorphic dolerite sills) may be found in their correct "stratigraphic" positions in a mass of rather inhomogeneous rocks consisting of quartz, felspars, and micas in greatly varying proportions. These changes, which were most pronounced in a restricted number of stratigraphic units, and decreased in intensity both upwards and downwards, were attributed by Read and Cheng to the soaking of the original metamorphic rocks in "liquids of magmatic and anatectic origin" (Cheng 1943, p. 147).

The bulk chemical changes necessary for the granitisation of such a heterogeneous set of rocks are difficult to estimate, but are probably restricted to addition of Na and K and possibly removal/

removal of Ca and Mg. Such changes could have been produced by addition of or reaction with a granitic liquid derived by partial melting of metasedimentary or basement material. The preferential enrichment of Na as against K in the "intermediate stages of granitisation", a feature also noted by Read (1927) in the Cromar district of Aberdeenshire could be the result of the action of an aqueous vapour phase upon rocks having a higher Ca/Na ratio than the surrounding rocks (see p. 150), or possibly to the squeezing out of a potash-rich partial melt from pelitic material.

Thus the Sutherland migmatites differ considerably from the Ardgour rocks in that, instead of simple addition of a granitic liquid to material which had crystallised to phases in equilibrium with the liquid, reactions such as the conversion of hornblende or garnet to biotite and sillimanite to muscovite occurred. Also the original liquid may have originally been more soda-rich and have preferentially crystallised oligoclase during these reactions, leaving a potash-rich liquid to form the intrusive granitic sills, etc. In addition, the formation of the migmatites probably occurred relatively late in the structural history of the region and the relationships of the migmatites have not been obscured by later fold epidsodes.

The Carn Chuinneag and Inchbae augen gneisses (Harker 1954a,b,) represent intrusive granites, one part at least of which is very similar in composition (Harker 1954a) to the Ardgour granitic gneiss, and which hornfelsed the surrounding rocks before the regional/

regional metamorphism of the area. Even though a relatively low grade of regional metamorphism was attained in the surrounding rocks, considerable remetamorphism of the aureole took place. Differences in the regional metamorphic minerals inside, and outside the aureole are ascribed to differences in the water contents of the hornfelses and the unmetamorphosed sediments outside. Even if the Ardgour granitic gneiss were a pre-metamorphic intrusion, against which theory Dalziel (1963a) advanced considerable evidence, it is doubtful whether its aureole would have left any traces considering the high grade maintained by the regional metamorphism in Ardgour over an extensive period.

The Older Granodiorite of Donegal (Pitcher 1952) is believed to be an intrusive body, even though its south-eastern part was designated as "migmatitic". It has a well-developed aureole, in which a high-grade contact metamorphism is superimposed on a previous greenschist-facies regional metamorphism; near the granodiorite, the biotite-muscovite-chlorite pelitic rocks have been converted to sillimanite-cordierite hornfelses, tremolite marbles have developed clinozoisite and diopside, and ophitic quartz-dolerite sills unaffected by the regional metamorphism have become amphibolites. In the south-eastern part of the "granodiorite", which is hereof quartz-diorite composition, there are numerous horn-felsed relic masses of country rock which, for the most part, retain the strike and dip which they had before the "granodiorite" was formed/

formed, and in which a "ghost stratigraphy" can easily be made out. Towards the north-west, the "granodiorite" changes in composition from quartz-diorite through granodiorite to adamellite, and the xenoliths become very much scarcer.

Pitcher (1952) was unable to decide whether the Older Granodiorite formed by partial mobilisation of country rock granitised in place or by piecemeal stoping and marginal assimilation of country rock by a granitic magma, and Mercy (1960a) considered that the geochemistry of the body was consistent with either of Pitcher's theories, though (1963, pp. 209-210) he later came to regard a magmatic theory as the more likely alternative. The marked northwesterly trend of compositional variation was ascribed largely to crystallisation differentiation, and the assimilation of country rock was only responsible for the slight departures from the trend found by Whitten (1960) in areas containing xenoliths of varying sedimentary types.

Compared with the Donegal granodiorite, the compositional trend within the Ardgour granitic gneiss is very much smaller and less well defined, and the body as a whole has a much higher $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio (pp. 67-70). In both cases the influence of the metasediments on the bulk chemical composition, whether, as in Donegal, they have been assimilated by solution by the magma, or, as probably happened in Ardgour, they have been bodily incorporated into the granitic gneiss, has been very slight. The chief differences between the Donegal and/

and Ardgour bodies can probably be accounted for by the formation of the Ardgour magma at a lower temperature and a much higher level in the crust, its gradual intrusion in several thin sheets, and the folding and metamorphism which occurred after the formation of the granitic gneiss.

In mode of formation the Ardgour granitic gneiss is in some ways intermediate between the Sutherland and Donegal examples. It differs from both in that it was formed at a relatively early stage in the structural history of the region where it was emplaced, whereas neither of the other examples was significantly affected by later folding and metamorphism. The Carn Chuinneag and Inchbae augen gneisses were not subjected to such a degree of post-consolidation metamorphism as to seriously affect their relations with the surrounding rocks, though they were sheared sufficiently to give them their augen-gneiss texture, but in the case of the Ardgour granitic gneiss, the subsequent folding and high-grade metamorphism are held responsible for the apparent disappearance of the metasedimentary host rock.

VIII. CONCLUSIONS.

The average bulk chemical composition of the granitic gneiss of Western Ardgour is very similar in its proportions of normative Q, Or and Ab to that found experimentally to have the lowest melting point of all compositions with the same Ca/Na ratio as the granitic gneiss. Under the pressures likely to have prevailed during the regional metamorphism of the area and the formation of the granitic gneiss, the minimum temperature for melting was 650 - 660°C. The mineral assemblages in the granitic gneiss and surrounding rocks indicate that the temperature throughout Ardgour at the time of granitic gneiss formation was 600-700°C, and that the total pressure was 3 - 11Kb, corresponding to depths of 10 - 33 kilometres, while the water vapour pressure was at least 3Kb. The distribution of Ca, Na and K between co-existing felspar phases in the granitic gneiss and surrounding rocks suggests that the temperature of crystallisation of the granitic gneiss lay in the range 580 - 700°C, while the surrounding rocks crystallised at 545 - 630°C.

The variation in chemical composition of the granitic gneiss from place to place within its outcrop was studied in order to decide whether there was a well-defined trend of composition variation within the granitic gneiss or whether the variation was largely random. This was first attempted by splitting the analysed samples into seven regional groups and comparing the means/

means and standard deviations of the seven groups for each element. A considerable increase in Si and K and a decrease in Na, Ba, Sr, Mg, Mn, Fe^{2+}, Ti, P and Zr from north to south was found while Ca, Fe^{3+} and Rb remained roughly constant in this direction. There was much less variability between neighbouring samples in the northern part of the granitic gneiss outcrop than in the southern part, this difference being statistically significant at the 0.1% level for Si, Ti, Al, Mg, P, Rb and Ba, and at the 1% level for Fe^{3+}, Mn, Ca, and $(Fe^{2+} + Fe^{3+})$.

The attempt to define the trend of composition variation within the granitic gneiss was continued for Si, Al, $(Fe^{2+} + Fe^{3+})$, Mg, Ca, Na and K by means of the statistical technique known as trend surface analysis. The geographical positions of the individual sample points were represented by two co-ordinates, and equations of the first, (first and second), and (first, second and third) degrees in these co-ordinates were calculated. These equations were calculated in such a way that the sum of the squares of the deviations of the actual values for any one element from the values predicted by the surface for each of the sample localities in turn was a minimum. This was done for all of the elements studied, and it was found that in all cases the linear trend surface was much more significant than the quadratic or cubic surfaces. The linear surfaces were found to be significant at the 0.1% level for Si, Al, $(Fe^{2+} + Fe^{3+})$ and Mg, at the 1% level for/

for Ca, and at the 5% level for Na and K. These linear surfaces all had a roughly E-W strike except that for Ca, which struck N-S. Ca had been found not to vary significantly in a N-S direction.

The granitic gneiss analyses were then considered independently of the locality of the samples which they represented and the distributions about the mean of the values for different elements were studied. The prediction that all elements except Si and K would be lognormally distributed was not fulfilled and, in fact, the only element whose distribution approached lognormality was Sr. In fact the trends of regional compositional variation and differences in homogeneity combined to give distributions very different from those reported by other workers who generally considered granites from one petrogenetic province or sometimes from a large arbitrarily defined area and even included granites of widely differing ages. It was therefore concluded that, though lognormality may be approached if a large and diverse enough set of rocks is considered, elements are not lognormally distributed within one granitic body and trace elements tend to show similar distributions to elements with which they are geochemically associated.

A considerable number of analyses of pelitic and psammitic metasedimentary rocks from the Ardgour district were performed. The pelites and psammites are very much more variable in composition than the granitic gneiss, and have almost certainly undergone isochemical metamorphism; their "injected" appearance is due to a process of metamorphic differentiation, probably aided by an aqueous vapour phase during crystallisation. Analysis of biotite and/

and hornblende from different layers in a banded semipelite with calc-silicate ribs suggests that equilibrium distribution of elements between coexisting minerals was attained over distances of 5-10 cm., but there is no evidence of any homogenisation of bulk compositions in the mass of pelites and psammites.

Comparison with the average bulk composition of the granitic gneiss shows that if psammitic and pelitic material were mixed together in the same proportions which they probably had in the part of the Beinn an Tuim striped group from which the granitic gneiss was believed to have been formed, there would be insufficient alkalis (the shortage of K being probably the greater), and an excess of Mg, Sr, and Ba.

Micas and felspars were separated from selected samples of granitic gneiss and psammitic and pelitic rocks, and analysed. The granitic gneiss biotites contained less Si and more Fe^{3+} and had higher $Fe^{2+}/(Fe^{2+} + Mg)$ ratios than those from the surrounding rocks. The distribution of Rb, Ba, and Sr between coexisting biotite, muscovite, potash felspar and plagioclase, or such of these phases as were present, was consistent from rock to rock, suggesting that equilibrium was attained, but the minerals from the granitic gneiss generally contained more Rb, and less Ba and Sr than the corresponding minerals from the surrounding rocks. The distribution coefficients for Rb, Ba, and Sr between any two of the minerals studied in any one rock could not be correlated with the crystallisation temperature estimated from the distribution of Na between the/

the coexisting felspars of that rock. An attempt to decide whether the geochemical behaviour of Sr in metamorphic and igneous rocks was nearer to that of Ca or Na was inconclusive. With regard to the distribution of Rb, K, Ba, Na, Sr and Ca, the minerals biotite, muscovite, potash felspar, plagioclase and hornblende (in that order) form a geochemical series in which the elements nearest the left in the series listed above are most strongly concentrated in the minerals nearest the left in the mineral sequence.

These geochemical findings strongly indicate that a relatively low-temperature (about 660°C) silicate melt, produced at a level below that at present exposed from material containing less Al and having a higher $(Fe^{2+} + Fe^{3+})/(Fe^{2+} + Fe^{3+} + Mg)$ ratio, was involved in the origin of the granitic gneiss, and that a relatively small degree of crystallisation differentiation occurred during its formation. There is, however, no direct geochemical evidence for the incorporation of a metasedimentary host rock into the granitic gneiss. These conclusions are difficult to reconcile with the structural and petrographic evidence which is in favour of the derivation of much of the granitic gneiss from semipelitic metasediments and shows that substantial amounts of granitic liquid can only have been present during a restricted part of the structural and metamorphic history of Ardgour. A hypothesis involving the intrusion of thin veins or sills of granitic liquid into the top of the Beinn an Tuim striped group, followed by partial homogenisation during subsequent episodes of folding and metamorphism, is favoured.

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A P P E N D I X

CHEMICAL ANALYSES

AND NORMS.

(TABLES I - XX).

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	VIII	Whole Rock Analyses.	Pelitic schists and oligoclase gneisses.
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TABLE I.

Whole Rock Analyses. Granitic Gneisses. Group I.

Weight Percentages.

	GP112	G116	G117	G121	G122	G123	G131	G132	G133	G141	G142	G151	G152	G161	G162	G163
SiO_2	71.4	71.8	73.1	72.8	74.9	75.6	76.3	75.3	71.3	75.4	70.4	73.3	71.9	76.1	70.6	74.0
TiO_2	0.67	0.50	0.50	0.45	0.42	0.28	0.38	0.36	0.74	0.50	0.86	0.40	0.60	0.33	0.75	0.49
Al_2O_3	13.1	13.1	12.7	13.1	11.6	12.4	11.9	12.4	12.8	11.4	13.5	13.8	13.2	12.1	12.6	12.5
Fe_2O_3	1.73	1.38	1.22	1.04	1.35	0.55	0.58	1.10	0.86	1.38	1.23	0.64	1.60	0.73	1.98	0.51
FeO	2.30	1.73	1.68	1.70	1.05	1.03	1.36	1.01	3.94	1.71	3.30	1.64	2.32	1.31	3.01	2.91
MnO	0.081	0.066	0.056	0.060	0.039	0.026	0.034	0.036	0.086	0.057	0.080	0.045	0.097	0.037	0.076	0.049
MgO	0.55	0.38	0.41	0.48	0.30	0.22	0.39	0.27	0.81	0.36	1.13	0.37	0.57	0.19	0.62	0.50
CaO	1.48	1.50	1.60	1.51	1.24	1.25	1.05	1.09	2.00	1.04	2.31	0.99	2.15	1.17	2.16	1.37
Na_2O	2.73	3.30	3.40	3.34	2.99	3.02	2.80	2.78	2.82	2.57	2.91	3.53	3.28	3.08	3.25	3.11
K_2O	5.38	4.98	4.64	4.67	5.04	4.85	5.28	5.68	3.52	5.03	4.24	5.17	3.76	5.15	4.27	4.73
P_2O_5	0.152	0.104	0.122	0.103	0.100	0.061	0.059	0.080	0.149	0.080	0.162	0.072	0.161	0.085	0.170	0.098
H_2O^+	0.59	0.45	0.21	0.22	0.19	0.25	0.28	0.29	0.81	0.25	0.54	0.37	0.59	0.18	0.44	0.36
Total	100.16	99.36	99.60	99.47	99.30	99.55	100.42	100.41	99.87	99.55	100.75	100.39	99.96	100.43	100.01	99.99
T.Fe*	4.29	3.30	3.09	2.93	2.52	1.69	2.09	2.22	5.24	3.28	4.90	2.46	4.18	2.19	5.33	2.94
Rb (ppm)	180	174	153	212	216	166	204	206	93	193	142	208	111	161	115	176
Ba (ppm)	1080	950	750	630	470	460	470	720	950	640	1200	620	1050	570	920	800
Sr (ppm)	175	135	105	150	130	100	100	105	205	150	155	80	180	75	225	170
Zr (ppm)	458	372	366	322	218	164	276	180	644	396	640	266	444	252	642	304

One-cation Percentages.

	GP112	G116	G117	G121	G122	G123	G131	G132	G133	G141	G142	G151	G152	G161	G162	G163	Arith. Mean.	S.D.‡
Si^{4+}	67.77	68.27	69.13	68.85	71.37	71.65	71.87	70.96	68.07	71.93	66.57	68.56	68.07	71.52	66.99	69.99	69.47	1.79
Ti^{4+}	0.48	0.35	0.35	0.32	0.30	0.20	0.27	0.25	0.53	0.36	0.61	0.28	0.43	0.23	0.54	0.35	0.37	0.12
Al^{3+}	14.69	14.70	14.15	14.65	13.08	13.87	13.24	13.74	14.36	12.79	15.07	15.26	14.79	13.38	14.08	13.97	14.11	0.70
Fe^{3+}	1.24	0.99	0.87	0.74	0.97	0.39	0.41	0.78	0.62	0.99	0.87	0.44	1.15	0.51	1.41	0.36	0.80	0.31
Fe^{2+}	1.82	1.38	1.33	1.35	0.84	0.81	1.07	0.80	3.14	1.36	2.61	1.28	1.84	1.03	2.39	1.73	1.55	0.66
Mn^{2+}	0.06	0.05	0.05	0.05	0.03	0.02	0.03	0.03	0.07	0.05	0.06	0.03	0.07	0.03	0.06	0.04	0.046	0.015
Mg^{2+}	0.76	0.54	0.57	0.68	0.43	0.31	0.55	0.37	1.15	0.51	1.59	0.51	0.80	0.27	0.87	0.71	0.66	0.38
Ca^{2+}	1.51	1.52	1.62	1.53	1.26	1.27	1.06	1.10	2.36	1.06	2.03	0.99	2.18	1.18	2.19	1.39	1.52	0.43
Na^{+}	5.02	6.08	6.23	6.12	5.52	5.54	5.11	5.08	5.22	4.75	5.33	6.40	6.02	5.61	6.16	5.69	5.62	0.48
K^{+}	6.51	6.03	5.60	5.63	6.12	5.86	6.34	6.82	4.36	6.12	5.11	6.17	4.54	6.17	5.17	5.70	5.77	0.66
P^{5+}	0.12	0.09	0.10	0.09	0.08	0.05	0.05	0.06	0.12	0.06	0.13	0.06	0.13	0.07	0.14	0.08	0.089	0.031

*T.Fe = Total iron as Fe_2O_3

‡ S.D. = standard deviation uncorrected either for experimental error or for small sample effect. No sample excluded in calculating mean or S.D.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE II.

White Rock Analyses. Granitic Gneisses. Group II.

TABLE II.

Weight Percentages.

	G51	G52	G53	G54	G55	G56	G63	G64	G65	G101	G102	G103	G105	G111	G112	G113
SiO_2	72.7	71.4	74.9	71.7	71.6	73.3	72.4	71.6	69.7	76.2	73.1	73.5	70.3	73.3	75.7	71.5
TiO_2	0.39	0.65	0.62	0.59	0.48	0.67	0.61	0.58	0.64	0.33	0.62	0.22	0.60	0.53	0.37	0.68
Al_2O_3	13.2	12.8	10.7	12.9	13.4	12.0	12.6	13.1	14.4	11.9	12.9	14.3	13.1	12.8	12.0	12.9
Fe_2O_3	0.62	0.98	1.87	1.01	1.63	1.35	0.98	1.13	1.40	0.33	0.80	0.29	1.60	0.95	0.99	1.61
FeO	1.93	3.39	2.37	2.96	1.64	2.36	2.59	2.71	2.33	1.39	2.65	0.65	2.40	2.09	1.03	2.83
MnO	0.051	0.072	0.087	0.078	0.060	0.078	0.060	0.073	0.088	0.029	0.061	0.032	0.060	0.051	0.042	0.070
MgO	0.23	0.61	0.65	0.61	0.42	0.80	0.67	0.63	0.65	0.25	0.64	0.20	0.61	0.55	0.20	0.87
CaO	1.01	1.42	1.72	1.43	1.44	1.44	1.35	1.55	1.63	0.99	1.28	1.49	1.36	1.22	1.22	2.05
Na_2O	2.92	2.92	2.48	2.92	3.57	3.24	2.31	2.99	3.02	2.64	2.97	3.88	3.46	3.10	3.03	2.94
K_2O	5.61	4.22	3.60	4.60	4.95	3.88	4.72	4.84	3.97	5.47	4.15	4.36	4.85	5.26	4.70	3.88
P_2O_5	0.129	0.188	0.191	0.192	0.200	0.169	0.139	0.194	0.199	0.050	0.138	0.045	0.141	0.105	0.081	0.144
H_2O+	0.46	0.57	0.40	0.40	0.25	0.40	0.58	0.55	0.55	0.48	0.67	0.42	0.49	0.39	0.15	0.68
Total	99.27	99.20	100.01	99.47	99.71	99.71	99.24	99.90	99.51	100.09	99.98	99.58	99.04	100.33	99.53	100.13
T.Fe*	2.76	4.75	4.50	4.30	3.45	3.97	3.86	4.14	4.99	1.87	3.74	1.03	4.27	3.26	2.13	4.76
Rb (ppm)	168	165	118	204	172	173	161	172	197	201	176	122	195	254	189	165
Ba (ppm)	920	660	1380	900	860	880	1080	860	750	560	940	2610	770	530	380	730
Sr (ppm)	180	145	240	130	125	205	180	165	115	110	200	565	105	85	120	120
Zr (ppm)	284	410	476	450	410	430	486	430	526	150	362	60	424	330	212	568

One-cation percentages.

	G51	G52	G53	G54	G55	G56	G63	G64	G65	G101	G102	G103	G105	G111	G112	G113	Arith. Mean.	S.D.‡
Si^{4+}	69.19	68.50	72.13	68.45	67.64	69.74	68.65	67.64	65.40	72.37	69.53	69.13	67.05	69.27	71.88	68.05	69.04	1.80
Ti^{4+}	0.29	0.47	0.45	0.42	0.34	0.48	0.43	0.41	0.45	0.23	0.44	0.15	0.42	0.38	0.26	0.49	0.38	0.10
Al^{3+}	14.86	14.47	12.17	14.51	14.96	13.45	14.03	14.59	15.90	12.98	14.47	15.84	14.76	14.23	13.46	14.44	14.32	0.93
Fe^{3+}	0.45	0.71	1.35	0.73	1.16	0.97	0.56	0.57	0.66	0.25	0.57	0.20	1.15	0.59	0.71	1.15	0.74	0.32
Fe^{2+}	1.54	2.72	1.91	2.36	1.29	1.88	2.20	2.37	2.86	1.10	2.11	0.51	1.91	1.67	0.82	2.25	1.84	0.56
Mn^{2+}	0.04	0.06	0.07	0.06	0.05	0.06	0.05	0.06	0.07	0.02	0.05	0.03	0.05	0.04	0.03	0.06	0.050	0.014
Mg^{2+}	0.33	0.87	0.93	0.87	0.45	1.13	0.95	0.89	0.90	0.35	0.90	0.28	0.86	0.74	0.28	1.23	0.75	0.30
Ca^{2+}	1.02	1.46	1.78	1.46	1.46	1.47	1.37	1.57	1.64	0.99	1.30	1.50	1.39	1.20	1.24	2.09	1.43	0.26
Na^{+}	5.39	5.43	4.63	5.40	6.53	5.97	5.86	5.73	6.43	5.05	5.47	7.07	6.39	5.56	5.57	5.42	5.74	0.59
K^{+}	6.81	5.17	4.42	5.60	5.96	4.71	5.79	6.00	5.55	6.61	5.03	5.23	5.90	6.23	5.69	4.70	5.59	0.70
P^{5+}	0.10	0.15	0.16	0.15	0.16	0.14	0.11	0.15	0.16	0.04	0.11	0.03	0.11	0.08	0.06	0.11	0.114	0.041

*T.Fe = Total iron as Fe_2O_3

‡ S.D. = standard deviation uncorrected either for experimental error or for small sample effect. No sample excluded in calculating mean or S.D.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE III.

Whole Rock Analyses. Granitic Gneisses. Group III.

Weight Percentages.

	G61	G62	G71	G72	G73	G74	G81	G82	G83	G91	G92	G93	G96	G97	G98
SiO_2	70.7	73.1	73.7	73.0	73.4	72.5	71.2	71.0	72.0	74.8	71.0	70.8	75.0	73.1	71.9
TiO_2	0.65	0.59	0.52	0.67	0.56	0.56	0.60	0.64	0.73	0.37	0.59	0.58	0.34	0.50	0.60
Al_2O_3	12.7	11.5	12.5	12.4	12.4	12.8	13.1	13.5	12.9	12.3	13.4	13.6	12.2	13.5	13.0
Fe_2O_3	1.48	1.11	1.01	0.99	0.84	0.77	0.64	0.73	0.97	0.50	0.78	0.59	0.45	0.45	0.64
FeO	3.26	2.35	2.52	2.84	2.73	2.90	2.70	3.44	4.13	1.59	3.29	3.38	1.55	2.33	2.51
MnO	0.087	0.058	0.087	0.065	0.057	0.065	0.068	0.069	0.110	0.036	0.078	0.078	0.038	0.062	0.059
MgO	0.70	0.67	0.50	0.73	0.51	0.48	0.78	0.55	0.65	0.30	0.66	0.64	0.34	0.54	0.51
CaO	1.48	1.35	1.46	1.27	1.20	1.46	1.24	1.36	1.41	0.83	1.44	1.48	0.97	1.32	1.43
Na_2O	2.85	2.76	3.37	2.80	2.89	3.11	3.08	3.31	2.72	2.92	3.18	3.21	2.42	3.10	3.35
K_2O	4.75	4.43	3.60	4.13	4.82	4.71	5.75	4.61	3.66	4.96	4.58	5.13	5.00	4.95	4.94
P_2O_5	0.148	0.163	0.109	0.155	0.167	0.168	0.187	0.217	0.175	0.056	0.171	0.186	0.054	0.128	0.125
H_2O+	0.78	0.66	0.62	0.74	0.50	0.49	0.68	0.62	0.71	0.48	0.53	0.50	0.47	0.57	0.59
Total	99.64	99.03	100.13	99.72	100.14	100.00	100.04	99.96	100.10	99.13	99.71	100.18	99.89	100.52	99.68
T.Fe*	5.10	3.72	3.81	4.17	3.87	3.99	3.64	4.58	5.56	2.27	4.44	4.35	2.17	3.04	3.43
Rb (ppm)	198	133	164	168	212	179	225	225	146	250	178	211	223	157	142
Ba (ppm)	800	810	520	750	700	870	950	750	940	540	810	750	500	840	860
Sr (ppm)	155	170	120	165	100	150	165	120	165	95	155	120	65	160	170
Zr (ppm)	592	358	408	348	356	352	488	460	552	210	456	428	228	280	476

One-cation Percentages.

	G61	G62	G71	G72	G73	G74	G81	G82	G83	G91	G92	G93	G96	G97	G98	Arith. Mean	S.D.‡
Si^{4+}	67.74	70.59	70.00	69.85	69.68	68.76	67.29	67.22	68.85	71.51	67.40	66.77	72.25	68.71	68.17	68.99	1.53
Ti^{4+}	0.47	0.43	0.38	0.48	0.40	0.40	0.43	0.46	0.53	0.26	0.42	0.41	0.24	0.36	0.43	0.41	0.07
Al^{3+}	14.37	13.08	14.01	13.94	13.90	14.26	14.57	15.05	14.51	13.80	15.01	15.11	13.72	14.96	14.52	14.37	0.56
Fe^{3+}	0.80	0.67	0.58	0.71	0.60	0.55	0.45	0.52	0.70	0.36	0.56	0.42	0.32	0.32	0.46	0.54	0.14
Fe^{2+}	2.82	2.07	2.14	2.27	2.17	2.30	2.13	2.73	3.30	1.27	2.61	2.66	1.25	1.83	1.99	2.24	0.53
Mn^{2+}	0.07	0.05	0.07	0.05	0.05	0.05	0.06	0.06	0.09	0.03	0.06	0.06	0.03	0.05	0.05	0.055	0.015
Mg^{2+}	1.00	0.96	0.71	1.04	0.72	0.67	1.10	0.77	0.93	0.42	0.93	0.90	0.49	0.76	0.72	0.81	0.19
Ca^{2+}	1.52	1.40	1.48	1.30	1.22	1.48	1.25	1.38	1.44	0.85	1.47	1.50	1.00	1.33	1.45	1.34	0.19
Na^{+}	5.28	5.16	6.20	5.19	5.31	5.71	5.64	6.08	5.04	5.41	5.85	5.87	4.52	5.65	6.15	5.54	0.44
K^{+}	5.79	5.45	4.36	5.04	5.83	5.69	6.93	5.57	4.47	6.04	5.54	6.17	6.14	5.93	5.97	5.66	0.63
P^{5+}	0.14	0.12	0.09	0.13	D.14	0.14	0.15	0.18	0.14	0.05.	0.14	0.15	0.05	0.10	0.10	0.121	0.037

* T.Fe = Total iron as Fe_2O_3

‡ S.D. = standard deviation uncorrected either for experimental error or for small sample effect. No sample excluded in calculating mean or S.D.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE IV.

Whole Rock Analyses. Granitic Gneisses. Group IV.

Weight Percentages.

	G11	G12	G13	G21	G22	G23	G31	G32	G33	G41	G42	G43	G171	G172	G173
SiO_2	73.0	69.8	71.0	71.4	70.2	71.7	71.7	72.6	71.2	72.2	71.4	71.6	71.4	70.4	72.6
TiO_2	0.52	0.56	0.56	0.54	0.73	0.65	0.62	0.45	0.64	0.62	0.56	0.65	0.70	0.57	0.63
Al_2O_3	13.0	13.6	14.0	13.2	13.7	12.9	13.2	13.0	13.2	12.9	13.5	12.9	13.3.	13.4	12.6
Fe_2O_3	0.87	1.08	0.89	0.78	1.30	1.72	0.59	0.56	1.89	0.32	0.87	1.17	0.83	0.74	1.23
FeO	2.94	3.07	3.04	2.78	3.12	2.74	3.58	2.23	2.38	3.96	2.62	2.88	3.51	3.31	2.60
MnO	0.062	0.086	0.077	0.064	0.092	0.095	0.082	0.072	0.088	0.079	0.069	0.089	0.075	0.078	0.076
MgO	0.53	0.65	0.66	0.49	0.63	0.50	0.54	0.42	0.64	0.51	0.53	0.50	0.92	0.60	0.53
CaO	1.31	1.73	2.23	1.36	2.16	2.06	1.52	1.29	1.85	1.44	1.55	1.49	1.78	1.76	1.79
Na_2O	3.72	3.20	4.29	3.11	3.18	3.28	3.40	3.22	2.90	2.63	3.26	3.42	3.34	3.49	3.28
K_2O	4.17	4.69	2.11	4.82	4.63	4.13	3.91	5.11	5.16	4.51	5.03	3.84	3.80	4.33	4.07
P_2O_5	0.187	0.194	0.177	0.196	0.171	0.145	0.167	0.191	0.161	0.187	0.158	0.148	0.171	0.207	0.164
H_2O+	0.34	0.48	0.51	0.49	0.47	0.43	0.64	0.47	0.41	0.60	0.44	0.59	0.43	0.56	0.41
Total	100.56	99.54	99.55	99.17	100.39	100.37	99.93	99.63	100.52	99.94	99.97	99.36	100.26	99.45	99.94
T.Fe *	4.14	4.49	4.27	3.87	4.77	4.77	4.49	3.08	4.54	4.70	3.78	4.38	4.73	4.42	4.13
Rb (ppm)	159	181	122	201	175	174	236	190	175	155	193	159	174	181	140
Ba (ppm)	600	790	330	900	1010	920	900	730	1010	770	860	640	680	800	810
Sr (ppm)	125	130	170	125	180	125	135	210	195	150	140	205	120	130	200
Zr (ppm)	372	490	422	396	580	650	454	316	500	490	390	554	328	406	476

One-cation Percentages.

	G11	G12	G13	G21	G22	G23	G31	G32	G33	G41	G42	G43	G171	G172	G173	Arith. Mean.	S.D.‡
Si^{4+}	68.42	66.59	67.19	68.11	66.18	67.81	68.05	68.80	67.19	68.87	67.37	68.43	67.35	66.85	68.81	67.74	0.71
Ti^{4+}	0.37	0.41	0.40	0.38	0.52	0.46	0.44	0.32	0.45	0.44	0.39	0.46	0.49	0.41	0.45	0.43	0.05
Al^{3+}	14.31	15.26	15.56	14.81	15.26	14.37	14.75	14.53	14.65	14.54	15.03	14.53	14.82	15.04	14.09	14.77	0.36
Fe^{3+}	0.61	0.77	0.64	0.56	0.93	1.23	0.42	0.40	1.34	0.23	0.62	0.84	0.59	0.53	0.88	0.71	0.29
Fe^{2+}	2.30	2.45	2.40	2.22	2.46	2.16	2.84	1.76	1.88	3.16	2.07	2.30	2.77	2.63	2.06	2.36	0.36
Mn^{2+}	0.05	0.07	0.06	0.05	0.07	0.07	0.07	0.06	0.07	0.06	0.06	0.07	0.06	0.06	0.06	0.063	0.007
Mg^{2+}	0.74	0.92	0.93	0.70	0.88	0.70	0.76	0.59	0.90	0.72	0.74	0.71	1.29	0.85	0.75	0.81	0.16
Ca^{2+}	1.32	1.76	2.26	1.39	2.18	2.08	1.54	1.31	1.87	1.47	1.57	1.53	1.80	1.79	1.82	1.71	0.29
Na^{+}	6.76	5.91	7.86	5.75	5.81	6.01	6.25	5.91	5.30	4.86	5.96	6.33	6.11	6.42	6.03	6.08	0.62
K^{+}	4.98	5.71	2.55	5.86	5.57	4.98	4.73	6.17	6.21	5.49	6.06	4.68	4.57	5.24	4.92	5.18	0.88
P^{5+}	0.15	0.15	0.14	0.16	0.14	0.11	0.14	0.15	0.13	0.15	0.12	0.12	0.14	0.16	0.13	0.139	0.017

* T.Fe = Total iron as Fe_2O_3

‡ S.D. = standard deviation uncorrected either for experimental error or for small sample effect. No sample excluded in calculating mean or S.D.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE V.

Whole Rock Analyses. Granitic Gneisses. Group V.

TABLE V.

Weight Percentages.

	G181	G182	G183	G184	G185	G186	G187	G191	G192	G193	G201	G202	G203	G211	G212	G213
SiO_2	72.6	72.8	72.2	70.7	72.0	71.5	72.8	71.1	72.4	70.0	71.9	70.9	70.2	73.0	71.7	70.6
TiO_2	0.45	0.54	0.61	0.67	0.69	0.64	0.54	0.62	0.55	0.60	0.59	0.64	0.67	0.52	0.59	0.56
Al_2O_3	13.1	13.2	13.0	13.3	12.6	12.9	12.4	13.3	13.0	13.5	13.0	13.7	13.4	12.6	13.1	13.4
Fe_2O_3	0.81	1.16	0.83	1.33	1.23	0.64	0.81	1.03	0.71	1.21	0.87	1.16	1.32	1.93	1.42	0.99
FeO	2.86	2.35	3.29	3.29	2.98	3.64	2.79	3.17	2.79	2.97	2.99	3.00	3.28	1.56	2.39	2.88
MnO	0.059	0.066	0.077	0.098	0.085	0.079	0.074	0.085	0.072	0.080	0.072	0.083	0.102	0.055	0.074	0.085
MgO	0.52	0.50	0.62	0.75	0.70	0.60	0.51	0.60	0.55	0.56	0.57	0.69	0.67	0.41	0.56	0.56
CaO	1.32	1.23	1.36	1.86	1.72	1.26	1.36	1.73	1.39	1.57	1.41	1.61	1.61	1.15	1.35	1.48
Na_2O	3.30	3.47	3.16	3.53	3.04	3.28	3.28	3.47	3.41	3.28	3.01	3.71	3.24	3.64	3.27	2.98
K_2O	4.57	4.69	4.41	4.00	4.18	4.42	4.70	4.35	4.49	5.18	4.68	3.81	4.59	4.27	4.58	5.55
P_2O_5	0.144	0.168	0.205	0.226	0.184	0.212	0.185	0.197	0.166	0.189	0.172	0.163	0.202	0.179	0.193	0.217
H_2O+	0.56	0.40	0.49	0.58	0.48	0.64	0.44	0.40	0.37	0.36	0.42	0.39	0.28	0.43	0.27	0.39
Total	100.31	100.63	100.23	100.30	99.85	99.88	99.84	100.00	99.97	99.47	99.68	99.80	99.48	99.74	99.47	100.12
T.Fe*	3.99	3.77	4.49	4.97	4.54	4.69	3.91	4.55	3.81	4.51	4.19	4.50	4.97	3.65	4.08	4.19
Rb (ppm)	200	183	190	166	165	181	167	161	178	198	196	175	197	110	189	216
Ba (ppm)	800	750	820	990	900	770	760	920	820	790	810	620	870	790	900	720
Sr (ppm)	165	150	140	150	200	160	135	155	170	135	115	110	140	200	135	120
Zr (ppm)	338	382	422	498	590	390	384	486	380	494	352	592	448	382	550	392

One-cation Percentages.

	G181	G182	G183	G184	G185	G186	G187	G191	G192	G193	G201	G202	G203	G211	G212	G213	Arith. Mean.	S.D.‡
Si^{4+}	68.50	68.33	68.34	66.75	68.48	67.95	69.03	67.12	68.38	66.36	68.41	66.94	66.66	69.21	68.09	66.92	67.84	0.87
Ti^{4+}	0.32	0.38	0.43	0.48	0.49	0.46	0.38	0.44	0.39	0.42	0.42	0.46	0.47	0.37	0.42	0.40	0.42	0.04
Al^{3+}	14.60	14.60	14.41	14.76	14.14	14.49	13.79	14.81	14.52	15.10	14.52	15.23	14.97	14.03	14.64	14.96	14.60	0.38
Fe^{3+}	0.57	0.82	0.59	0.95	0.88	0.46	0.58	0.73	0.50	0.87	0.62	0.82	0.94	1.38	1.02	0.71	0.78	0.22
Fe^{2+}	2.26	1.84	2.60	2.60	2.37	2.89	2.21	2.50	2.20	2.35	2.38	2.37	2.61	1.24	1.90	2.28	2.29	0.32
Mn^{2+}	0.05	0.05	0.06	0.08	0.07	0.06	0.06	0.07	0.06	0.06	0.06	0.07	0.08	0.05	0.06	0.07	0.063	0.009
Mg^{2+}	0.73	0.70	0.88	1.06	0.99	0.85	0.72	0.85	0.77	0.79	0.81	0.97	0.95	0.58	0.79	0.79	0.83	0.12
Ca^{2+}	1.33	1.23	1.38	1.88	1.75	1.28	1.37	1.75	1.41	1.60	1.43	1.63	1.64	1.17	1.37	1.50	1.48	0.20
Na^{+}	6.04	6.30	5.79	6.46	5.61	6.04	6.02	6.35	6.24	6.03	5.55	6.79	5.96	6.68	6.02	5.47	6.08	0.37
K^{+}	5.50	5.61	5.32	4.82	5.07	5.35	5.68	5.24	5.40	6.27	5.68	4.59	5.59	5.16	5.54	6.71	5.47	0.49
P^{5+}	0.11	0.14	0.16	0.18	0.15	0.17	0.15	0.16	0.13	0.15	0.14	0.13	0.16	0.14	0.15	0.18	0.150	0.015

* T.Fe = Total iron as Fe_2O_3

‡ S.D. = standard deviation uncorrected either for experimental error or for small sample effect. No sample excluded in calculating mean or S.D.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE VI.

Whole Rock Analyses. Granitic Gneisses. Group VI.

Weight Percentages.

	G221	G222A	G223	G231	G232	G233	G241	G242	G243	G244	G251	G252	G253	G254
SiO_2	71.3	72.2	69.2	68.9	69.7	68.7	69.9	70.3	69.6	70.3	69.9	68.8	71.5	69.8
TiO_2	0.44	0.56	0.65	0.74	0.66	0.64	0.69	0.62	0.73	0.52	0.69	0.72	0.61	0.62
Al_2O_3	14.5	13.1	13.9	13.7	14.1	14.6	13.8	13.7	13.6	14.1	14.2	13.5	13.6	13.9
Fe_2O_3	0.49	0.71	1.31	0.99	0.92	1.08	0.95	0.71	1.18	1.08	1.05	0.91	1.00	1.01
FeO	2.17	2.94	3.25	4.14	4.05	3.40	3.85	3.48	3.74	2.70	3.94	4.28	3.59	3.76
MnO	0.063	0.091	0.088	0.089	0.091	0.068	0.092	0.080	0.076	0.054	0.092	0.088	0.083	0.096
MgO	0.65	0.54	0.70	0.73	0.65	0.62	0.76	0.63	0.80	0.62	0.74	0.89	0.59	0.79
CaO	2.42	1.30	1.50	1.56	1.55	1.90	1.47	1.40	1.21	1.31	1.51	1.42	1.45	1.45
Na_2O	4.81	2.92	3.08	3.15	3.37	3.72	3.06	3.26	3.21	3.63	2.77	3.32	3.14	3.09
K_2O	1.57	4.96	4.98	4.45	4.50	4.20	4.17	4.64	4.48	5.01	4.33	3.60	4.24	4.80
P_2O_5	0.207	0.165	0.199	0.197	0.197	0.199	0.204	0.193	0.241	0.217	0.205	0.210	0.211	0.231
H_2O+	0.56	0.56	0.53	0.68	0.50	0.55	0.67	0.77	0.94	0.55	0.73	1.56	0.64	0.64
Total	99.19	100.11	99.39	99.32	100.29	99.69	99.51	99.81	99.83	100.07	100.09	99.34	100.59	100.18
T.Fe *	2.90	3.98	4.92	5.59	4.97	4.48	5.23	4.58	5.34	4.08	5.40	5.67	4.99	5.19
Rb (ppm)	87	177	176	164	186	176	175	157	169	165	176	138	151	180
Ba (ppm)	340	820	900	930	880	1050	790	970	1220	870	790	730	780	990
Sr (ppm)	430	200	135	195	120	275	160	150	315	195	155	220	160	155
Zr (ppm)	512	382	560	486	538	314	650	512	586	416	654	496	496	486

One-cation Percentages.

	G221	G222A	G223	G231	G232	G233	G241	G242	G243	G244	G251	G252	G253	G254	Arith. Mean.	S.D.‡
Si^{4+}	67.23	68.46	65.95	65.87	65.70	64.97	66.71	66.81	66.36	66.12	66.53	66.40	67.51	66.07	66.48	0.79
Ti^{4+}	0.31	0.40	0.46	0.53	0.47	0.45	0.50	0.45	0.53	0.36	0.50	0.52	0.43	0.44	0.45	0.06
Al^{3+}	16.15	14.64	15.58	15.44	15.70	16.23	15.47	15.33	15.27	15.59	15.88	15.35	15.12	15.48	15.52	0.37
Fe^{3+}	0.35	0.51	0.94	0.71	0.65	0.77	0.68	0.51	0.85	0.76	0.75	0.66	0.71	0.72	0.68	0.14
Fe^{2+}	1.71	2.33	2.59	3.31	3.19	2.69	3.07	2.76	2.98	2.12	3.13	3.45	2.84	2.97	2.80	0.46
Mn^{2+}	0.05	0.07	0.07	0.07	0.07	0.06	0.07	0.06	0.06	0.05	0.07	0.07	0.07	0.08	0.066	0.007
Mg^{2+}	0.91	0.80	1.00	1.04	0.91	0.87	1.08	0.89	1.13	0.87	1.05	1.28	0.83	1.11	0.98	0.13
Ca^{2+}	2.44	1.32	1.53	1.60	1.56	1.92	1.50	1.43	1.24	1.32	1.54	1.46	1.47	1.47	1.56	0.29
Na^{+}	8.79	5.36	5.69	5.83	6.16	6.81	5.66	6.00	5.93	6.61	5.11	6.20	5.75	5.67	6.11	0.86
K^{+}	1.89	5.99	6.05	5.43	5.41	5.06	5.08	5.62	5.44	6.01	5.25	4.43	5.11	5.79	5.18	1.01
P^{5+}	0.16	0.13	0.16	0.16	0.16	0.16	0.17	0.15	0.19	0.18	0.17	0.17	0.17	0.18	0.165	0.014

* T.Fe = Total iron as Fe_2O_3

‡ S.D. = standard deviation uncorrected either for experimental error or for small sample effect. No sample excluded in calculating mean or S.D.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE VII.

Whole Rock Analyses. Granitic Gneisses. Group VII.

Weight Percentages.

	G261	G262	G263	G271	G271A	G272	G273	G274	G281	G282	G283	G284	G291	G292	G293
SiO_2	69.4	70.1	70.3	69.1	69.8	68.8	71.2	73.3	71.1	70.2	71.9	70.3	71.4	71.2	69.7
TiO_2	0.69	0.63	0.60	0.79	0.73	0.80	0.57	0.45	0.59	0.66	0.56	0.71	0.59	0.55	0.58
Al_2O_3	14.0	13.5	13.6	13.6	13.1	13.5	13.1	13.9	13.5	13.7	14.0	13.6	12.9	13.2	13.9
Fe_2O_3	0.89	0.53	0.71	0.91	0.67	1.05	1.06	0.77	0.93	0.80	0.85	1.08	1.59	1.16	1.11
FeO	3.97	3.99	3.57	4.37	4.19	4.38	2.97	2.34	3.30	3.98	3.14	4.04	2.82	2.89	3.13
MnO	0.087	0.104	0.083	0.116	0.092	0.117	0.079	0.038	0.096	0.101	0.085	0.103	0.072	0.065	0.083
MgO	0.78	0.69	0.65	0.96	0.84	0.98	0.53	0.63	0.47	0.53	0.60	0.67	0.59	0.60	0.66
CaO	1.47	1.71	1.48	1.62	1.90	1.48	1.34	2.43	1.81	1.68	2.83	1.84	1.29	1.19	1.55
Na_2O	2.91	2.91	3.27	2.73	2.62	3.35	3.26	4.47	3.15	3.11	4.08	2.97	3.10	2.96	3.43
K_2O	4.29	4.62	4.52	4.41	4.70	4.75	4.81	1.44	4.58	4.63	1.26	4.07	4.41	4.77	4.57
P_2O_5	0.197	0.167	0.202	0.221	0.219	0.226	0.191	0.060	0.149	0.106	0.166	0.256	0.200	0.206	0.198
H_2O^+	0.71	0.88	0.43	0.61	0.58	0.75	0.47	0.40	0.43	0.61	0.41	0.66	0.61	0.66	0.59
Total	99.33	99.85	99.42	99.45	99.39	100.17	99.57	100.21	100.10	100.09	99.84	100.26	99.53	99.46	99.42
T.Fe *	5.30	4.96	4.68	5.77	5.33	5.92	4.36	3.37	4.60	5.22	4.34	5.57	4.72	4.37	4.59
Rb (ppm)	165	155	170	152	163	171	182	92	188	198	77	155	152	167	180
Ba (ppm)	890	910	920	1150	1160	940	820	200	810	870	360	880	940	950	840
Sr (ppm)	130	195	135	270	180	215	145	210	130	135	180	200	245	155	190
Zr (ppm)	552	656	568	578	586	568	452	408	616	714	370	626	450	476	450

One-cation Percentages.

	G261	G262	G263	G271	G271A	G272	G273	G274	G281	G282	G283	G284	G291	G292	G293	Group VII. only. Arith. Mean	Group VII. only. S.D.‡	All granitic gneisses. Arith Mean	All granitic gneisses. S.D.‡
Si^{4+}	66.41	66.85	66.84	66.18	66.88	65.16	67.62	68.73	67.22	66.60	68.01	66.80	68.21	67.98	66.20	67.05	0.80	68.12	1.66
Ti^{4+}	0.49	0.45	0.43	0.56	0.52	0.57	0.41	0.32	0.42	0.47	0.40	0.51	0.42	0.40	0.42	0.45	0.06	0.41	0.08
Al^{3+}	15.80	15.21	15.23	15.36	14.75	15.03	14.70	15.33	15.09	15.28	15.59	15.19	14.47	14.91	15.51	15.16	0.32	14.67	0.73
Fe^{3+}	0.64	0.38	0.51	0.66	0.48	0.75	0.76	0.54	0.66	0.57	0.61	0.77	1.14	0.83	0.79	0.67	0.17	0.70	0.26
Fe^{2+}	3.18	3.18	2.84	3.50	3.36	3.46	2.36	1.84	2.61	3.16	2.48	3.21	2.26	2.31	2.49	2.82	0.50	2.26	0.67
Mn^{2+}	0.07	0.09	0.07	0.09	0.07	0.09	0.07	0.03	0.08	0.08	0.07	0.09	0.06	0.05	0.07	0.072	0.016	0.059	0.015
Mg^{2+}	1.11	0.98	0.92	1.37	1.20	1.38	0.75	0.88	0.66	0.75	0.85	0.95	0.84	0.85	0.94	0.96	0.21	0.82	0.24
Ca^{2+}	1.51	1.75	1.51	1.66	1.95	1.50	1.36	2.44	1.84	1.71	2.87	1.87	1.32	1.22	1.57	1.74	0.42	1.54	0.34
Na^{+}	5.40	5.38	6.02	5.07	4.86	6.15	6.00	8.12	5.77	5.71	7.48	5.47	5.74	5.48	6.32	5.93	0.83	5.87	0.66
K^{+}	5.23	5.62	5.48	5.38	5.74	5.73	5.83	1.72	5.52	5.60	1.52	4.93	5.37	5.81	5.53	5.00	1.34	5.42	0.89
P^{5+}	0.16	0.14	0.16	0.18	0.18	0.18	0.15	0.05	0.12	0.09	0.13	0.21	0.16	0.17	0.16	0.139	0.052	0.130	0.042

* T.Fe = Total iron as Fe_2O_3

‡ S.D. = standard deviation uncorrected either for experimental error or for small sample effect. No samples excluded from mean or S.D.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE VIII.

Whole Rock Analyses. Pelitic schists and oligoclase gneisses.

Weight Percentages.

	P42	P53	P81	P96	P102	P125	P132	P142	P161	P166	P181	P192	P223	P233	P273	P291
SiO_2	65.7	66.0	67.3	59.0	60.8	59.3	59.4	65.9	73.9	59.0	69.0	57.9	66.6	64.8	61.1	66.2
TiO_2	0.90	0.79	0.75	1.34	1.10	1.22	1.15	0.86	0.80	1.07	0.71	1.01	0.75	0.88	0.96	0.86
Al_2O_3	15.3	17.3	15.8	17.1	17.5	18.0	17.2	16.5	12.6	18.6	14.8	18.0	15.5	16.0	18.0	15.6
Fe_2O_3	0.87	0.70	0.65	0.74	1.07	0.99	0.67	0.97	0.56	1.17	0.61	1.24	0.67	1.01	1.62	1.16
FeO	4.99	2.98	4.06	6.74	5.03	6.15	6.38	3.90	3.57	6.21	3.73	5.97	4.04	4.19	4.81	4.04
MnO	0.086	0.071	0.094	0.128	0.109	0.128	0.122	0.103	0.104	0.137	0.082	0.140	0.135	0.092	0.136	0.091
MgO	1.82	1.43	1.68	2.72	2.29	2.54	2.21	1.59	0.81	2.16	1.45	2.18	1.64	1.89	2.48	1.95
CaO	2.31	3.66	2.32	2.82	3.61	3.17	2.36	1.46	2.17	2.32	2.17	1.81	2.75	2.89	2.02	3.04
Na_2O	4.07	4.17	3.96	3.46	4.08	3.83	4.28	2.92	3.15	3.22	3.70	4.49	4.38	3.95	2.94	3.26
K_2O	2.54	1.87	2.59	3.46	2.62	2.84	4.86	3.88	2.31	3.31	2.51	5.84	2.26	2.64	3.76	2.70
P_2O_5	0.157	0.053	0.190	0.211	0.210	0.230	0.189	0.138	0.127	0.238	0.147	0.318	0.226	0.322	0.283	0.181
H_2O+	0.97	0.86	1.00	1.37	1.17	1.60	0.79	1.60	0.48	1.62	0.89	1.06	0.83	1.39	1.82	0.90
Total	99.77	99.87	100.38	99.04	99.63	100.04	99.56	99.84	100.60	99.09	99.86	99.96	99.81	100.03	99.85	99.95
T.Fe *	6.42	4.01	5.16	8.23	6.66	7.85	7.76	5.30	4.33	8.08	4.76	7.21	5.16	5.67	7.06	5.65
Rb (ppm)	79	51	94	133	102	121	309	107	91	156	131	221	122	127	160	108
Ba (ppm)	840	560	720	880	810	800	780	1160	610	890	460	1370	550	630	930	590
Sr (ppm)	275	495	420	280	545	360	240	275	270	300	400	325	355	350	310	405
Zr (ppm)	296	514	314	378	316	334	250	302	470	236	346	180	262	256	302	282

One-cation Percentages.

	P42	P53	P81	P96	P102	P125	P132	P142	P161	P166	P181	P192	P223	P233	P273	P291	Arith. Mean.	S.D.‡
Si^{4+}	62.12	61.93	63.16	56.28	57.32	56.27	55.54	62.89	70.03	56.60	65.31	53.82	62.61	61.20	58.22	62.68	59.73	3.41
Ti^{4+}	0.64	0.56	0.53	0.96	0.78	0.87	0.81	0.62	0.57	0.77	0.50	0.70	0.53	0.62	0.69	0.61	0.68	0.13
Al^{3+}	17.06	19.09	17.46	19.23	19.48	20.13	18.94	18.60	14.07	21.00	16.55	19.76	17.17	17.83	20.17	17.39	18.66	1.30
Fe^{3+}	0.62	0.50	0.46	0.53	0.76	0.71	0.47	0.70	0.40	0.85	0.43	0.87	0.47	0.72	1.16	0.82	0.67	0.20
Fe^{2+}	3.94	2.34	3.18	5.37	3.96	4.88	4.99	3.11	2.83	4.97	2.95	4.64	3.17	3.31	3.84	3.20	3.86	0.89
Mn^{2+}	0.07	0.06	0.07	0.10	0.08	0.10	0.10	0.09	0.09	0.11	0.07	0.11	0.11	0.07	0.11	0.07	0.088	0.018
Mg^{2+}	2.56	2.00	2.35	3.87	3.21	3.59	3.08	2.26	1.14	3.09	2.05	3.02	2.30	2.66	3.52	2.75	2.82	0.56
Ca^{2+}	2.34	3.68	2.33	2.88	3.65	3.22	2.37	1.49	2.20	2.38	2.20	1.80	2.77	2.92	2.06	3.08	2.61	0.62
Na^{+}	7.45	7.58	7.20	6.39	7.45	7.04	7.76	5.40	5.78	5.98	6.79	8.09	7.97	7.23	5.43	5.98	6.92	0.86
K^{+}	3.06	2.24	3.10	4.20	3.15	3.00	5.80	4.72	2.79	4.05	3.03	6.93	2.71	3.18	4.58	3.26	3.80	1.22
P^{5+}	0.13	0.04	0.15	0.17	0.17	0.18	0.15	0.11	0.10	0.20	0.12	0.25	0.18	0.26	0.23.	0.15	0.166	0.055

* T.Fe = Total iron as Fe_2O_3

‡ S.D. = standard deviation uncorrected either for experimental error or for small sample effect. Sample P161 excluded from all means and S.D.s

Samples dried at 110°C for 12 hours prior to analysis.

TABLE IX.

Whole Rock Analyses. Psammites from the Glen Garvan Psammitic Group.

Weight Percentages.

	S42	S52	S71	S77	S96	S101	S116	S121	S141	S142	S151	S162	S191	S213	S252	S261
SiO_2	67.3	76.0	76.4	82.2	83.4	86.2	76.5	76.4	72.6	90.4	81.1	90.7	88.3	80.3	75.8	83.8
TiO_2	0.97	0.27	0.41	0.37	0.33	0.17	0.29	0.59	0.75	0.18	0.39	0.13	0.30	0.10	0.46	0.33
Al_2O_3	15.5	12.0	12.3	8.43	8.48	7.27	12.0	11.9	13.4	5.10	10.3	4.80	6.08	11.4	12.7	8.47
Fe_2O_3	0.69	0.36	0.67	0.25	0.27	0.31	0.48	0.55	0.41	0.16	0.22	0.12	0.24	0.22	0.49	0.16
FeO	3.24	1.27	1.30	1.14	0.78	0.30	1.81	1.70	2.34	0.43	1.20	0.62	0.54	0.42	1.61	0.66
MnO	0.051	0.027	0.059	0.030	0.026	0.021	0.052	0.044	0.065	0.013	0.035	0.025	0.016	0.016	0.104	0.022
MgO	1.06	0.46	0.65	0.23	0.09	0.11	0.30	0.59	0.81	0.09	0.16	0.12	0.11	0.21	0.89	0.15
CaO	1.24	0.87	2.24	0.60	0.48	0.45	2.11	2.50	1.92	0.24	0.81	0.15	0.23	0.73	3.23	0.36
Na_2O	3.33	2.36	3.35	1.74	1.49	1.68	3.52	3.18	3.51	0.82	2.21	0.54	1.53	2.57	2.83	1.32
K_2O	6.01	5.56	2.15	4.44	4.35	3.23	1.54	1.61	3.24	2.50	2.45	2.33	2.66	4.37	1.05	4.46
P_2O_5	0.090	0.026	0.031	0.027	0.053	0.017	0.032	0.059	0.054	0.025	0.044	0.016	0.009	0.015	0.036	0.045
H_2O+	0.71	0.47	0.47	0.20	0.27	0.14	0.46	0.20	0.50	0.07	0.49	0.12	0.16	0.09	0.39	0.11
Total	100.21	99.69	100.04	99.66	100.04	99.86	99.13	99.36	99.58	100.05	99.45	99.65	100.14	100.38	99.59	99.11
T.Fe *	4.29	1.73	2.11	1.52	1.14	0.64	2.49	2.44	3.01	0.64	1.55	0.81	0.84	0.69	2.28	0.89
Rb (ppm)	149	137	58	128	113	97	65	76	115	67	80	78	55	105	57	113
Ba (ppm)	1690	1200	1310	860	1050	770	300	600	950	680	740	640	920	1340	660	1380
Sr (ppm)	270	205	550	180	165	160	350	335	360	100	260	80	160	270	560	260
Zr (ppm)	558	172	292	316	322	206	148	402	614	200	384	80	442	76	270	500

One-cation Percentages.

	S42	S52	S71	S77	S96	S101	S116	S121	S141	S142	S151	S162	S191	S213	S252	S261	Arith. Mean	S.D.‡
Si^{4+}	63.12	72.38	72.47	79.32	80.46	83.34	73.34	73.07	68.85	88.42	78.53	89.39	85.63	75.76	72.48	80.89	77.34	7.00
Ti^{4+}	0.68	0.19	0.29	0.27	0.24	0.12	0.21	0.42	0.54	0.14	0.28	0.09	0.22	0.07	0.33	0.24	0.27	0.16
Al^{3+}	17.10	13.47	13.71	9.59	9.64	8.28	13.57	13.42	14.91	5.88	11.75	5.58	6.95	12.66	14.33	9.63	11.28	2.33
Fe^{3+}	0.48	0.26	0.48	0.18	0.20	0.23	0.35	0.40	0.29	0.12	0.16	0.09	0.17	0.16	0.35	0.12	0.25	0.12
Fe^{2+}	2.54	1.01	1.03	0.92	0.63	0.24	1.45	1.36	1.86	0.35	0.97	0.51	0.44	0.33	1.29	0.53	0.97	0.60
Mn^{2+}	0.04	0.02	0.05	0.02	0.02	0.02	0.04	0.03	0.05	0.01	0.03	0.02	0.01	0.01	0.09	0.02	0.030	0.020
Mg^{2+}	1.48	0.65	0.92	0.33	0.13	0.16	0.43	0.84	1.14	0.13	0.23	0.18	0.16	0.29	1.27	0.21	0.53	0.44
Ca^{2+}	1.24	0.89	2.27	0.62	0.50	0.46	2.16	2.56	1.95	0.25	0.84	0.16	0.24	0.74	3.31	0.37	1.16	0.95
Na^{+}	6.05	4.35	6.16	3.25	2.79	3.15	6.54	5.89	6.45	1.56	4.14	1.03	2.88	4.70	5.24	2.47	4.17	1.73
K^{+}	7.18	6.75	2.60	5.47	5.35	3.98	1.88	1.96	3.92	3.12	3.02	2.93	3.29	5.26	1.28	5.49	3.97	1.71
P^{5+}	0.07	0.02	0.02	0.02	0.04	0.01	0.03	0.05	0.05	0.02	0.03	0.01	0.01	0.01	0.03	0.03	0.028	0.017

*T.Fe = Total iron as Fe_2O_3

‡ S.D. = standard deviation uncorrected for experimental error or small sample effect.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE X.

Whole Rock Analyses. Psammites and striped schists from the Beinn an Tuim Striped Group and Druim na Sailleach Pelitic Group.

Weight Percentages.

	Q187	K191	SK211	SK221	SK222	SK241	SK242	K271	K281	SK291
SiO_2	90.7	85.9	80.8	88.3	84.1	88.7	91.9	81.3	77.4	81.1
TiO_2	0.24	0.28	0.34	0.22	0.36	0.23	0.20	0.40	0.54	0.23
Al_2O_3	5.02	7.87	10.7	6.12	8.81	6.04	4.76	9.90	11.3	10.5
Fe_2O_3	0.28	0.42	0.32	0.29	0.42	0.34	0.23	0.35	0.77	0.47
FeO	0.51	0.55	0.97	0.59	0.92	0.49	0.56	1.25	1.97	0.89
MnO	0.015	0.027	0.042	0.018	0.033	0.014	0.021	0.038	0.054	0.024
MgO	0.12	0.21	0.33	0.22	0.37	0.13	0.17	0.50	0.79	0.46
CaO	0.39	0.46	0.72	0.65	1.53	0.24	0.41	1.66	1.68	1.39
Na_2O	1.32	1.50	2.51	1.74	3.16	1.14	0.88	2.80	3.17	3.46
K_2O	1.24	2.49	3.69	1.79	0.66	2.98	1.00	1.61	1.60	1.19
P_2O_5	0.022	0.035	0.037	0.029	0.133	0.020	0.019	0.081	0.108	0.029
H_2O+	0.13	0.21	0.17	0.04	0.16	0.15	0.20	0.16	0.46	0.34
Total	100.03	99.98	100.60	99.95	100.64	100.51	100.31	100.08	99.84	100.04
T.Fe *	0.85	1.03	1.40	0.95	1.44	0.88	0.85	1.74	2.96	1.46
Rb (ppm)	25	77	82	41	27	69	25	42	58	31
Ba (ppm)	370	590	1040	590	220	650	200	580	460	440
Sr (ppm)	100	125	250	230	265	90	85	285	300	590
Zr (ppm)	386	182	324	286	418	206	210	304	356	208

One-cation Percentages.

	Q187	K191	SK211	SK221	SK222	SK241	SK242	K271	K281	SK291	Arith. Mean	S.D.‡
Si^{4+}	88.74	83.32	76.56	85.64	79.99	85.99	90.09	77.68	73.93	77.10	81.90	5.32
Ti^{4+}	0.18	0.20	0.24	0.16	0.26	0.17	0.15	0.29	0.39	0.17	0.22	0.07
Al^{3+}	5.79	8.99	11.90	7.00	9.88	6.90	5.50	11.14	12.75	11.76	9.16	2.57
Fe^{3+}	0.21	0.31	0.23	0.21	0.30	0.25	0.17	0.25	0.55	0.34	0.28	0.10
Fe^{2+}	0.42	0.45	0.77	0.48	0.73	0.40	0.46	1.00	1.57	0.71	0.70	0.34
Mn^{2+}	0.01	0.02	0.03	0.02	0.03	0.01	0.02	0.03	0.05	0.02	0.024	0.011
Mg^{2+}	0.18	0.30	0.47	0.32	0.53	0.18	0.25	0.71	1.12	0.65	0.47	0.28
Ca^{2+}	0.41	0.48	0.73	0.68	1.56	0.25	0.43	1.70	1.72	1.42	0.94	0.56
Na^{+}	2.50	2.81	4.61	3.27	5.82	2.14	1.67	5.18	5.87	6.37	4.02	1.65
K^{+}	1.54	3.08	4.45	2.21	0.80	3.68	1.25	1.96	1.95	1.44	2.24	1.10
P^{5+}	0.02	0.03	0.03	0.02	0.11	0.02	0.02	0.06	0.09	0.02	0.042	0.032

* T.Fe = Total iron as Fe_2O_3

‡S.D. = standard deviation uncorrected for experimental error or for small sample effect.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE XI

Whole Rock Analyses. Calcareous psammites

Weight Percentages.

	L116	S172
SiO_2	76.9	65.1
TiO_2	0.10	0.77
Al_2O_3	13.1	15.6
Fe_2O_3	0.21	1.58
FeO	1.41	3.56
MnO	0.164	0.148
MgO	0.50	1.84
CaO	3.89	5.52
Na_2O	2.50	2.64
K_2O	1.14	1.65
P_2O_5	0.052	0.275
H_2O	0.58	0.65
Total	100.58	99.31
* T.Fe	1.78	5.54
Rb (ppm)	43	81
Ba (ppm)	50	640
Sr (ppm)	280	445
Zr (ppm)	72	338

One-cation Percentages.

	L116	S172
Si^{4+}	73.11	62.32
Ti^{4+}	0.07	0.56
Al^{3+}	14.72	17.59
Fe^{3+}	0.15	1.14
Fe^{2+}	1.12	2.85
Mn^{2+}	0.13	0.12
Mg^{2+}	0.71	2.62
Ca^{2+}	3.96	5.66
Na^{+}	4.61	4.90
K^{+}	1.38	2.01
P^{5+}	0.04	0.22

*T.Fe = Total iron as Fe_2O_3

Samples dried at 110°C for 12 hours prior to analysis.

TABLE XII.

Whole Rock Analyses. Pegmatites, aplites, etc.

Weight Percentages.

	GS166	G222	G262A	G274A
SiO_2	69,6	72.6	70.6	73.2
TiO_2	0.46	0.38	0.50	0.30
Al_2O_3	15.6	13.8	13.8	14.4
Fe_2O_3	0.35	0.69	0.47	0.51
FeO	1.66	1.74	2.94	1.04
MnO	0.038	0.057	0.111	0.050
MgO	0.66	0.28	0.51	0.39
CaO	1.89	1.33	1.77	1.59
Na_2O	3.49	3.25	3.69	3.42
K_2O	5.32	5.13	4.12	4.07
P_2O_5	0.096	0.101	0.125	0.033
H_2O	0.52	0.45	0.54	0.37
Total	99.73	99.83	99.25	99.47
*T.Fe	2.19	2.62	3.74	1.67
Rb (ppm)	127	177	115	118
Ba (ppm)	2440	860	940	1030
Sr (ppm)	580	165	315	350
Zr (ppm)	138	276	498	78

One-cation Percentages.

	GS166	G222	G262A	G274A
Si^{4+}	65.26	68.54	67.03	69.07
Ti^{4+}	0.33	0.27	0.35	0.22
Al^{3+}	17.24	15.35	15.46	16.17
Fe^{3+}	0.25	0.49	0.34	0.36
Fe^{2+}	1.30	1.37	2.33	0.82
Mn^{2+}	0.03	0.05	0.09	0.04
Mg^{2+}	0.92	0.40	0.72	0.55
Ca^{2+}	1.90	1.34	1.80	1.61
Na^{+}	6.34	5.94	6.78	6.25
K^{+}	6.36	6.17	4.99	4.89
P^{5+}	0.08	0.08	0.10	0.03

*T.Fe = Total iron as Fe_2O_3

Samples dried at 110°C for 12 hours prior to analysis.

TABLE XIII.

Eskola molecular norms using C.I.P.W. normative minerals.

(a) Granitic gneisses.

	G11	G12	G13	G21	G22	G23	G31	G32	G33	G41	G42	G43	G51	G52	
Q	28.69	26.11	29.19	28.67	25.71	29.38	29.43	28.59	27.68	31.96	26.41	29.50	29.46	31.40	Q
Or	24.90	28.55	12.75	29.30	27.85	24.90	23.65	30.85	31.05	27.45	30.30	23.40	34.05	25.85	Or
Ab	33.80	29.55	39.30	28.75	29.05	30.05	31.25	29.55	26.50	24.30	29.80	31.65	26.95	27.15	Ab
An	5.25	7.45	10.25	5.75	9.70	8.45	6.65	5.20	7.85	6.20	6.95	6.75	4.35	6.20	An
C	0.47	0.66	1.05	0.90	-	-	1.11	0.37	-	1.71	0.23	1.82	0.92	1.39	C
Di	-	-	-	-	0.12	0.92	-	-	0.44	-	-	-	-	-	Di
Hy	4.82	5.28	5.34	4.62	4.78	3.24	6.04	3.78	3.24	6.76	4.24	4.40	2.78	5.64	Hy
Mt	0.92	1.16	0.96	0.84	1.40	1.85	0.63	0.60	2.01	0.35	1.01	1.26	0.68	1.07	Mt
Il	0.74	0.82	0.80	0.76	1.04	0.92	0.88	0.64	0.90	0.88	0.78	0.92	0.58	0.94	Il
Ap	0.42	0.42	0.35	0.40	0.35	0.27	0.35	0.42	0.32	0.38	0.30	0.30	0.25	0.37	Ap

	G53	G54	G55	G56	G61	G62	G63	G64	G65	G71	G72	G73	G74	G81	
Q	40.12	30.47	26.86	33.08	28.95	34.03	28.81	27.13	23.61	33.39	34.42	32.00	29.68	24.92	Q
Or	22.10	28.00	29.80	23.55	28.95	27.25	28.95	30.00	27.75	21.80	25.20	29.15	28.45	34.65	Or
Ab	23.15	27.00	32.65	29.85	26.40	25.80	29.30	28.65	32.15	31.00	25.95	26.55	28.55	28.20	Ab
An	7.70	6.20	6.10	6.30	6.40	6.05	5.95	6.75	7.00	6.70	5.55	5.05	6.35	5.00	An
C	0.04	1.03	0.03	0.25	0.74	0.05	-	0.16	1.12	0.77	1.49	0.74	0.32	-	C
Di	-	-	-	-	-	-	0.08	-	-	-	-	-	-	0.12	Di
Hy	3.56	5.00	1.74	4.20	6.04	4.62	4.94	5.24	6.10	4.50	5.04	4.48	4.68	5.20	Hy
Mt	2.03	1.10	1.74	1.46	1.20	1.01	0.84	0.86	0.99	0.87	1.07	0.90	0.83	0.68	Mt
Il	0.90	0.84	0.68	0.96	0.94	0.86	0.86	0.82	0.90	0.76	0.96	0.80	0.80	0.86	Il
Ap	0.40	0.37	0.40	0.35	0.36	0.32	0.27	0.37	0.40	0.23	0.32	0.35	0.35	0.37	Ap

	G82	G83	G91	G92	G93	G96	G97	G98	G101	G102	G103	G105	G111	G112	
Q	27.21	34.42	34.32	27.81	25.09	37.04	29.49	27.21	34.69	33.46	29.22	25.90	29.97	35.34	Q
Or	27.85	22.35	30.20	27.70	30.85	30.70	29.65	29.85	33.05	25.15	26.15	29.50	31.15	28.45	Or
Ab	30.40	25.20	27.05	29.25	29.35	22.60	28.25	30.75	25.25	27.35	35.35	31.95	27.80	27.85	Ab
An	5.55	6.15	3.90	6.30	6.40	4.65	5.90	6.00	3.30	5.60	6.30	6.15	5.40	5.50	An
C	1.18	2.54	0.79	1.10	0.51	1.20	1.02	-	-	1.73	1.09	0.01	0.28	-	C
Di	-	-	-	-	-	-	-	0.40	1.08	-	-	-	-	0.20	Di
Hy	5.68	6.88	2.56	5.80	6.00	2.74	4.24	4.00	1.68	4.66	1.14	3.64	3.54	0.92	Hy
Mt	0.78	1.05	0.54	0.84	0.63	0.48	0.48	0.69	0.38	0.86	0.30	1.73	0.89	1.07	Mt
Il	0.92	1.06	0.52	0.84	0.82	0.48	0.72	0.86	0.46	0.88	0.30	0.84	0.76	0.52	Il
Ap	0.45	0.35	0.12	0.35	0.37	0.12	0.25	0.25	0.10	0.29	0.11	0.27	0.20	0.15	Ap

(Continued on next page)

TABLE XIII. (Continued)

Eskola molecular norms using C.I.P.W. normative minerals.

(a) Granitic gneisses, (Continued)

	GP112	G113	G116	G117	G121	G122	G123	G131	G132	G133	G141	G142	G151	G152	
Q	28.98	31.36	28.15	29.85	29.43	34.08	34.31	34.47	32.77	31.45	36.33	28.36	27.74	30.70	Q
Or	32.55	23.50	30.15	28.00	28.15	30.60	29.30	31.70	34.10	21.80	30.60	25.55	30.85	22.70	Or
Ab	25.10	27.10	30.40	31.15	30.60	27.60	27.70	25.55	25.40	26.10	23.75	26.65	32.00	30.10	Ab
An	6.65	9.65	6.48	5.80	6.95	3.60	6.00	4.50	4.60	10.90	4.80	9.20	4.50	9.95	An
C	0.50	0.46	-	-	0.12	-	0.07	-	-	0.42	-	0.95	0.89	0.25	C
Di	-	-	0.32	1.24	-	1.68	-	0.36	0.36	-	0.04	-	-	-	Di
Hy	3.08	4.94	2.08	1.70	2.78	0.18	1.48	2.16	0.94	7.04	2.10	6.42	2.62	3.42	Hy
Mt	1.86	1.73	1.49	1.31	1.11	1.46	0.59	0.62	1.17	0.93	1.49	1.31	0.68	1.71	Mt
Il	0.96	0.98	0.70	0.70	0.64	0.60	0.40	0.54	0.50	1.06	0.72	1.22	0.56	0.86	Il
Ap	0.30	0.27	0.23	0.25	0.23	0.20	0.12	0.12	0.15	0.30	0.15	0.32	0.15	0.32	Ap

	G161	G162	G163	G171	G172	G173	G181	G182	G183	G184	G185	G186	G187	G191	
Q	33.46	27.58	31.33	28.80	25.91	30.78	29.11	28.76	29.92	26.91	30.98	28.61	29.47	26.72	Q
Or	30.85	25.85	28.50	22.85	26.20	24.60	27.50	28.05	26.60	24.10	25.35	26.75	28.40	26.20	Or
Ab	28.05	30.80	28.45	30.55	32.10	30.15	30.20	31.50	28.95	32.30	28.05	30.20	30.10	31.75	Ab
An	4.00	6.90	6.35	7.95	7.75	7.85	5.85	5.10	5.70	8.05	7.40	5.15	5.23	7.55	An
C	-	-	0.04	0.96	0.28	-	0.72	0.65	1.02	0.26	0.50	1.04	-	0.20	C
Di	1.12	2.40	-	-	-	0.24	-	-	-	-	-	-	0.20	-	Di
Hy	1.12	2.94	3.90	6.66	5.72	3.84	4.86	3.60	5.62	5.56	5.00	6.22	4.54	5.22	Hy
Mt	0.77	2.12	0.54	0.89	0.80	1.32	0.86	1.23	0.39	1.43	1.32	0.69	0.87	1.10	Mt
Il	0.46	1.08	0.70	0.98	0.82	0.90	0.64	0.76	0.86	0.96	0.98	0.92	0.76	0.88	Il
Ap	0.17	0.35	0.20	0.35	0.40	0.32	0.27	0.35	0.40	0.45	0.42	0.42	0.42	0.40	Ap

	G192	G193	G201	G202	G203	G211	G212	G213	G221	G222A	G223	G231	G232	G233	
Q	28.63	24.36	29.76	27.38	26.51	30.96	29.39	25.54	28.61	29.61	25.42	25.84	24.98	23.22	Q
Or	27.00	31.35	28.40	22.95	27.95	25.80	27.70	33.55	9.45	29.95	30.25	27.15	27.05	25.30	Or
Ab	31.20	30.15	27.75	33.95	29.80	33.40	30.10	27.35	43.95	26.80	28.45	29.15	30.80	34.05	Ab
An	6.10	6.90	6.10	7.20	7.00	4.80	5.50	6.15	11.00	5.65	6.45	6.80	6.60	8.40	An
C	0.44	0.04	0.85	0.97	0.62	0.27	0.88	0.32	1.07	1.03	1.26	1.46	1.49	1.00	C
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Hy	4.78	4.68	5.04	5.08	5.40	1.62	3.64	4.76	4.36	5.08	5.46	7.06	6.74	5.56	Hy
Mt	0.75	1.31	0.93	1.23	1.41	2.07	1.53	1.07	0.53	0.77	1.41	1.07	0.98	1.16	Mt
Il	0.78	0.84	0.84	0.92	0.94	0.74	0.84	0.80	0.62	0.80	0.92	1.06	0.94	0.90	Il
Ap	0.32	0.37	0.35	0.32	0.40	0.35	0.42	0.45	0.40	0.32	0.40	0.40	0.40	0.40	Ap

(Continued on next page)

TABLE XIII. (Continued)

Eskola molecular norms using C.I.P.W. normative minerals.

(a) Granitic gneisses, (Continued)

	G241	G242	G243	G244	G251	G252	G253	G254	G261	G262	G263	G271	G271A	G272	
Q	28.63	26.65	27.14	23.86	29.52	28.16	29.56	25.92	28.43	27.16	26.66	27.98	27.85	23.08	Q
Or	25.40	28.10	27.20	30.05	26.25	22.15	25.55	28.95	26.15	28.10	27.40	26.90	28.70	28.65	Or
Ab	28.30	30.00	29.65	33.05	25.55	31.00	28.75	28.35	27.00	26.90	30.10	25.35	24.30	30.75	Ab
An	6.20	5.75	4.75	5.25	6.40	6.00	6.05	6.00	6.35	7.70	6.35	6.95	8.40	6.15	An
C	2.25	1.41	2.00	0.87	2.96	2.32	1.84	1.62	2.63	1.13	1.19	2.13	0.79	0.69	C
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Hy	6.76	6.00	6.42	4.60	6.74	7.90	5.90	6.72	7.10	7.22	6.28	8.14	7.74	7.96	Hy
Mt	1.02	0.77	1.28	1.14	1.13	0.99	1.07	1.08	0.96	0.57	0.77	0.99	0.72	0.13	Mt
Il	1.00	0.90	1.06	0.72	1.00	1.04	0.86	0.88	0.98	0.90	0.86	1.12	1.04	1.14	Il
Ap	0.43	0.43	0.48	0.45	0.43	0.43	0.43	0.45	0.40	0.35	0.40	0.45	0.45	0.45	Ap

	G273	G274	G281	G282	G283	G284	G291	G292	G293	(b) Pegmatites, aplites, etc. GS166	G222	G262A	G274A	
Q	27.58	32.33	27.43	26.34	32.97	29.15	30.55	29.80	25.31	21.81	28.47	25.80	31.54	Q
Or	29.15	8.60	27.60	28.00	7.60	24.65	26.85	29.05	27.65	31.80	30.85	24.95	24.45	Or
Ab	30.00	40.60	28.85	28.55	37.40	27.35	28.70	27.40	31.60	31.70	29.70	33.90	31.25	Ab
An	5.40	11.80	8.30	7.85	13.35	7.75	5.40	4.80	6.65	8.90	6.10	8.25	7.75	An
C	0.71	0.77	0.48	0.83	1.25	1.69	1.20	1.70	1.00	0.98	0.80	0.39	1.93	C
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Hy	4.78	4.32	5.20	6.38	5.38	6.70	4.34	4.78	5.36	3.58	2.60	5.24	2.02	Hy
Mt	1.14	0.81	0.99	0.86	0.92	1.16	1.71	1.25	1.19	0.38	0.74	0.51	0.54	Mt
Il	0.82	0.64	0.84	0.94	0.80	1.02	0.84	0.80	0.84	0.66	0.54	0.70	0.44	Il
Ap	0.43	0.13	0.30	0.23	0.33	0.53	0.40	0.43	0.40	0.20	0.20	0.25	0.08	Ap

	(c) Psammites and striped schists from the Beinn an Tuim striped group and the Druim na Saille pelitic group. Q187	K191	SK211	SK221	SK222	SK241	SK242	K271	K281	SK291	(d) Psammites from the Glen Garvan Psammitic group. S42	S52	S71	
Q	75.54	64.38	47.11	67.35	56.45	67.80	80.04	51.72	45.24	49.85	18.03	36.00	40.24	Q
Or	7.70	15.40	22.25	11.05	4.00	18.40	6.25	9.80	9.75	7.20	35.90	33.75	13.00	Or
Ab	12.50	14.05	23.05	16.35	29.10	10.70	8.35	25.90	29.35	31.85	30.25	21.75	30.80	Ab
An	1.90	2.15	3.40	3.25	7.00	1.10	2.00	8.05	7.90	6.95	5.70	4.30	11.20	An
C	0.99	2.24	1.48	0.22	0.46	0.64	1.78	0.78	1.77	1.17	1.59	0.65	0.47	C
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Hy	0.64	0.82	1.82	1.08	1.76	0.58	0.98	2.64	4.14	2.08	6.28	2.72	2.94	Hy
Mt	0.32	0.47	0.35	0.32	0.45	0.38	0.26	0.38	0.83	0.51	0.72	0.39	0.72	Mt
Il	0.36	0.40	0.48	0.32	0.52	0.34	0.30	0.58	0.78	0.34	1.36	0.38	0.58	Il
Ap	0.05	0.08	0.08	0.05	0.27	0.05	0.05	0.15	0.23	0.05	0.17	0.05	0.05	Ap

(Continued on next page)

TABLE XIII. (Continued)

Eskola molecular norms using C.I.P.W. normative minerals.

(d) Psammites from the Glen Garvan Psammitic group (Continued)

	S77	S96	S101	S116	S121	S141	S142	S151	S162	S191	S213	S252	S261	
Q	51.23	54.72	60.87	42.31	42.91	31.62	73.65	54.58	76.64	66.38	43.96	44.26	55.95	Q
Or	27.35	26.75	19.90	9.40	9.80	19.60	15.60	15.10	14.65	16.45	26.30	6.40	27.45	Or
Ab	16.25	13.95	15.75	32.70	29.45	32.25	7.80	20.70	5.15	14.40	23.50	26.20	12.35	Ab
An	2.20	2.20	2.25	10.60	12.45	9.40	1.10	4.00	0.75	1.10	3.60	16.30	1.50	An
C	-	0.62	0.25	0.91	0.59	0.78	0.76	2.99	1.32	0.34	1.26	1.29	1.07	C
Di	0.60	-	-	-	-	-	-	-	-	-	-	-	-	Di
Hy	1.52	0.88	0.36	3.06	3.26	4.72	0.58	1.74	1.14	0.60	0.96	4.30	0.92	Hy
Mt	0.27	0.30	0.35	0.53	0.60	0.44	0.18	0.24	0.14	0.26	0.24	0.53	0.18	Mt
Il	0.54	0.48	0.24	0.42	0.84	1.08	0.28	0.56	0.18	0.44	0.14	0.66	0.48	Il
Ap	0.05	0.10	0.02	0.07	0.12	0.12	0.05	0.07	0.02	0.03	0.03	0.08	0.10	Ap

(e) Pelitic schists and oligoclase gneisses.

	P42	P53	P81	P96	P102	P125	P132	P142	P161	
Q	20.69	21.64	23.22	11.14	12.63	12.91	3.44	25.36	36.93	Q
Or	15.30	11.20	15.50	21.00	15.75	15.00	29.00	23.60	13.95	Or
Ab	37.25	37.90	36.00	31.95	37.25	35.20	38.80	27.00	28.90	Ab
An	10.70	18.10	10.50	13.15	17.00	14.75	10.75	6.70	10.25	An
C	2.27	2.03	2.96	3.38	2.08	4.19	1.08	5.80	1.40	C
Di	-	-	-	-	-	-	-	-	-	Di
Hy	11.24	7.18	9.68	16.22	12.18	14.68	14.24	8.98	6.58	Hy
Mt	0.93	0.75	0.69	0.80	1.14	1.07	0.71	1.05	0.60	Mt
Il	1.28	1.12	1.06	1.92	1.56	1.74	1.62	1.24	1.14	Il
Ap	0.33	0.10	0.38	0.42	0.42	0.45	0.37	0.27	0.25	Ap

	P166	P181	P192	P223	P233	P273	P291	(f) Calcareous psammites. L116	S172	
Q	15.38	27.46	-	20.76	19.85	18.79	24.34	45.53	26.47	Q
Or	20.25	15.15	34.65	13.55	15.90	22.90	16.30	6.90	10.05	Or
Ab	29.90	33.95	40.45	39.85	36.15	27.15	29.90	23.05	24.50	Ab
An	10.40	10.10	7.15	12.50	12.65	8.00	14.05	19.50	26.65	An
C	6.81	2.69	1.88	1.49	2.36	6.96	2.53	0.93	0.02	C
Di	-	-	-	-	-	-	-	-	-	Di
Hy	13.94	8.70	10.34	9.62	10.12	12.40	10.00	3.62	8.92	Hy
Ol	-	-	2.19	-	-	-	-	-	-	Ol
Mt	1.28	0.65	1.31	0.71	1.08	1.74	1.23	0.23	1.71	Mt
Il	1.54	1.00	1.40	1.06	1.24	1.38	1.22	0.14	1.12	Il
Ap	0.50	0.30	0.62	0.45	0.65	0.69	0.42	0.10	0.55	Ap

TABLE XIV.

Mineral formulae and assemblages used in calculating Eskola molecular norms with high-grade metamorphic minerals.

Minerals used	Formula.
Quartz	SiO_2.
Potash Felspar	$Na\,AlSi_3O_8 \cdot\ 4KAlSi_3O_8$ $(= Or_{80}Ab_{20})$
Plagioclase	$x\,Ca\,Al_2Si_2O_8 \cdot\ (100-x)\,Na\,AlSi_3O_8$.
Muscovite	$KAl_3Si_3O_{10}(OH)_2$ *.
Garnet	$(Fe^{2+}, Mg)_3Al_2Si_3O_{12}$.
Biotite	$K_{1.85}((Fe^{2+},Mg)_{4.35}Fe^{3+}_{0.3}Ti_{0.45}Al_{0.5})\,Al_{2.75}Si_{5.25}O_{20}(OH)_4$. *
Hornblende	$Ca_{2.0}Na_{0.3}((Fe^{2+},Mg)_{3.5}\,Fe^{3+}_{0.8}Ti_{0.2}Al_{0.5})\,Al_2Si_6O_{22}(OH)_2$. *
Diopside	$Ca\,(Fe^{2+}, Mg)\,Si_2O_6$.
Magnetite	$Fe^{2+}\,Fe^{3+}_2O_4$.
Ilmenite	$Fe^{2+}\,Ti\,O_3$.
Apatite	$Ca_3P_2O_8$.

* The amount of OH required for the formation of these minerals in the correct quantities was assumed to be present, and the molecular percentages of these minerals were calculated on a water-free basis.

Permitted mineral assemblages.

	New Scheme	C. I. P. W. scheme	
Type C	Q + Plag + Musc + Gt + Bi + Mt + Il + Ap	$Q^{\ddagger}$+Or+Ab+An+C+Hy+Mt+Il+Ap.	Type I
Type A	Q + Kfel + Plag + Musc +Bi +Mt + Il + Ap		
Type B / Type D	Q + Kfel + Plag + Bi + Horn +Mt +Il + Ap		
Type E	Q + Kfel + Plag + Horn +Di +Mt + il + Ap	$Q^{\ddagger}$+Or+Ab+An+Di+Hy+Mt+Il+Ap	Type II

‡ In many cases, Q-bearing assemblages in the new scheme are represented by Ol-bearing assemblages in the C.I.P.W. scheme.

TABLE XV.

Eskola molecular norma using high-grade metamorphic minerals.

(a) Granitic gneisses.

	G11	G12	G13	G21	G22	G23	G31	G32	G33	G41	G42	G43	G51	G52	
Q	31.10	28.81	32.07	31.16	27.85	30.62	32.68	30.48	29.04	35.61	28.44	31.88	30.96	34.45	Q
Kfels	25.03	28.56	7.64	29.45	30.34	28.99	20.31	33.78	36.16	23.33	33.08	21.22	37.77	23.19	Kf
Plag	32.32	29.98	48.02	28.61	28.99	29.25	33.77	26.65	24.21	25.83	27.89	34.16	23.75	28.71	Pl
Musc	-	-	0.10	0.07	-	-	-	-	-	0.89	-	1.99	0.80	0.67	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Biot	8.15	9.53	11.00	9.52	5.98	2.85	12.37	6.39	3.54	13.93	6.41	9.06	5.73	11.62	Bi
Horn	2.28	1.73	-	-	5.11	6.11	0.09	1.78	4.60	-	2.98	-	-	-	Hb
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Mt	0.50	0.75	0.64	0.56	0.83	1.29	0.26	0.27	1.55	-0.06	0.59	1.00	0.51	0.73	Mt
Il	0.20	0.22	0.16	0.21	0.56	0.59	0.16	0.22	0.57	0.07	0.33	0.39	0.25	0.26	Il
Ap	0.42	0.42	0.35	0.40	0.35	0.27	0.35	0.42	0.32	0.38	0.30	0.30	0.25	0.37	Ap
% An in plag	12	21	21	20	22	19	20	15	22	24	18	20	18	22	X

	G53	G54	G55	G56	G61	G62	G63	G64	G65	G71	G72	G73	G74	G81	
Q	41.75	33.17	27.66	35.10	32.03	36.15	31.03	29.58	26.90	35.79	37.14	34.38	31.95	27.25	Q
Kfels	24.07	27.17	35.48	24.62	28.08	29.46	31.52	31.96	25.34	20.53	23.04	29.83	30.06	38.44	Kf
Plag	23.54	27.77	30.48	29.09	25.63	22.72	25.17	25.74	34.00	33.31	26.89	25.24	26.68	21.50	Pl
Musc	-	0.19	-	-	-	-	-	-	-	-	1.10	-	-	-	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Biot	4.74	10.30	2.36	6.44	10.83	6.15	6.24	7.40	12.49	8.98	10.38	8.83	7.34	6.52	Bi
Horn	3.32	-	1.57	2.84	2.06	4.31	5.15	4.34	0.10	0.37	-	0.51	2.94	5.53	Hb
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Mt	1.63	0.80	1.55	1.05	0.72	0.49	0.26	0.30	0.62	0.58	0.77	0.60	0.39	0.06	Mt
Il	0.54	0.24	0.50	0.51	0.25	0.39	0.36	0.27	0.17	0.23	0.35	0.27	0.29	0.33	Il
Ap	0.40	0.37	0.40	0.35	0.36	0.32	0.27	0.37	0.40	0.23	0.32	0.35	0.35	0.37	Ap
% An in plag	23	22	17	15	20	14	11	15	20	19	21	19	17	7	X

(Continued on next page)

TABLE XV. (Continued)

Eskola molecular norms using high-grade metamorphic minerals.

(a) Granitic gneisses, (Continued)

	G82	G83	G91	G92	G93	G96	G97	G98	G101	G102	G103	G105	G111	G112	
Q	30.28	38.14	35.70	30.94	28.05	38.52	31.78	28.92	35.18	35.98	29.83	27.56	31.71	35.71	Q
Kfels	25.90	15.76	33.41	25.66	31.21	33.29	30.24	33.89	40.85	23.18	29.83	33.34	34.66	34.89	Kf
Plag	30.77	28.20	24.27	30.42	27.07	20.59	28.10	26.55	17.92	28.31	35.69	28.75	24.74	25.47	Pl
Musc	0.24	2.50	0.61	0.03	-	1.38	0.45	-	-	1.72	1.76	-	-	-	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Biot	11.70	14.17	5.28	11.95	9.83	5.64	8.73	4.57	0.62	9.60	2.34	4.72	5.71	0.90	Bi
Horn	-	-	-	-	3.24	-	-	5.22	5.07	-	-	3.55	2.02	1.54	Hb
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Mt	0.44	0.64	0.38	0.49	0.09	0.32	0.23	0.15	-0.03	0.58	0.24	1.31	0.57	0.92	Mt
Il	0.24	0.23	0.21	0.14	0.16	0.15	0.21	0.46	0.29	0.32	0.17	0.47	0.37	0.43	Il
Ap	0.45	0.35	0.12	0.35	0.37	0.12	0.25	0.25	0.10	0.29	0.11	0.27	0.20	0.15	Ap
% An in plag	18	22	16	21	16	23	21	12	7	20	18	13	16	19	X

	GP112	G113	G116	G117	G121	G122	G123	G131	G132	G133	G141	G142	G151	G152	
Q	30.61	33.82	29.01	30.31	30.75	34.12	35.01	35.36	33.11	34.84	37.29	31.71	29.15	32.37	Q
Kfel	36.17	23.19	36.03	34.69	32.14	38.25	34.98	37.93	42.11	19.17	36.27	22.81	34.02	24.31	Kf
Plag	24.21	30.26	27.76	27.36	29.53	23.24	25.87	20.45	20.47	29.59	19.69	30.29	29.70	33.63	Pl
Musc	-	-	-	-	-	-	-	-	-	-	-	-	0.80	-	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Biot	6.03	8.26	2.22	0.41	4.07	-	2.19	2.26	0.69	10.79	2.64	12.19	5.40	5.43	Bi
Horn	0.40	2.45	3.07	5.59	2.12	0.69	1.10	3.27	2.07	4.75	2.21	1.32	-	2.06	Hb
Di	-	-	-	-	-	1.52	-	-	-	-	-	-	-	-	Di
Mt	1.65	1.30	1.19	0.86	0.83	1.41	0.44	0.30	0.99	0.24	1.24	0.85	0.52	1.39	Mt
Il	0.60	0.43	0.49	0.53	0.35	0.58	0.24	0.32	0.41	0.31	0.51	0.47	0.25	0.49	Il
Ap	0.30	0.27	0.23	0.25	0.23	0.20	0.12	0.12	0.15	0.30	0.15	0.32	0.15	0.32	Ap
% An in plag	26	27	18	14	19	14	20	14	18	26	17	28	15	26	X

(Continued on next page)

TABLE XV. (Continued)

Eskola molecular norms using high grade metamorphic minerals.

(a) Granitic gneisses, (Continued)

	G161	G162	G163	G171	G172	G173	G181	G182	G183	G184	G185	G186	G187	G191	
Q	33.71	28.31	33.12	32.27	28.64	32.46	31.65	30.69	32.94	29.55	33.48	31.91	31.48	29.18	Q
Kfels	38.56	32.05	31.75	19.18	26.38	27.31	27.46	29.58	24.68	23.97	25.32	24.23	31.34	27.11	Kf
Plag	22.38	26.42	25.70	33.53	31.43	29.41	29.80	30.60	29.59	32.45	28.65	30.01	25.45	30.79	Pl
Musc	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Biot	-	0.34	5.18	12.54	8.52	4.60	9.23	7.32	11.46	8.23	8.50	12.31	5.55	7.54	Bi
Horn	4.28	10.47	3.65	1.50	4. 18	4.56	0.99	0.12	0.16	4.13	2.30	0.64	5.13	4.11	Hb
Di	0.11	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Mt	0.40	1.29	0.10	0.41	0.23	0.83	0.51	1.01	0.55	0.87	0.89	0.28	0.31	0.56	Mt
Il	0.35	0.79	0.30	0.21	0.21	0.51	0.07	0.33	0.19	0.37	0.42	0.18	0.30	0.33	Il
Ap	0.17	0.35	0.20	0.35	0.40	0.32	0.27	0.35	0.40	0.45	0.42	0.42	0.42	0.40	Ap
% An in plag.	11	12	15	21	16	18	17	16	19	16	21	16	8	16	X

	G192	G193	G201	G202	G203	G211	G212	G213	G221	G222A	G223	G231	G232	G233	
Q	31.00	26.50	32.44	30.12	29.25	31.82	31.36	27.85	30.96	32.35	28.37	29.65	28.62	26.21	Q
Kfels	27.78	34.56	28.01	20.83	27.82	29.85	28.76	36.36	4.77	29.51	29.09	22.87	23.13	23.17	Kf
Plag	29.93	26.80	27.88	36.98	29.68	32.10	29.85	23.95	54.00	26.55	29.08	31.38	32.77	37.65	Pl
Musc	-	-	-	0.04	-	-	0.40	-	0.51	0.16	0.47	0.28	0.46	-	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Biot	7.97	6.18	10.00	10.46	9.51	3.20	7.50	7.45	8.98	10.46	11.25	14.54	13.88	11.30	Bi
Horn	2.40	4.43	0.48	-	2.06	0.17	-	3.02	-	-	-	-	-	0.20	Hb
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Mt	0.33	0.78	0.60	0.93	0.97	1.96	1.31	0.62	0.27	0.47	1.08	0.64	0.58	0.82	Mt
Il	0.25	0.36	0.24	0.31	0.33	0.55	0.40	0.29	0.10	0.19	0.27	0.21	0.13	0.24	Il
Ap	0.32	0.37	0.35	0.32	0.40	0.35	0.42	0.45	0.40	0.32	0.40	0.40	0.40	0.40	Ap
% An in plag	15	15	21	19	19	15	18	17	20	21	22	22	20	22	X

(Continued on next page)

TABLE XV. (Continued)

Eskola molecular norms using high grade metamorphic minerals.

(a) Granitic gneisses (Continued)

	G241	G242	G243	G244	G251	G252	G253	G254	G261	G262	G263	G271	G271A	G272	
Q	32.28	29.89	30.61	26.34	33.16	32.43	32.75	29.55	32.26	30.96	30.05	32.38	31.73	27.01	Q
Kfel	20.09	25.51	23.10	30.45	20.29	14.45	21.92	25.37	20.11	24.67	24.54	20.31	25.98	26.01	Kf
Plag	30.48	30.65	29.78	32.21	27.89	34.11	30.42	29.28	29.33	28.79	31.54	28.24	24.89	28.52	Pl
Musc	1.99	0.58	1.60	0.02	3.40	1.68	1.47	0.73	2.60	-	0.03	1.21	-	-	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Bi	13.93	12.36	13.23	9.48	13.88	16.27	12.15	13.84	14.63	13.97	12.94	16.77	13.23	13.10	Bi
Horn	-	-	-	-	-	-	-	-	-	1.16	-	-	3.48	4.23	Hb
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Mt	0.61	0.41	0.89	0.86	0.73	0.52	0.72	0.68	0.53	0.07	0.39	0.50	0.06	0.42	Mt
Il	0.19	0.18	0.29	0.17	0.19	0.09	0.15	0.07	0.13	0.05	0.11	0.14	0.18	0.26	Il
Ap	0.43	0.43	0.48	0.45	0.43	0.43	0.43	0.45	0.40	0.35	0.40	0.45	0.45	0.45	Ap
% An in plag	20	19	16	16	23	18	20	20	22	24	20	25	22	12	X

(b) Pegmatites, aplites, etc. (see p. 26)

	G273	G274	G281	G282	G283	G284	G291	G292	G293	GS166	G222	G262A	G274A	
Q	30.08	34.65	30.01	29.62	35.88	32.77	32.89	32.38	28.20	23.74	29.87	28.36	32.63	Q
Kfel	29.63	4.20	28.01	26.29	0.90	19.93	26.38	27.93	26.30	33.84	34.16	24.93	25.51	Kf
Plag	28.74	51.41	29.59	29.70	50.57	31.11	28.82	26.61	32.99	33.84	28.97	34.81	33.90	Pl
Musc	-	-	-	-	0.49	0.87	0.78	1.61	-	0.62	0.63	-	3.10	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Bi	9.09	8.75	8.68	11.65	11.08	13.80	8.94	9.85	11.04	7.37	5.36	8.36	4.16	Bi
Horn	0.97	0.19	2.61	1.92	-	-	-	-	-	-	-	3.12	-	Hb
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Mt	0.80	0.54	0.53	0.37	0.60	0.76	1.45	0.96	0.87	0.17	0.58	0.02	0.42	Mt
Il	0.26	0.12	0.27	0.21	0.15	0.22	0.32	0.23	0.20	0.23	0.23	0.13	0.20	Il
Ap	0.43	0.13	0.30	0.23	0.33	0.53	0.40	0.43	0.40	0.20	0.20	0.25	0.08	Ap
% An in plag	17	23	22	22	26	25	19	18	20	26	21	18	23	X

(Continued on next page)

TABLE XV. (Continued)

Eskola molecular norms using high grade metamorphic minerals.

(c) Psammites and striped schists from the Beinn an Tuim striped group and the Druim na Saille pelitic group.

(d) Psammites from the Glen Garvan psammitic group.

	Q187	K191	SK211	SK221	SK222	SK241	SK242	K271	K281	SK291	S42	S52	S71	
Q	75.89	64.82	48.09	67.93	57.40	68.11	80.57	53.15	47.48	50.97	21.42	37.47	41.79	Q
Kfel	7.55	15.38	23.58	12.12	2.12	21.44	4.30	7.82	4.56	4.82	34.67	37.82	11.96	Kf
Plag	12.89	13.12	21.73	17.18	35.68	7.51	9.49	32.39	36.34	37.84	29.02	18.49	39.29	Pl
Musc	1.74	4.17	2.28	0.04	0.26	1.06	3.19	0.57	1.99	1.56	0.83	0.28	-	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Bi	1.32	1.69	3.75	2.22	3.63	1.19	2.02	5.44	8.53	4.28	12.94	5.60	5. 73	Bi
Horn	-	-	-	-	-	-	-	-	-	-	-	-	0.42	Hb
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Mt	0.28	0.42	0.24	0.26	0.34	0.35	0.20	0.22	0.58	0.39	0.34	0.23	0.52	Mt
Il	0.28	0.30	0.26	0.20	0.31	0.27	0.18	0.26	0.28	0.09	0.60	0.05	0.23	Il
Ap	0.05	0.08	0.08	0.05	0.27	0.05	0.05	0.15	0.23	0.05	0.17	0.05	0.05	Ap
% An in plag.	15	16	16	19	20	15	21	25	22	18	20	23	28	X

	S77	S96	S101	S116	S121	S141	S142	S151	S162	S191	S213	S252	S261	
Q	51.77	55.20	61.06	43.96	44.66	34.12	73.96	55.52	77.26	66.70	44.48	46.58	56.45	Q
Kfel	33.37	31.51	24.09	6.61	7.28	17.54	17.79	12.86	15.17	19.35	30.04	0.76	31.77	Kf
Plag	9.97	9.85	13.18	41.98	40.37	37.73	5.34	22.13	2.87	11.63	21.09	42.35	7.50	Pl
Musc	-	0.91	0.36	0.68	-	-	1.30	5.33	2.21	0.46	2.16	0.97	1.80	Mu
Gt	-	-	-	-	-	-	-	-	-	-	-	-	-	Gt
Bi	1.10	1.81	0.74	6.30	6.64	9.30	1.19	3.58	2.35	1.24	1.98	8.86	1.90	Bi
Horn	3.40	-	-	-	0.10	0.54	-	-	-	-	-	-	-	Hb
Di	-	-	-	-	-	-	-	-	-	-	-	-	-	Di
Mt	-0.03	0.25	0.33	0.35	0.40	0.13	0.15	0.14	0.07	0.22	0.18	0.27	0.12	Mt
Il	0.39	0.37	0.20	0.05	0.45	0.52	0.21	0.35	0.04	0.37	0.02	0.14	0.37	Il
Ap	0.05	0.10	0. 02	0.07	0.12	0.12	0.05	0.07	0.02	0.03	0.03	0.08	0.10	Ap
% An in plag.	7	22	17	25	31	24	21	18	26	9	17	38	20	X

(Continued on next page)

TABLE XV. (Concluded)

Eskola molecular norms using high grade metamorphic minerals.

	(e) Pelitic schists and oligoclase gneisses.									
	P42	P53	P81	P96	P102	P125	P132	P142	P161	
Q	26.76	25.52	28.45	19.89	19.12	19.40	10.40	30.21	40.48	Q
Kfel	1.58	2.07	3.01	0.80	1.51	-	19.19	10.50	7.08	Kf
Plag	47.63	55.59	45.90	44.94	53.16	49.95	39.40	31.60	37.73	Pl
Musc	0.34	1.37	2.30	0.69	-	0.34	-	8.24	0.34	Mu
Gt	-	-	-	-	-	4.43	-	-	-	Gt
Bi	23.15	14.79	19.94	33.42	24.29	24.78	22.79	18.50	13.55	Bi
Horn	-	-	-	-	1.03	-	8.38	-	-	Hb
Di	-	-	-	-	-	-	-	-	-	Di
Mt	0.26	0.32	0.11	-0.17	0.35	0.33	-0.61	0.51	0.21	Mt
Il	-0.07	0.26	-0.10	-0.01	0.12	0.32	0.07	0.16	0.35	Il
Ap	0.33	0.10	0.38	0.42	0.42	0.45	0.37	0.27	0.25	Ap
% An in plag	22	33	23	29	31	30	13	21	27	X

								(f) Calcareous psammites.		
	P166	P181	P192	P223	P233	P273	P291	L116	S172	
Q	22.54	32.16	6.16	25.82	25.31	25.49	29.74	47.48	30.53	Q
Kfel	-	4.19	24.72	3.06	3.68	3.70	4.13	2.72	3.91	Kf
Plag	40.30	43.21	40.27	50.51	48.06	34.41	43.12	42.01	43.80	Pl
Musc	7.76	2.13	-	-	0.94	9.28	1.32	0.51	-	Mu
Gt	1.11	-	-	-	-	-	-	-	-	Gt
Bi	27.35	17.92	24.84	18.55	20.85	25.54	20.60	7.46	11.55	Bi
Horn	-	-	3.16	1.62	-	-	-	-	8.73	Hb
Di	-	-	-	-	-	-	-	-	-	Di
Mt	0.46	0.13	0.34	0.04	0.47	1.00	0.63	0.01	0.69	Mt
Il	-0.03	-0.04	-0.13	-0.07	0.03	-0.11	0.02	-0.29	0.22	Il
Ap	0.50	0.30	0.62	0.45	0.65	0.69	0.42	0.10	0.55	Ap
% An in plag.	26	23	13	23	26	23	33	46	48	X

TABLE XVI.

Effect of experimental error on Eskola molecular norms.

		C.I.P.W. scheme		High-grade metamorphic scheme.			
		Type I	Type II	Type C	Type A	Types B & D	Type E
Table lists the error produced in the amount of a mineral by an error of 2xS.D. in the analytical determination of all the elements involved in the mineral in question.	Q	±2.06	±1.97	±1.72	±1.87	±2.28	±1.91
	Or	±0.60	±0.60	-	-	-	-
	Ab	±1.10	±1.60	-	-	-	-
	An	±0.47	±1.75	-	-	-	-
	Kfels	-	-	-	±2.39	±3.90	±0.75
	Plag	-	-	±1.57	±1.85	±3.07	±2.25
	C	±0.89	-	-	-	-	-
	Musc	-	-	±1.66	±1.93	-	-
	Gt	-	-	±1.85	-	-	-
	Biot	-	-	±2.12	±0.84	±4.20	-
	Hy	±0.41	±1.60	-	-	-	-
	Horn	-	-	-	-	±5.12	±4.96
	Di	-	±1.78	-	-	-	±2.94
	Mt	±0.13	±0.13	±0.19	±0.15	±0.44	±0.42
	Il	±0.03	±0.03	±0.15	±0.08	±0.16	±0.16
	Ap	±0.03	±0.03	±0.03	±0.03	±0.03	±0.03
Total Dark Mins.*		±1.25	±1.20	±1.58	±2.23	±1.52	±1.36

* Including muscovite or corundum.

		Type I	Type II	Type C	Type A	Types B & D	Type E
Error in An		±0.47	±1.75	±0.47	±0.47	±3.22	±1.61
Error in Total Plag.		±1.57	±1.75	±1.85	±1.57	±3.07	±2.25
Variation in calculated composition of plagioclase when actual plagioclase is An	(Amount of plag. 50%	18-22	16-24	18-22	18-22	13-28	16-24
	40%	18-22	15-25	17-22	18-22	11-30	15-25
	30%	18-23	13-27	17-23	18-23	8-34	14-27
	20%	16-24	10-32	16-25	16-24	3-43	11-32
	10%	13-29	2-44	13-30	13-29	0-75	3-47
	5%	8-43	0-85	8-47	8-43	0-100	0-95

TABLE XVII.

Analyses of Biotites and Hornblende.

(a) From Granitic Gneiss - Biotites.

Rock	G12	G13	G62	G92	G103	G163	G183	G241	G262	G281
SiO_2	33.49	33.75	32.90	33.86	34.02	33.99	33.42	33.88	31.50	33.41
TiO_2	3.98	3.81	3.57	3.63	3.65	4.20	3.68	3.66	3.69	3.66
Al_2O_3	16.75	17.24	17.37	16.90	17.66	15.60	17.64	18.14	17.28	15.94
Fe_2O_3	3.73	3.43	3.53	2.59	3.05	3.03	3.67	2.68	1.84	3.92
FeO	23.81	23.38	23.31	25.42	21.93	24.62	24.12	23.54	27.42	24.94
MnO	0.52	0.51	0.37	0.46	0.38	0.40	0.43	0.43	0.38	0.44
MgO	4.82	4.97	5.81	4.11	6.33	5.74	4.34	4.50	4.07	3.79
Li_2O	0.18	0.19	0.14	0.16	0.15	0.14	0.19	0.18	0.16	0.19
CaO*	0.13	0.19	0.37	0.11	0.66	0.20	0.05	0.22	1.31	0.86
Na_2O	0.18	0.21	0.27	0.20	0.18	0.17	0.17	0.14	0.16	0.17
K_2O	9.09	9.14	7.81	8.62	7.56	8.14	8.82	9.30	6.41	8.60
P_2O_5*	0.01	0.07	0.04	0.04	n.d.	0.04	0.01	0.15	0.36	0.45
H_2O^+	2.90	2.99	3.71	3.14	3.71	3.03	3.03	3.05	4.94	2.95
F	0.32	0.34	0.22	0.29	0.38	0.37	0.32	0.32	0.23	0.29
Total	99.91	100.22	99.44	99.53	99.66	99.67	99.89	100.19	99.75	99.61
less O for F	0.13	0.14	0.09	0.12	0.16	0.16	0.13	0.13	0.10	0.13
Total	99.78	100.08	99.35	99.41	99.50	99.51	99.76	100.06	99.65	99.48
ppm Rb	630	605	510	595	n.d.	600	565	560	410	610
ppm Ba	705	770	625	575	n.d.	635	580	710	495	635
ppm Sr	10	16	22	16	n.d.	8	10	14	26	24

* P_2O_5 determined in order to correct Ca in mineral formula for apatite in separation; the amount of purified biotite obtained from G103 was insufficient for determination of P, Ba, Rb and Sr.

Samples dried at 110°C for 12 hours prior to analysis.

(Continued on next page)

TABLE XVII. (Continued)

Analyses of Biotites and Hornblende.

	(b) From Pelitic schists and oligoclase gneisses - Biotites				(c) From psammites - Biotites				(d) From calcareous psammite	
Rock	P125	P192	P223	P291	S42	S77	S141	SK211	Biotite S172	Hornblende S172
SiO_2	34.65	34.77	35.17	35.45	33.09	34.86	34.57	34.78	35.83	40.07
TiO_2	3.65	3.42	3.45	2.83	4.88	3.84	4.13	3.17	3.53	1.26
Al_2O_3	18.97	18.28	18.26	18.41	17.99	19.42	17.57	19.06	17.11	13.65
Fe_2O_3	1.48	1.16	1.00	2.12	2.50	1.17	2.22	1.62	3.80	6.61
FeO	20.13	20.14	20.41	18.67	21.21	20.88	21.39	21.17	17.99	13.29
MnO	0.20	0.44	0.46	0.28	0.21	0.31	0.32	0.62	0.44	0.85
MgO	7.63	7.73	7.80	8.60	7.02	6.96	7.08	6.63	8.63	7.56
Li_2O	0.13	0.18	0.16	0.14	0.15	0.18	0.15	0.16	0.17	0.06
CaO*	0.12	0.23	0.06	0.16	0.29	0.24	0.21	0.03	0.71	12.01
Na_2O	0.31	0.23	0.20	0.27	0.28	0.23	0.17	0.21	0.26	1.23
K_2O	9.10	9.65	9.31	9.19	8.32	8.93	8.66	9.15	8.81	1.09
P_2O_5*	0.01	0.12	0.01	0.07	0.06	0.00	0.01	0.01	0.08	0.85
H_2O^+	2.89	2.72	2.92	3.01	3.21	3.16	3.03	2.94	2.81	1.33
F	0.17	0.29	0.23	0.23	0.23	0.29	0.14	0.24	0.30	0.20
Total	99.44	99.34	99.44	99.43	99.44	100.47	99.65	99.79	100.47	100.06
Less O for F	0.07	0.12	0.10	0.10	0.10	0.12	0.06	0.10	0.13	0.08
Total	99.37	99.22	99.34	99.33	99.34	100.35	99.59	99.69	100.34	99.98
ppm Rb	365	460	450	335	415	645	455	515	340	11
ppm Ba	2300	760	1740	1670	760	350	810	745	2700	35
ppm Sr	8	14	14	26	8	18	20	14	22	48

* P_2O_5 determined in order to correct Ca in mineral formula for apatite in separation.

Samples dried at 110°C for 12 hours prior to analysis.

TABLE XVIII.

Mineral Formulae for Biotites and Hornblende,

Calculated to 24 (O, OH, F) by the method of Foster (1960).

(a) From granitic gneisses - biotites.

Rock.		G12	G13	G62	G92	G103	G163	G183	G241	G262	G281
Z	Si	5.235	5.250	5.164	5.330	5.262	5.321	5.219	5.267	5.099	5.323
	Al^{IV}	2.765	2.750	2.836	2.670	2.738	2.679	2.781	2.733	2.901	2.677
Y	Al^{VI}	.319	.409	.376	.463	.479	.198	.464	.589	.394	.315
	Ti	.468	.445	.421	.429	.424	.494	.432	.428	.449	.438
	Fe^{3+}	.438	.401	.417	.307	.354	.357	.431	.313	.224	.469
	Fe^{2+}	3.111	3.039	3.058	3.341	2.835	3.221	3.148	3.058	3.710	3.321
	Mn	.069	.067	.049	.061	.050	.053	.057	.056	.052	.059
	Mg	1.122	1.151	1.358	.964	1.458	1.338	1.009	1.042	.981	.899
	Li	.113	.119	.089	.101	.093	.088	.119	.112	.104	.121
X	Ca*	.020	.016	.054	.010	.109	.025	.007	.004	.145	.045
	Na	.054	.064	.082	.061	.054	.052	.052	.042	.051	.053
	K	1.811	1.812	1.563	1.730	1.491	1.624	1.755	1.843	1.323	1.747
W	OH	3.020	3.099	3.880	3.294	3.823	3.161	3.154	3.160	5.328	3.132
	F	.158	.167	.110	.145	.186	.183	.158	.157	.118	.146
Total Z		8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Total Y		5.640	5.631	5.768	5.666	5.693	5.749	5.660	5.598	5.914	5.622
Total X		1.885	1.892	1.699	1.801	1.654	1.701	1.814	1.889	1.519	1.845
Total W		3.178	3.266	3.990	3.439	4.009	3.344	3.312	3.317	5.446	3.278
Tetrahedral charge		-2.765	-2.750	-2,836	-2.670	-2.738	-2.679	-2.781	-2.733	-2.901	-2.677
Octahedral charge		+ .859	+0.846	+1.082	+ .860	+0.974	+ .952	+ .960	+ .841	+1.239	+ .787
Composite charge		-1.906	-1.904	-1.754	-1.810	-1.764	-1.727	-1.821	-1.892	-1.662	-1.890
Charge on X group		+1.905	+1.908	+1.753	+1.811	+1.763	+1.726	+1.821	+1.892	+1.664	+1.890

* Corrected for apatite in analysed material.

(Continued on next page)

TABLE XVIII. (Continued)

Mineral Formulae for Biotites and Hornblende

Calculated to 24 (O, OH, F) by the method of Foster (1960).

		(b) From pelitic schists and oligoclase gneisses - Biotites				(c) From Psammites - Biotites				(d) From calcareous psammite	
										Biotite	Hornblende
Rock		P125	P192	P223	P291	S42	S77	S141	SK211	S172	S172
Z	Si	5.264	5.319	5.355	5.371	5.106	5.263	5.289	5.303	5.377	6.130
	Al^{IV}	2.736	2.681	2.645	2.629	2.894	2.737	2.711	2.697	2.623	1.870
Y	Al^{VI}	.659	.613	.630	.657	.376	.716	.456	.726	.402	.590
	Ti	.417	.393	.395	.322	.566	.436	.475	.363	.398	.145
	Fe^{3+}	.169	.133	.115	.241	.290	.133	.255	.186	.429	.760
	Fe^{2+}	2.556	2.575	2.597	2.364	2.735	2.634	2.735	2.698	2.256	1.699
	Mn	.026	.057	.059	.036	.027	.039	.041	.080	.056	.110
	Mg	1.726	1.761	1.769	1.941	1.613	1.564	1.613	1.505	1.928	1.722
	Li	.079	.110	.098	.086	.093	.109	.092	.098	.103	.037
X	Ca*	.018	.012	.008	.011	.035	.038	.033	.003	.097	1,783
	Na	.091	.068	.059	.079	.083	.067	.051	.062	.076	.365
	K	1.762	1.882	1.808	1.755	1.636	1.719	1.689	1.779	1.685	.212
W	OH	2.925	2.773	2.963	3.039	3.300	3.179	3.090	2.987	2.811	1.356
	F	.081	.141	.111	.110	.112	.139	.068	.115	.142	.096
Total Z		8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Total Y		5.632	5.642	5.663	5.647	5.700	5.631	5.667	5.656	5.572	5.063
Total X		1.871	1.962	1.875	1.865	1.754	1.824	1.773	1.844	1.858	2.360
Total W		3.006	2.914	3.074	3.149	3.412	3.318	3.158	3.102	2.953	1.452
Tetrahedral Charge		-2.736	-2.681	-2.645	-2.629	-2.894	-2.737	-2,711	-2.697	-2.623	-1.870
Octahedral Charge		+ .845	+ .706	+ .762	+ .751	+1.105	+ .874	+ .905	+ .851	+ .669	+1.728
Composite Charge		-1.891	-1.975	-1.883	-1.878	-1.789	-1.863	-1.806	-1.846	-1.954	- .142
Charge on X group		+1.889	+1,974	+1,883	+1,876	+1,789	+1.862	+1.806	+1.847	+1.955	+4.143

* Corrected for apatite in analysed material.

TABLE XIX.

Partial Analyses of Potash Felspars and Muscovites.

(a) From granitic gneisses - Potash Felspars.										(b) From oligoclase gneiss - Potash felspar	(c) From psammites-Potash Felspare.			
Rock	G12	G13	G62	G92	G103	G163	G183	G241	G262	G281	P.192	S42	S77	S141	SK211
CaO	0.117	0.412	0.159	0.113	0.403	0.097	0.079	0.064	0.253	0.166	0.127	0.101	0.080	0.209	0.109
Na_2O	1.61	2.49	1.80	1.53	2.85	1.42	1.41	1.36	2.68	1.51	1.49	1.41	1.79	1.62	1.47
K_2O	14.64	12.62	14.29	14.68	12.14	14.71	14.56	14.51	12.72	14.33	14.65	14.81	14.44	14.16	14.69
Mol.% An	0.6	2.1	0.8	0.6	2.0	0.5	0.4	0.3	1.2	0.8	0.6	0.5	0.4	1.0	0.5
Mol.% Ab	14.2	22.6	15.9	13.6	25.8	12.7	12.8	12.4	24.0	13.7	13.3	12.6	15.8	14.7	13.1
Mol.% Or	85.2	75.3	83.3	85.8	72.2	86.8	86.8	87.3	74.8	85.5	86.1	86.9	83.8	84.3	86.3
ppm Rb	344	276	285	280	237	360	310	261	272	313	213	210	327	235	150
ppm Ba	2800	3000	3100	3150	6750	2600	3100	4200	2800	3050	4350	4250	2700	4700	4700
ppm Sr	190	260	328	218	715	252	216	296	266	234	456	350	316	598	506

	Muscovites		
Rock	Granitic gneiss. G241	"Pelite" P291	Psammite. SK211
CaO	0.048	0.107	0.028
Na_2O	0.58	0.71	0.57
K_2O	10.69	10.47	10.71
H_2O+	3.93	3.82	3.84
F	0.14	0.06	0.11
Mol.% Marg	0.3	0.8	0.2
Mol.% Par	7.6	9.3	7.5
Mol.% Musc	92.1	90.0	92.3
ppm Rb	349	210	314
ppm Ba	1065	3450	1295
ppm Sr	28	128	34

Samples dried at 110°C for 12 hours prior to analysis.

TABLE XX.

Partial Analyses of Plagioclases.

(a) From granitic gneisses.

Rock		G12	G13	G62	G92	G103	G163	G183	G241	G262	G281
In original powder	CaO	2.29	3.07	2.24	3.15	1.85	2.44	2.89	3.42	2.31	2.79
	Na_2O	4.49	6.55	4.88	6.78	3.99	5.28	6.92	8.31	5.12	5.76
	K_2O	0.35	0.31	0.35	0.42	0.23	0.32	0.36	0.34	0.85	0.43
In analysed powder	ppm Rb	11	8	18	15	6	12	14	7	40	15
	ppm Ba	65	45	75	85	95	45	35	60	95	75
	ppm Sr	132	210	222	208	372	194	214	318	244	174
% Plag in separation		51.4	72.5	54.5	75.5	44.3	58.7	75.0	89.3	59.8	65.1
In plagioclase	CaO	4.45	4.24	4.11	4.17	4.17	4.16	3.86	3.83	3.86	4.29
	Na_2O	8.73	9.04	8.96	8.98	9.00	9.00	9.23	9.31	8.57	8.85
	K_2O	0.68	0.43	0.64	0.56	0.52	0.55	0.48	0.38	1.42	0.66
	Mol % An	21.1	20.1	19.5	19.8	19.8	19.7	18.2	18.1	18.4	20.3
	Mol % Ab	75.0	77.5	76.9	77.1	77.3	77.2	79.1	79.7	73.6	75.9
	Mol % Or	3.8	2.4	3.6	3.1	2.9	3.1	2.7	2.2	8.0	3.7
	ppm Rb	21	11	33	20	14	20	19	8	67	23
	ppm Ba	126	62	138	113	214	77	47	67	159	115
	ppm Sr	257	290	408	276	839	331	285	356	408	267

Samples dried at 110° C for 12 hours prior to analysis.

(Continued on next page)

TABLE XX. (Continued)

Partial Analyses of Plagioclases.

Rock		(b) From pelitic schists and oligoclase gneisses				(c) From psammites				(d) From calcareous psammite.
		P125	P192	P223	P291	S42	S77	S141	SK211	S172
In analysed powder	CaO	5.85	3.13	2.96	6.35	3.09	0.45	2.52	1.66	12.74
	Na_2O	7.42	9.28	5.47	7.24	7.95	1.11	4.76	5.34	2.94
	K_2O	0.35	0.37	0.37	0.45	0.44	0.15	0.57	0.23	0.46
	ppm Rb	8	7	11	16	5	7	16	3	19
	ppm Ba	225	95	100	65	120	45	95	15	135
	ppm Sr	745	572	466	985	368	54	422	348	805
% Plag in separation		93.9	96.2	63.1	95.4	85.2	12.5	56.1	54.8	90.8
In plagioclase	CaO	6.24	3.25	4.69	6.65	3.63	3.60	4.49	3.03	14.04
	Na_2O	7.91	9.64	8.66	7.59	9.33	8.87	8.48	9.75	3.24
	K_2O	0.37	0.38	0.59	0.47	0.52	1.20	1.02	0.42	0.51
	Mol % An	29.7	15.4	22.3	31.8	17.2	17.1	21.3	14.3	68.5
	Mol % Ab	68.2	82.5	74.4	65.5	79.9	76.2	72.9	83.3	28.6
	Mol % Or	2.1	2.2	3.3	2.7	2.9	6.8	5.7	2.4	2.9
	ppm Rb	9	7	22	17	6	56	28	5	21
	ppm Ba	240	99	158	68	141	360	169	27	149
	ppm Sr	794	594	738	1032	432	432	752	635	887

Samples dried at 110°C for 12 hours prior to analysis.

GEOLOGICAL MAP OF THE GRANITIC GNEISS AND ASSOCIATED ROCKS OF WESTERN ARDGOUR, ARGYLL.

after Harry (1953), Dalziel (1963a), and Johnson (unpublished).

LEGEND

METASEDIMENTS OF THE MOINE SERIES

- Upper (Glen Garvan) Psammitic Group.
- Druim na Saille Pelitic Group (including oligoclase-quartz-biotite gneisses).
- Beinn an Tuim Striped Group.

ARDGOUR GRANITIC GNEISS

- Ordinary Granitic Gneiss.
- Augen Gneiss.

INTRUSIVE ROCKS AND ROCKS MOBILISED BY INTRUSIONS

- Diorite.
- Zone of Mobilisation in schists around Diorite.
- Strontian "Granite" and other granitic intrusions.

70 Strike and dip of bedding schistosity and foliation.

Fault.

Geological Boundary.

°G273 Analysed Sample.

Numbers round the margin of the map are those of the National Grid lines for the area.

SCALE

0 1 2 MILES.

0 1 2 3 KILOMETRES.

N

Beinn an Tuim
Druim na Saille
Glentinnan
LOCH SHIEL
Callop River
GLEANN DUBH LIGHE
Finnan Dubh Lighe
LOCH EIL
Sgor Craobh a' Chaoruinn
Sgurr Ghiubhsachain
Glas Bheinn
Meall nan Damh
Finnan
GLEN GARVAN
Cona River
Meall Mor
Stob a'Chuir
Hurich
GLEN HURICH
Lochan Dubh
Beinn Mheadhoin
Druim Garbh
Sgurr Dhomhnuill
Druim Glas
Meall Iain
Loch Doilet
River Strontian
Sgurr a' Chaoruinn
Sgurr nan Conbhairean
GLEN COUR
River Cour
Loch Coire na Creiche
Strontian
Garbh Bheinn
Meall a' Chulinn
GLEN TARBERT
LOCH SUNART
LOCH LINNHE
GLEN SCADDLE