

PETROLOGY AND GEOCHEMISTRY OF IGNEOUS ROCKS
IN THE COMBOYNE PLATEAU - LORNE BASIN AREA,
NEW SOUTH WALES

by

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THESIS SUBMITTED FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY
IN THE SCHOOL OF EARTH SCIENCES,
MACQUARIE UNIVERSITY,

SYDNEY

MAY, 1975

This thesis is based on studies carried out in the School of Earth Sciences, Macquarie University, from March, 1971 to May, 1975. I certify that the thesis has not been submitted to any other university or institution. All sources of information and assistance given in the preparation of the thesis have been clearly acknowledged.

J. Knutson

28-5-75

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ACKNOWLEDGEMENTS

This study was carried out while holding a Commonwealth Postgraduate Award, and this assistance is gratefully acknowledged.

I wish to express my sincere appreciation to Dr. T.H. Green, who supervised this work, and to Dr. S.Y. Wass, for their valuable guidance, encouragement and helpful discussions throughout the study. Their critical reading of the manuscript is acknowledged with many thanks.

Sincere thanks are also due to Drs. R.H. Flood, S.E. Shaw, and Associate Professor R.H. Vernon for providing valuable discussions and advice on many aspects of the work. In particular, the assistance of Drs. R.H. Flood and S.E. Shaw in X-ray fluorescence techniques is much appreciated.

I would also like to express my appreciation to Dr. I. McCougall of the Research School of Earth Sciences, Australian National University, Canberra, for undertaking K/Ar age dating of some of the Lorne Basin calc-alkaline rocks.

Access to the TPD electron microprobe at the Research School of Earth Sciences, Australian National University, is much appreciated. Sincere thanks are due to Dr. S.J.B. Reed and Mr. N.G. Ware of the above School for their assistance in electron microprobe techniques.

I also wish to thank the postgraduate students and technical staff of the School of Earth Sciences, Macquarie University, for both helpful discussions and practical assistance. In particular, the help of Mr. H.S. Ackland is gratefully acknowledged.

I am grateful to Mr. K.T. Kendall and the cartographic staff of the School of Earth Sciences, Macquarie University, for assistance with drafting and photography.

The assistance of Mrs. M.E. Allen in the field, and Mrs. P. Knowles in collating the manuscript, is much appreciated.

Finally, I am forever grateful to my mother for her constant encouragement and understanding.

SUMMARY

Mineralogical and geochemical data indicate that the igneous rocks in the Lorne Basin - Comboyne Plateau area on the mid-north coast of New South Wales consist primarily of alkaline and calc-alkaline types, with minor occurrences of tholeiitic andesite.

Alkaline rocks are centred on the Comboyne Plateau, and include hawaiite, mugearite, benmoreite, mafic trachyte, anorthoclase trachyte and comendite. The rocks form domes, plugs and flows and there is a dominance of intermediate and silicic types. There are also rare occurrences of tholeiitic andesite on the Comboyne Plateau. The calc-alkaline rocks are predominantly high level intrusive granodiorites and granites and extrusive rhyolites. More basic types include basaltic andesite and diorite.

Data presented in this study indicate that the Comboyne alkaline rocks represent a mildly alkaline-transitional association in which strong fractionation has resulted in the formation of peralkaline rhyolites. Geochemically the Comboyne association has tholeiitic affinities with normative hypersthene occurring in all rock types. Mineralogically they are alkaline, although analyses and normative components for clinopyroxenes in the more basic rocks confirm the mildly alkaline-transitional character of the host rocks.

Textural and chemical criteria suggest that megacrysts and/or cumulates of orthopyroxene, olivine, clinopyroxene, plagioclase, ilmenite, titanomagnetite and apatite in the hawaiites, and less commonly, in the mugearites, crystallised at elevated pressures. Similarly, crystal aggregates of anorthoclase, olivine, edenitic amphibole, Fe-Ti oxides and apatite in the benmoreites indicate that these also formed at depth and that fractionation in near-surface magma chambers was not significant, at least in obtaining

compositions as evolved as benmoreite.

100Mg/Mg + Fe²⁺ values of less than 50 in the hawaiites indicate that these rocks do not represent primitive mantle melts, and that fractionation of magnesium-rich phases occurred prior to the crystallisation of the present megacryst/cumulate assemblage.

The derivation of the hawaiites at elevated pressures is supported by experimental work undertaken in this study. The near-duplication of the natural megacryst/cumulate assemblage in runs at 6.5 kb with 2% H₂O indicates that the hawaiite megacryst/cumulate assemblages precipitated at a depth of 20 - 24 km before rapid ascent to higher crustal levels.

It is suggested that megacrysts of orthopyroxene and olivine, and gabbroic aggregates of plagioclase, clinopyroxene, olivine and ilmenite, in the Comboyne tholeiitic andesite precipitated at pressures up to 6 kb. This is supported by mineralogical and experimental data on similar compositions from other basaltic associations.

Both major and trace element data indicate that the Lorne Basin granodiorites, granites and rhyolites are not genetically related to the more basic diorites and basaltic andesite. Partial melting of andesitic material in the crust best explains the origin of the silicic rocks, but the relatively high MgO, FeO and CaO in the diorites and basaltic andesite points to their origin from the mantle by a two-stage melting process.

CHAPTER 1 INTRODUCTION

1.1 Introduction

The primary and deep level origin of basaltic magmas has been postulated for some time (Bowen, 1928; Tilley, 1950; Powers, 1955) and is supported by experimental work (Yoder and Tilley, 1962; Green and Ringwood, 1967; O'Hara, 1968; Ito and Kennedy, 1968; Kushiro, 1968; Bultitude and Green, 1971) and the recognition of lherzolite nodules and high-pressure cumulate and megacryst phases in silica-undersaturated basaltic rocks (Green and Ringwood, 1967; Binns et al., 1970. Aoki, 1970; Best, 1970; Wass, 1971, Irving, 1974). High-pressure phases have also been reported from more evolved alkaline as well as tholeiitic basalts (Kuno, 1964; Binns et al., 1970; Wright, 1969; Green et al., 1974; Duggan and Wilkinson, 1973), and the occurrence of cumulates and megacrysts in several near-saturated and saturated hawaiites in the Comboyne area on the mid-north coast of New South Wales gives the opportunity to test experimentally the hypothesis that these relatively evolved rocks have been generated at elevated pressures.

The dominance of intermediate and acid rocks relative to basic types in the Comboyne alkaline association highlights one of the major problems relating to the petrogenesis of some alkaline provinces in both continental and, more rarely, oceanic environments, where intermediate and/or silicic lavas are greatly in excess of the associated basic types. Similarly, the close association of proportionally large volumes of peralkaline rhyolites (comendites) with the Comboyne mildly alkaline-transitional rocks emphasises the problem of the genetic relationship, if any, of peralkaline rhyolites with associated basic and intermediate rocks in this and other locations.

The aim of this petrological, geochemical and experimental study is to outline the petrogenesis of these near-saturated and saturated alkaline and peralkaline rocks, and to determine whether there is geochemical continuity and a demonstrable genetic link between the major occurrences of igneous rocks in the area. Such links have generally been inferred up to the present time.

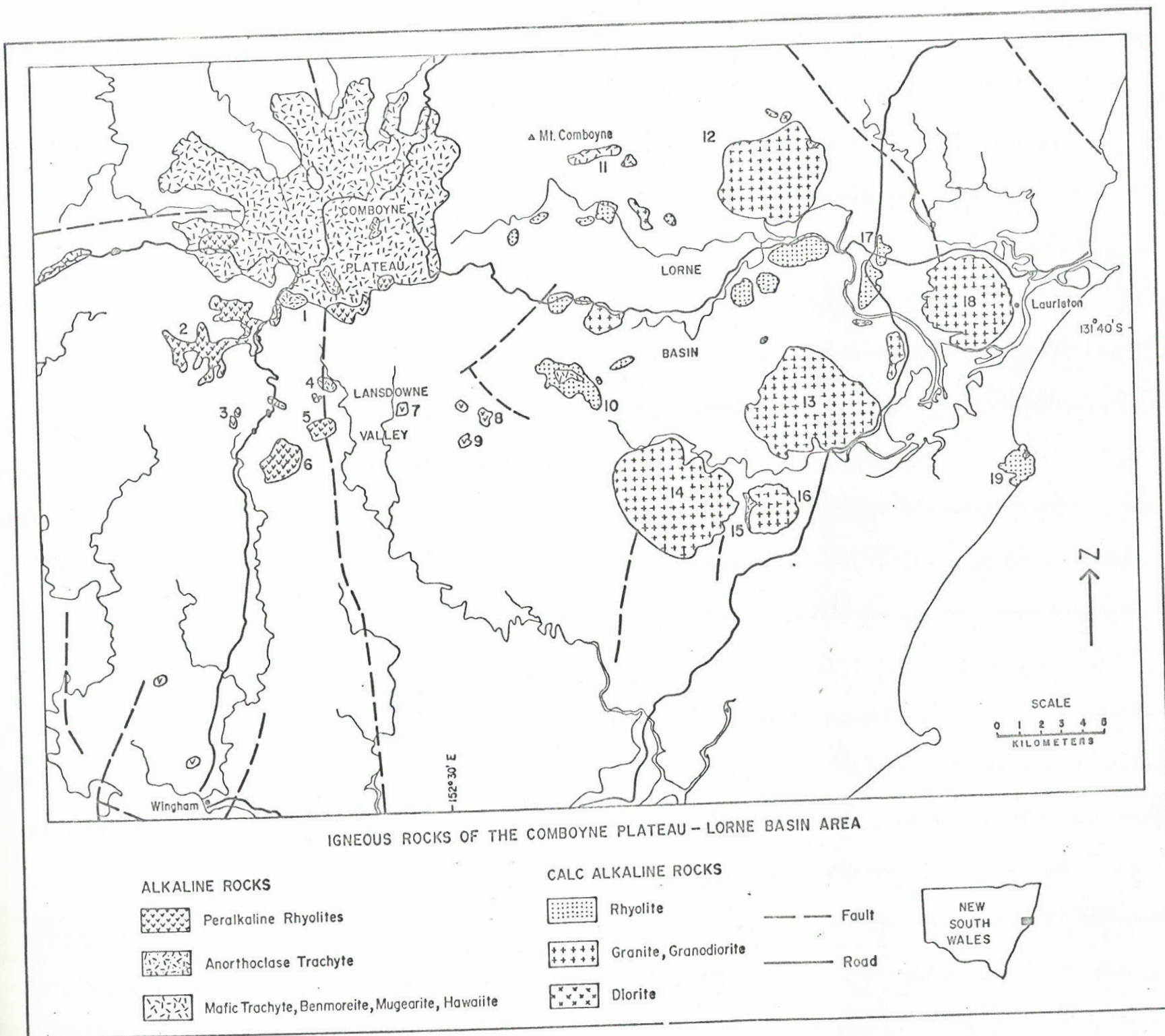
The area of this study totals approximately 2,400 km², and is bounded by latitudes 31°30' and 32°00' E and longitudes 152°10' and 152°50' S. Widespread igneous activity in the Lorne Basin, Lansdowne Valley and Comboyne Plateau (Fig. 1.1) is reflected by the prominence of stocks, plugs and lava flows. Alkaline rock types occurring in the area include hawaiite, mugearite, benmoreite, mafic trachyte, anorthoclase trachyte, comendite and alkaline pitchstone. Sub-alkaline rocks consist primarily of silicic calc-alkaline types, but rare occurrences of tholeiitic rocks have been found on the Comboyne Plateau. Volcanic calc-alkaline rocks include andesite, dacite and rhyolite, with diorite, granodiorite, and granite forming high-level intrusive bodies.

1.2 Previous Work

The Three Brothers (Fig. 1.1) were named by James Cook on May 11, 1770 (Cook, 1773) when he calculated their latitude as 31°40' S. They were again mentioned by Oxley in 1817-1818, and Carne (1897) noted diorite, quartz-felsite and rhyolite at Diamond Head. Sussmilch (1932), Browne (1933) and Chalmers (1934) referred to the volcanic plugs in the Lansdowne Valley and Browne identified comendite, arfvedsonite anorthoclase trachyte and

Fig. 1.1 Locality map for the igneous rocks in the Comboyne Plateau - Lorne Basin area. Map compiled from Hastings 1:250,000 Geological Sheet, Comboyne, Wingham, Camden Haven and Cundle 1:63,360 Military Maps, and Wingham North and South and Camden Haven North and South 1:50,000 Photomaps.

1. Mt. Gibraltar
2. Mt. Killabahk
3. The Monastery
4. Mt. Baldy
5. Mt. Coxcomb
6. Mt. Goonook
7. Mt. Oliver
8. Big Nellie
9. Little Nellie
10. Hannon Vale
11. Lorne Basin hawaiiite
12. Lorne Forest
13. Middle Brother
14. Juhle Mountain
15. Moorland
16. South Brother
17. Kew
18. North Brother
19. Diamond Head



bostonite. Voisey (1938, 1939a, 1939b, 1950) described the geology of the area and referred to the stratigraphy in some detail. More recently, Stewart (1953), in a general study of the geology of the Camden Haven area, described the stratigraphy, as well as the petrography of the igneous and metamorphic rocks. McDougall and Wilkinson (1967) determined the age (K/Ar) of the Lansdowne Valley Mt. Oliver comendite, and their paper includes the only rock analysis known to exist from the area prior to the present study. The New South Wales Department of Mines has mapped the general geology of the area and this is presented in the Hastings 1:250,000 Geological Series Sheet dated 1970. In a reappraisal of the Lorne Basin sedimentary rocks, Pratt and Herbert (1973) correlate the Camden Haven Group with the top part of the Sydney Basin Narrabeen Group which dates these rocks as early Triassic.

Before the present study the igneous rocks in the Lorne Basin - Comboyne Plateau area have generally been described as alkaline, although references by Carne (1897) to diorite and rhyolite imply that he considered the Diamond Head rocks to be sub-alkaline. It is now apparent that whereas the majority of rocks in the Comboyne-Lansdowne area have alkaline affinities, a substantial proportion of those in the Lorne Basin are sub-alkaline.

1.3 Distribution and Field Relationships of the Igneous Rocks

1.3.1 Alkaline Rocks

As already noted, the bulk of the alkaline rocks occurs on the Comboyne Plateau, although there are significant occurrences in the Lansdowne Valley, Lorne Basin and to the north of Wirrham.

Hawaiite, the most basic alkaline rock located in the area,

generally forms boulder slopes relatively low down on the periphery of the Comboyne Plateau, and it is particularly common on the western and northwestern escarpments (Fig. 1.1).

Boulder slopes of hawaiite also occur on the northwestern rim of the Lorne Basin, to the east of Mt. Comboyne. The most spectacular occurrence of hawaiite is the Lansdowne Valley, Mt. Baldy plug, which rises steeply from the valley floor. Cumulate and megacryst (see page 24) phases are most abundant in the Mt. Baldy and Lorne Basin hawaiites and occur only sporadically in the other samples. With the exception of Mt. Baldy, the absence of in-situ outcrop makes it impossible to determine whether the hawaiites represent volcanic flows or high-level intrusives.

Only two specimens of mugearite have been positively identified from the Comboyne Plateau. One contains megacrysts and occurs near the southeastern escarpment, and the other forms a boulder slope relatively low down on the northwestern escarpment. Again, the absence of field relationships precludes the positive identification of these as high-level intrusives or volcanic flows, although the fine grain size of the groundmass suggests the latter. Mugearites also occur to the west of Hannon Vale in the Lorne Basin. These rocks form a number of terraces and overlie calc-alkaline rhyolites. Their fine grain size supports field evidence indicating they represent a series of volcanic flows.

Benmoreites outcrop extensively at a relatively high elevation on the Comboyne Plateau and a wide variation in grain size suggests that they formed high-level intrusives and flows. Mafic trachytes tend to be at a higher stratigraphic level than the benmoreites and their very fine grain size indicates they are volcanic flows. Good outcrops of mafic trachyte occur on the

eastern and northeastern escarpment of the plateau. However, lack of continuous outcrop associated with rugged terrain limits the correlation of individual trachyte flows. A small area of outcrop low down on the southern escarpment of the Comboyne Plateau on the Comboyne-Wingham road is isolated from the major occurrences of trachytic rock. Here porphyritic and vesicular benmoreite/trachyte appears to intrude the Carboniferous Giro Beds and is in close proximity to a volcanic breccia pipe which contains fragments of vesicular benmoreite/trachyte, comendite and sedimentary material ranging up to 1.5 metres in diameter, but more commonly approximating 1.0 cm. A second breccia pipe occurs on the eastern escarpment of the Comboyne Plateau and is surrounded on three sides at least by comendite.

Anorthoclase trachyte forms the prominent hills Mt. Bulli and Mt. Kophi on the Comboyne Plateau and the Mt. Gibraltar plug which intrudes the eastern escarpment of the plateau. The form of these bodies suggests they are intrusive, but the fine grain-size of their matrix indicates final crystallisation occurred in a near-surface environment.

The Comboyne Plateau comendites tend to be concentrated towards the plateau's southern margin as stratigraphic highs. In a number of locations these comendites overlie mafic trachytes, but contact relationships have not been observed. With the exception of the comendite forming the summit of Mt. Killabahk, the Comboyne comendites are fine grained and exhibit excellent flow banding due to the alignment of mafic and felsic minerals. The Mt. Killabahk comendite has a bostonitic texture with relatively coarse equidimensional feldspar, quartz, aegirine and riebeckite/arfvedsonite. The form and texture of the Mt. Killabahk comendite

indicate it is a high-level intrusive body and probably a volcanic centre for the surrounding fluidal types. These comendites are considered to be extrusive, although high silica contents and resultant viscosity probably limited flow after extrusion.

Comendite plugs occurring in the Lansdowne Valley, the western rim of the Lorne Basin and to the north of Wingham, exhibit some variation in composition and texture, but the majority are relatively coarse-grained, pointing to their intrusive nature. The large dome-like body of Mt. Goonook consists of fluidal, non-porphyrific comendite, and probably a large proportion of this mountain was formed from lava flows emanating from the summit. Somewhat similar fluidal comendite occurs near the base of the spectacular plug, Big Nellie, the central part of which consists of coarser material. The presence of brecciated feldspar and quartz with undulose extinction in some of the plugs suggest these have been intruded as a crystal mush.

A trachytic rock occurring at the western base of South Brother in the Lorne Basin shows marked variation in grain-size and degree of alteration over short distances, and chemically has more affinity with the Comboyne alkaline series than with the Lorne Basin granitic rocks with which it is associated. The texture is mainly trachytic, but shows great variation consistent with the rock being a high-level intrusive.

Age and Order of Emplacement of Alkaline Rocks

K/Ar dating of the Mt. Oliver comendite (McDougall and Wilkinson, 1967) gives an age of 15.7 ± 0.7 m.y., and similar dating of the fine grained trachytes by Wellman (1971) gives an age of 16 m.y. The similar ages of these two alkaline rock types

indicate they belong to the same volcanic episode, and the close proximity of other mildly alkaline near-saturated and saturated rocks suggests a Middle Miocene age for the Comboyne alkaline series as a whole.

This indicates that the Comboyne alkaline association is slightly younger than the closely comparable Nandewar Mountain province for which K/Ar dating gives an age of 17 to 20 m.y. (Abbott, 1969). The mildly alkaline rocks forming the Nandewar Mountains are chemically similar to those in the Comboyne area. However, the presence of rare alkali olivine basalt contrasts with the Comboyne association where hawaiite is the most basic rock found.

The actual order of emplacement of the Comboyne alkaline rocks can only be inferred from field relationships and geochemistry. The abundance of megacrysts and cumulates precipitated at elevated pressure in several of the hawaiites indicates that the melt rose rapidly from depth and marked fractionation did not subsequently occur prior to consolidation. The fine grain size and low stratigraphic position of the Comboyne hawaiites is consistent with their being erupted as lava flows prior to the more evolved lavas. The observed increase in stratigraphic level with degree of differentiation points strongly to volcanicity proceeding from basic hawaiite towards more evolved types.

The occurrence of isolated areas of alkaline volcanic rocks as topographic highs in the Lorne Basin suggests that these lavas may have initially been more extensive and that subsequent erosion has removed much of the volcanic material. Remnants of alkaline lavas also occur on highs to the west of the Comboyne Plateau and the escarpment of the plateau is cut by deep valleys, again

suggesting that erosion may have destroyed some of the original volcanic material. Small remnant spines of comendite in the Lansdowne Valley also indicate that erosion may have completely destroyed similar plugs.

Despite the suggestion that erosion has removed a proportion of the original volcanic material, the large areal extent and the limited total depth of the Comboyne lavas indicate that flows of restricted size were emitted from a number of volcanic centres and that at no time was there development of a large central-volcano type of complex.

1.3.2 Tholeiitic Rocks

Only two samples of tholeiitic rocks have been located, and only one of these crops out in-situ. Both occurrences are on the Comboyne Plateau (Grid References 546.8/083.3 and 548.2/086.2), in close association with alkaline rocks.

Age of Tholeiitic Rocks

By association with the alkaline rocks the tholeiitic andesite (MU6577) is considered to be of Tertiary age. The extreme alteration of the second rock suggests it might predate the Comboyne alkaline series.

1.3.3 Calc-Alkaline Rocks.

Calc-alkaline igneous activity centred on the Lorne Basin is now represented by laccoliths, sills, dykes and flows. There are also extensive occurrences of calc-alkaline rocks to the north of the Lorne Basin and limited outcrops of rhyolite occur to the south in the vicinity of Wingham.

The most prominent features in the eastern part of the Lorne Basin are the granitic laccoliths of North, Middle and South Brother, which rise steeply from the coastal plain to heights between 460 - 550 metres. Although these high-level intrusive bodies are in close proximity, they exhibit variations in chemical composition and texture which could partly result from differences in size and emplacement level. Rocks comprising low hills to the west and north are texturally and chemically similar to North and Middle Brother, indicating they belong to the same igneous episode. Middle Brother, South Brother and Juhle Mountain granites are intrusive into the Lorne Basin Triassic sedimentary rocks, whereas similar rocks in the Lorne Forest area intrude the Carboniferous Byabbarra Beds. Quaternary sediments tend to mask the intrusive relationships of the North Brother microgranite, although contact with Carboniferous sedimentary rocks occurs in a number of localities.

Middle Brother consists primarily of granite and granodiorite. Low down on Middle Brother angular blocks of the finer granodiorite can be seen enclosed in granite, and at higher elevations granodiorite becomes the dominant phase and granite mesh-veins and forms dykes in the granodiorite. Granophyric granite forms selvages between these dykes and the host, and also occurs as more extensive outcrops at Middle Brother and Juhle Mountain. Mixed rocks of granodiorite and granite also outcrop to the east of Mt. Comboyne. Field relationships suggest that the fine grained granodiorites may represent an earlier intrusion, which was subsequently stopped by the later granitic magma. Mesh-veining of the granodiorite indicates that some forceful injection of the granitic liquid occurred.

Two phases of the North Brother porphyritic microgranite can be recognised. The finer grain size of one gives it a darker, more glassy appearance than the slightly coarser phase. Contraction jointing, apparently parallel to the contact rocks, occurs in both types but is more closely spaced in the finer phase.

The South Brother (?)riebeckite microgranite is chemically distinct from the other intrusive rocks in the Lorne Basin. Marked secondary alteration of this small laccolith makes direct comparison difficult, although the near absence of ferromagnesian silicates in the South Brother microgranite, along with higher SiO_2 and K_2O and lower FeO (total iron) and Al_2O_3 than in the associated granites suggest it crystallised from a more evolved melt. The close proximity and similarity in form of the Three Brothers suggest that they are contemporaneous and derive from a common parental melt. A greater degree of oxidation in the South Brother microgranite may indicate the magma was enriched in volatiles relative to the other Lorne Basin granitic magmas, and this is supported by the presence of (?)riebeckite in the South Brother granite (Ernst, 1962).

Porphyritic and aphyric rhyolite forms, dykes and flows throughout the Lorne Basin and in the vicinity of Wingham. These rhyolites overlie and intrude the Triassic Camden Haven Beds and are considered to belong to the same magmatic cycle as the associated granitic rocks. Despite marked variation in degree of alteration, the calc-alkaline rhyolites show strong similarities over a wide area, and rhyolite flows in the western part of the Lorne Basin which underlie alkaline mugearites are closely comparable to those forming flows at Diamond Head. A breccia-pipe at Diamond Head may have been the source of the Diamond Head

rhyolites.

A 1.5 m wide pitchstone sill is mineralogically and chemically similar to the North Brother porphyritic microgranite and intrudes the Carboniferous Byabbarra Beds just east of Kew. A second pitchstone occurring in an abandoned quarry near Moorland also has chemical affinities with the Lorne Basin granitic rocks.

Three diorite bodies outcrop in and to the north of the Lorne Basin. The Diamond Head quartz diorite forms a plug approximately 30 m in diameter, and is in direct contact with rhyolite. Two small intrusions of diorite occur to the west of Herons Creek in the Lorne Forest area, the largest being approximately 170 x 20 m in area.

Two distinct rocks which have undergone marked secondary alteration are closely associated with the Diamond Head rhyolites. The more basic intrudes the Triassic sedimentary rocks and has large phenocrysts of plagioclase up to 8 cm long. The second type outcrops as a massive strongly altered dacitic rock which is surrounded by rhyolite and occurs near the southern end of Diamond Head. There is no conclusive evidence pointing to this rock predating or postdating the rhyolite.

Age and Order of Emplacement of Calc-Alkaline Rocks

Contact relationships of the calc-alkaline rocks with the Lorne Basin Triassic sedimentary rocks indicate that the igneous rocks postdate the sedimentary rocks. This is confirmed by K/Ar dating of hornblende from the Middle Brother granodiorite which gives an age of 204 m.y.¹. The close chemical affinity of the

1. K/Ar dating done by Dr. I McDougall, Research School of

majority of the granitic calc-alkaline rocks in the Lorne Basin area indicates that they are of a similar age. Rhyolite flows overlying the Triassic Camden Haven Beds and sills within these beds conclusively show that rhyolite postdates the sediments and Stewart (1953) reports indurated Triassic mudstones in contact with the Middle Brother igneous rocks.

The general absence of any contact relationships between the different igneous types restricts any interpretation on order of emplacement based on field relationships. As already noted, there is field evidence indicating the Middle Brother granodiorite crystallized ahead of the associated granite. Similarly, micro-granite veins in the Diamond Head diorite indicate the diorite predates the granitic rocks. The quartz diorite is surrounded by rhyolite, and the absence of any firm evidence of contact metamorphism of the rhyolites suggests that they postdate the diorite. Diorites to the north of the Lorne Basin intrude Carboniferous sedimentary rocks and on this basis can only be placed as post-Carboniferous.

The true extent of rhyolite volcanism in the Lorne Basin area has probably been destroyed by subsequent erosion. The common occurrence of isolated outcrops of rhyolite suggest that the volume of such material was originally greater and must have emanated from a number of volcanic vents.

1.4 Tectonics

The Lorne Basin Triassic Camden Haven Group unconformably overlies Palaeozoic strata which underwent deformation during the Hunter-Bowen Orogeny. Around the western escarpment of the Basin the Triassic beds dip inwards at angles from 10° to 15° (Stewart, 1953; Pratt and Herbert, 1973), whereas similar rocks

dip approximately 15° to the northeast at Diamond Head. Pratt and Herbert suggest the inward dipping beds in the western part of the Lorne Basin reflect the original depositional slopes and that this is supported by the thinning of the Triassic beds eastward. However, whereas the basal conglomerate reaches an elevation in excess of 300 m in the western margin of the Lorne Basin and on the Comboyne Plateau, it occurs at sea-level at Grants Head and Point Perpendicular. This has been interpreted by Voisey (1939a) and Stewart (1953) as indicating that the Lorne Basin has undergone an eastward tilt, although Stewart suggests the possibility of the present position of the sedimentary beds in the eastern part of the Basin being the result of faulting. The occurrence of high-level intrusives in close proximity with contemporaneous rhyolitic flows, but at elevations approximately 500 m higher, indicates that local faulting has occurred. It is envisaged that high-level emplacement of granitic laccoliths (probably to within several hundred metres of the surface), would be accompanied by doming of the country rocks resulting in relative downfaulting of the surrounding sedimentary rocks. To the east of the Three Brothers evidence for such faulting would be largely obscured by Quaternary alluvium deposits. Steeply dipping Triassic strata around the rim of Mt. Juhle has been interpreted by Pratt and Herbert (1973) to indicate the semi-diapiric intrusion of the Juhle Mountain granite. Similar disruption of the Camden Haven Group at Ross Glen to the north of Middle Brother granite also indicates some disruption of the country rock on emplacement of the high-level igneous material.

Post Triassic faulting is evidenced in a number of areas in the Lorne Basin and Stewart (1953) suggested this movement took place in the Tertiary. The tendency for alkaline plugs to be grouped in the Lansdowne Valley and the western rim of the Lorne Basin suggests that there may have been some tectonic control during their emplacement.

CHAPTER 2 PETROGRAPHY

2.1 Introduction

Petrographic descriptions are given of all rock types analysed in this study and the petrography of the alkaline rocks is summarised in Table 2.1. The rocks analysed are considered to be representative of the igneous rocks in the Lorne Basin - Comboyne Plateau area, and descriptions of the different rock types are grouped according to the major classification categories (see Chapter 4) of Alkaline, Tholeiitic and Calc-Alkaline rocks. Identification numbers refer to catalogue numbers for rocks held in the School of Earth Sciences, Macquarie University, North Ryde, New South Wales (Appendix I).

2.2 Alkaline Rocks

Hawaiite. Hawaiites contain up to twenty percent phenocrysts and/or megacryst/cumulates of plagioclase, olivine, clinopyroxene, orthopyroxene, Fe-Ti oxides and apatite (e.g. MU6498, MU6597). Plagioclase is the major phenocryst-megacryst phase and occurs as euhedral, as well as strongly corroded and sieved crystals (Plates 9 and 10). Rims of strongly zoned, untwinned feldspar are common. Olivine occurs as subhedral and anhedral phenocrysts-megacrysts, sometimes completely replaced by iddingsite, haematite, chlorite, calcite and siderite. Large subhedral crystals of clinopyroxene up to 1.0 cm long commonly have a broad sieved reaction rim (Plate 5b). Some small crystals are strongly corroded and sieved throughout (Plate 6a). Subhedral clinopyroxenes with no evidence of sieving range up to 2 mm in size. Relict cores of orthopyroxene are enclosed within granular reaction rims of olivine, clinopyroxene

Fig. 2.1 Summary of petrography of the Comboyne
alkaline rocks.

Hawaiite	Mugearite	Benmoreite Mafic Trachyte	Anorthoclase Trachyte	Comendite	
—•—x—•—x—	—x—x—x—x—	x x x x			Olivine
—•—x—•—x—	—•—x—•—x—	•••••	•••••		Clinopyroxene
x x x x	x x x x				Orthopyroxene
		x x x x		•••••	Aegirine Augite
				•••••	Brown Amphibole
				•••••	Riebeckite/ Arfvedsonite
—•—x—•—x—	—•—x—•—x—	—x—•—x—	•••••		Aenigmatite
—•—x—•—x—	—•—x—•—x—	—x—x—	•••••		Fe-Ti Oxides
•••••	•••••	—x—x—	•••••	•••••	Plagioclase
x • x • x •	x • x • x • x •	• x • x • x •	•••••	•••••	Alkali Feldspar
				•••••	Apatite

••••••••• Groundmass

----- Phenocrysts

x x x x x x x Corroded Megacrysts / cumulates

—•—x—•—x— Phenocrysts, Groundmass, Megacrysts / Cumulates

and haematite and, in some instances, the orthopyroxene has been completely replaced by these phases (Plate 1). Coarse-grained cumulates of 1) olivine-clinopyroxene-ilmenite, 2) olivine-orthopyroxene-plagioclase, 3) clinopyroxene-orthopyroxene-plagioclase, 4) clinopyroxene-orthopyroxene, are common in several of the hawaiites (Plates 2 and 3).

Megacryst and cumulate ilmenites are strongly corroded and are rimmed by and have lamellae of magnetite (Plate 14a). Rare corroded megacrysts of magnetite have lamellae of and are rimmed by ilmenite. Corroded crystals of apatite up to 1.5 mm long (Plate 6) have aligned elongate inclusions of a second phase, and occur as isolated grains and also associated with, or as inclusion in, clinopyroxene, olivine, plagioclase and magnetite. The hawaiite groundmass consists of plagioclase, alkali-feldspar, olivine, clinopyroxene, magnetite, ilmenite, colourless acicular apatite and rare pyrrhotite. Grain size varies from approximately 0.01 to 0.15 mm.

Mugearite. Mugearites are not common in the Comboyne-Lorne Basin area and generally are fine grained and aphyric. Phenocrysts-megacrysts of plagioclase, alkali-feldspar, olivine, clinopyroxene, striated apatite and Fe-Ti oxides are rare, although one specimen (MU6578) has abundant phenocryst and megacryst/cumulate phases including olivine, clinopyroxene (Plate 5a), orthopyroxene, plagioclase and Fe-Ti oxides. The groundmass of the mugearites consists of plagioclase, alkali-feldspar, clinopyroxene and Fe-Ti oxides, is generally strongly oxidised and sometimes includes brown interstitial glass (MU6578). The grain size varies from approximately 0.02 mm to 0.08 mm.

Benmoreite. The Comboyne benmoreites are relatively coarse-grained and contain phenocrysts and (?) megacrysts/coarse aggregates (up to 2 cm across) of anorthoclase, rare plagioclase, clinopyroxene, olivine, amphibole, ilmenite, magnetite and striated apatite (Plate 13b). The groundmass consists of equidimensional anorthoclase, pale-green clinopyroxene, ilmenite, magnetite and clear acicular apatite. Groundmass grain-size varies up to 0.1 mm.

Mafic Trachyte. Mafic trachytes (e.g. MU6501) can generally be distinguished from benmoreites on a textural basis as they tend to be aphyric with a fine-grained trachytic texture. Rare phenocrysts include anorthoclase, plagioclase, olivine and intergrowths of Fe-Ti oxides. The major groundmass phases are alkali-feldspar, green clinopyroxene, and finely disseminated magnetite and ilmenite. Oxidation of Fe-Ti oxides is common and oxidised orange interstitial glass is present in some specimens. Apatite occurs as striated subhedral crystals and as clear laths in the groundmass. Grain-size varies up to 0.1 mm. Secondary minerals include haematite, chlorite, calcite, siderite and iddingsite.

A quartz-bearing trachyte (MU6667) occurring at the western base of South Brother is distinct from the Comboyne benmoreites and mafic trachytes. This rock exhibits variable degrees of deuteric alteration over short distances. Plagioclase phenocrysts have marked normal and oscillatory zoning, with inclusion trails of pyroxenes and opaques near the rim. Olivine and pale-green clinopyroxene grains range up to 0.75 mm in size. Groundmass phases are plagioclase, alkali-feldspar, clinopyroxene, quartz, Fe-Ti oxides and acicular apatite. Secondary minerals include deep green amphibole, chlorite, epidote, haematite and calcite.

Anorthoclase Trachytes. Anorthoclase trachytes are restricted to plugs and dome-like intrusions on the Comboyne Plateau (Mt. Gibraltar - MU6630, Mt. Bulli - MU6635, Mt. Kophi - MU6627). These trachytes are considerably more leucocratic than the mafic trachytes, and phenocrysts include anorthoclase, minor plagioclase, green clinopyroxene, and rare aggregates of secondary minerals probably representing completely replaced olivine. Aggregates of anorthoclase, with minor plagioclase and clinopyroxene, range up to 5 mm across. The trachytic groundmass consists primarily of alkali-feldspar (up to 0.2 mm long), clinopyroxene, magnetite, ilmenite and clear apatite laths. Secondary minerals include chlorite and haematite. Mt. Bulli anorthoclase trachyte is essentially devoid of phenocrysts, but is chemically similar to the porphyritic anorthoclase trachytes forming Mt. Kophi and Mt. Gibraltar.

Comendite. Porphyritic comendite commonly forms plugs (e.g. MU6647, MU6649, MU6652) whereas comendites on the Comboyne Plateau generally have an even-grained fluidal texture (MU6636, MU6669, MU6573). The porphyritic comendites have phenocrysts of euhedral alkali-feldspar and rounded quartz, with a groundmass of alkali-feldspar, quartz, riebeckite/arfvedsonite, \pm aegirine, \pm aenigmatite and clear acicular apatite. Phenocrysts of plagioclase, riebeckite/arfvedsonite and aegirine occur in some plugs. The Mt. Coxcomb comendite (MU6645, MU6646) contains coarse-grained aggregates of anorthoclase, plagioclase (rimmed by alkali-feldspar), quartz, riebeckite/arfvedsonite, aegirine and Fe-Ti oxides, and similar fragments occur in a volcanic breccia to the northwest of Mt. Coxcomb. The Fe-Ti oxides are enclosed in a narrow rim of riebeckite/arfvedsonite and tend to be strongly corroded. The

non-porphyritic comendites have a similar mineralogy to the porphyritic varieties and a grain size of approximately 0.1 mm. Porphyritic comendites occurring at the southern escarpment of the Comboyne Plateau and at The Monastery to the south of the plateau are distinct from the majority of the comendites. These rocks have phenocrysts and crystal aggregates of euhedral and subhedral alkali-feldspar. Mafic minerals are less abundant than in the other comendites and riebeckite/arfvedsonite occurs as a relatively sparse groundmass phase (MU6658). An alkaline pitchstone (MU6636) in the Lansdowne Valley is chemically similar to the more mafic comendite, but contains soda-hedenbergite rather than aegirine. Alkali-feldspar is the only other phenocryst phase.

2.3 Tholeiitic Rocks

Tholeiitic Andesite. Two tholeiitic andesites have been identified on the Comboyne Plateau. A porphyritic type (MU6577) has phenocrysts of euhedral, zoned plagioclase, olivine, and clinopyroxene (some with sieved reaction rims). Additionally, it has orthopyroxene and olivine megacrysts and crystal aggregates (upto 1 cm across) comprising primarily of orthopyroxene, but with minor olivine. The orthopyroxene has distinct rims of twinned clinopyroxene (Plate 7). Gabbroic aggregates up to 2 cm across are also present and consist of plagioclase, clinopyroxene, olivine, ilmenite and interstitial glass (Plate 8). The groundmass consists of plagioclase laths, equidimensional augite, pigeonite, clear acicular apatite, ilmenite, magnetite and brown oxidised glass. The second (?) tholeiitic andesite (MU6568) is a quartz-bearing, non-porphyritic, fairly coarse-grained rock. Plagioclase and clinopyroxene are the main phases,

with less abundant quartz, chlorite, Fe-Ti oxides and epidote. The extreme secondary alteration of this rock makes any analyses of very limited value and it is therefore not considered further.

2.4 Calc-Alkaline Rocks

Microgranite. The North Brother porphyritic microgranite consists of two texturally distinct phases which show only limited mineralogical and chemical variation. A relatively coarse grained phase with typical microgranitic texture (e.g. MU6676) contrasts with a finer spherulitic phase (e.g. MU6673). Orthopyroxene and clinopyroxene occur in both, but are more abundant in the finer-grained phase. Minor pyroxene commonly occurs with plagioclase in crystal aggregates which range up to 4mm across and total 10-15% of the rock. Plagioclase exhibits strong sieving and alteration to white mica, clay minerals, calcite, epidote and chlorite. Groundmass feldspar is also strongly altered and is micrographically intergrown with clear quartz. Mafic xenoliths occur sporadically in the North Brother microgranite. The mineralogy of these xenoliths is similar, but the proportion of mafic minerals varies between xenoliths. Plagioclase and acicular clinopyroxene, almost totally replaced by amphibole and chlorite, are the major phases. Minor quartz, ⁺ alkali-feldspar are also present. Accessory and secondary minerals include apatite, zircon, Fe-Ti oxides, calcite, amphibole, chlorite and white mica.

Perthitic alkali-feldspar is the only phenocryst phase in the strongly altered South Brother porphyritic (?) riebeckite microgranite (MU6661, MU6663) and subhedral grains commonly occur in glomeroporphyritic aggregates up to 1.5 mm across. The ground-

mass consists of slightly spherulitic feldspar and quartz with rare 'mossy' grains of (?)riebeckite. Finely disseminated opaques are unevenly distributed throughout the groundmass. Secondary minerals include white mica, chlorite, haematite and clay minerals.

Graphic Granite. The graphic granites exhibit a wide variation in grain size and include porphyritic and non-porphyritic types. Strongly altered plagioclase phenocrysts total 5-15% of these rocks and range up to 2.5 mm in size. Groundmass feldspar exhibits similar strong alteration, but relict multiple twinning can be identified in some instances. Clear quartz is graphically intergrown with alkali-feldspar. Mafic minerals are rare and consist of microphenocrysts of hornblende. Accessory and secondary minerals include apatite, Fe-Ti oxides, zircon, biotite, epidote, chlorite and white mica. Even-grained granites have a similar mineralogy, but mafic minerals tend to be slightly more abundant.

Granodiorite. The major mineral phase in the granodiorites is plagioclase which occurs as subhedral laths exhibiting strong normal zoning. Quartz content varies from approximately 3 - 10%. Clinopyroxene (commonly replaced by green amphibole), orthopyroxene, brown amphibole, biotite, Fe-Ti oxides, apatite, zircon, chlorite, epidote and alkali-feldspar are minor phases. The rough alignment of plagioclase results in a relatively coarse-grained fluidal appearance.

Granophyre. Granophyres exhibit a wide variation in grain-size, texture and composition, and occur in a number of areas at Middle Brother and Juhle Mountain. They occur as isolated

boulders and as narrow selvages at contacts between graphic granite dykes and the host granodiorite. Euhedral, strongly zoned plagioclase is enclosed within broad areas of granophyric intergrowths of quartz and alkali-feldspar, with minor biotite and Fe-Ti oxides.

Pitchstones. Two calc-alkaline rhyolitic pitchstone sills have been located in the Lorne Basin. One of the pitchstones (MU6692) has phenocrysts and crystal aggregates (up to 3 mm across) of plagioclase, \pm alkali-feldspar, and minor clinopyroxene, orthopyroxene and Fe-Ti oxides (Plate 11b). The crystal aggregates include even-grained and porphyritic varieties. The second pitchstone (MU6690) has strongly corroded and sieved phenocrysts and aggregates of plagioclase (Plate 12a), \pm alkali-feldspar, with microphenocrysts of clinopyroxene, orthopyroxene and Fe-Ti oxides. Microlites grow out from the phenocrysts into the glassy matrix.

Rhyolite. The porphyritic and non-porphyritic rhyolites have generally undergone marked deuteric alteration. Phenocrysts consist of feldspar (some with relict multiple twinning) and, more rarely, quartz, orthopyroxene and clinopyroxene. The groundmass consists of alkali-feldspar, quartz, white mica, chlorite and finely disseminated Fe-Ti oxides. Flow banding is present in some specimens (MU6704).

Dacites. Strongly altered porphyritic dacites (MU6687, MU6686) have phenocrysts of plagioclase and cubic and skeletal pyrite up to 0.7 mm across. Groundmass phases include feldspar, quartz, and abundant secondary white mica, chlorite and minor biotite. The groundmass texture varies from spherulitic to

trachytic. A second dacitic rock type is more mafic with abundant groundmass amphibole and chlorite.

Basaltic Andesite. Basaltic andesite (MU6689) occurring at Diamond Head contains abundant euhedral plagioclase phenocrysts which vary greatly in size, but range up to 8 cm in length. Apart from plagioclase, the groundmass has undergone such extreme alteration to white mica, chlorite and Fe-Ti oxides that identification of primary minerals is impossible.

Diorite. Diorites have strongly zoned plagioclase, euhedral to subhedral clinopyroxene (commonly replaced by amphibole and chlorite), Fe-Ti oxides and minor apatite and zircon. Quartz is graphically intergrown with feldspar in the Diamond Head quartz diorite (MU6684).

CHAPTER 3 MINERALOGY

3.1 Introduction

All mineral analyses have been determined by a TPD electron microprobe following the method outlined by Reed and Ware (1973). This has allowed the detailed investigation of the major mineral phases present in the Comboyne alkaline rocks, with particular emphasis on mineral compositional trends within and between the various rock types.

The occurrence in the Comboyne hawaiites of at least two generations of "phenocryst" phases which can be distinguished on both textural and chemical criteria, leads to the suggestion that one of these generations crystallised at elevated pressures prior to the extrusion of the host melt to surface levels. In this study the terms "megacryst" and "cumulate" are used to describe phases considered to have crystallised at elevated pressures, whereas the term "phenocryst" is used when a crystal is considered to have precipitated in a low-pressure environment.

3.2 Olivine

3.2.1 Introduction

Olivine is a common megacryst/cumulate, phenocryst and micro-phenocryst phase in the hawaiites of the Comboyne Plateau - Lorne Basin alkaline province. It is less common in the mugearites and benmoreites and is rare in trachytes. The hawaiites also contain groundmass olivine. Both phenocryst and microphenocryst olivine is present in the tholeiitic andesite (MU6577), and green phyllosilicates possibly pseudomorphing olivine occur in the

Lorne Forest diorites (MU6666, MU6667).

3.3.2 Alkaline Rocks

3.2.2.1 Textural Relationships

At least three generations of olivine occur in the hawaiites.

These consist of:

1. a) Megacrysts
- b) Cumulates in which olivine is associated with orthopyroxene, ilmenite, clinopyroxene, plagioclase.
2. Phenocrysts
3. a) Groundmass olivine
- b) Olivine forming granular reaction rims around orthopyroxene and, less commonly, clinopyroxene.

Megacryst and cumulate olivines are devoid of zoning and deformation lamellae, show some evidence of resorption where in contact with the melt, and tend to be clearer and to have fewer inclusions than the associated olivine phenocrysts. Olivine phenocrysts in the hawaiites are euhedral to subhedral and show compositional zoning. Inclusions are not abundant and there is variable alteration of olivine to secondary minerals. Some crystals appear strongly sieved, but are chemically similar to the unsieved olivine phenocrysts. Groundmass olivine is less common than other major groundmass phases and exhibits more pronounced deuteric alteration than the megacryst/cumulate and phenocryst olivines. Olivine forms granular reaction rims (in association with clinopyroxene) around orthopyroxene and, to a lesser extent, clinopyroxene megacrysts (Plates 1a and 5a,b). In many instances these granular aggregates of olivine and clinopyroxene have completely replaced the orthopyroxene megacrysts and,

in turn, have undergone oxidation to form haematite,
+ iddingsite (Plate 1b).

In the near-aphyric mugearites olivine occurs as rare micro-phenocrysts and phenocrysts. These olivines are typically subhedral (a euhedral form possibly being masked by alteration to grain boundaries), show only limited zoning, and are usually more strongly oxidised than olivines in the hawaiites. A porphyritic mugearite contains rare olivine megacrysts. Olivine phenocrysts have inclusions of both striated and acicular apatite and Fe-Ti oxides.

Benmoreites contain rare subhedral (?) megacrysts of olivine up to 4 mm across, which exhibit minor zoning and inclusions of striated apatite. Olivine also occurs in crystal aggregates which contain anorthoclase, ilmenite, magnetite, amphibole and striated apatite in addition to olivine (Plate 13b).

Only rare relict olivine occurring as cores in pseudomorphs after olivine have been observed in the mafic and anorthoclase trachytes. In these rocks olivine is commonly completely replaced by iddingsite and haematite. The South Brother trachyte has subhedral and anhedral phenocrysts of olivine upto 1.0 mm across, which in some specimens is largely altered to iddingsite and green phyllosilicates. These phenocrysts total approximately 2% and are sometimes rimmed by granular green clinopyroxene similar to the phenocryst and groundmass pyroxene.

Alteration. Haggerty and Baker (1967) and Baker and Haggerty (1967) determined experimentally that olivines cooled under oxidising conditions exsolve iron oxides at high temperatures which, in turn, breakdown at low temperatures to essentially chloritic minerals. Olivines cooled under non-oxidising

conditions were found to alter at intermediate temperatures to complex assemblages of chlorite and phyllosilicates. In contrast, alteration under oxidising conditions at low temperatures (probably below 140° C) produces iddingsite which, during post-deuteric alteration, forms strongly pleochroic, highly ordered varieties which eventually breakdown to a green phyllosilicate assemblage.

Deuteric alteration of olivines in the Comboyne hawaiites has mainly resulted in the formation of phyllosilicates with only minor development of iddingsite, suggesting that alteration occurred in a non-oxidising environment at intermediate temperatures. The development of iddingsite and haematite in the majority of other alkaline rock types indicates that alteration occurred at low temperatures under oxidising conditions.

3.2.2.2 Analytical Results

Reference to Tables 3.1 and 4.1 shows that olivines in the Comboyne hawaiites exhibit a wide range in composition (Fo_{74-48}), with rather limited variation in host rock composition. However, if the proposed high-pressure and rim phases are not included, variation in olivine composition is more restricted (Fo_{59-48}).

The presence of olivine "phenocrysts" which fall into two distinct compositional groups in single hand-specimens suggests that not all "phenocrysts" crystallised in the same environment. This is also indicated by the occurrence of olivine in cumulates associated with orthopyroxene, clinopyroxene and plagioclase (Plate 2a), and the close similarity in form (page 25) of these to olivine megacrysts of high pressure origin described from other locations (Binns et al., 1970; Wass, 1971; Irving, 1974). The

TABLE 3.1

MICROPROBE ANALYSES OF OLIVINES IN ALKALINE ROCKS

Rock Type:	Hawaiiite																
Sample No:	6498	6597	6571	6572	6498	6498	6571	6570	6597	6572	6572	6597	6498	6597	6498	6597	6597
Phase:	C	C	M	M	P	P	P	P	S	S	G	G	G	Ropx	Ropx	Rcpix	Rcpix
SiO ₂	36.35	37.33	38.45	38.32	36.47	35.03	36.18	36.49	34.89	35.87	35.53	34.56	34.99	36.19	36.39	34.96	35.28
TiO ₂	-	-	-	-	.06	-	-	.09	-	-	.11	-	.18	-	.15	-	-
Al ₂ O ₃	-	.22	.73	-	.50	.11	.65	.51	.27	.15	-	.25	.16	.24	.34	.15	.18
Cr ₂ O ₃	.11	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
FeO	30.90	25.21	22.87	24.58	34.83	38.24	35.07	33.97	30.19	34.72	35.27	42.05	36.26	31.88	33.21	38.07	37.09
MnO	.34	.24	.19	.20	.55	.37	.37	.48	.69	.42	.60	.64	.65	.43	.55	.52	.47
MgO	32.05	36.04	36.54	36.88	26.71	25.36	26.63	27.29	24.76	27.91	27.59	21.77	25.36	29.24	28.77	24.53	25.86
CaO	.16	.23	.25	.24	.45	.23	.39	.41	.40	.39	.35	.33	.40	.30	.38	.28	.33
Na ₂ O	-	-	-	-	.23	-	-	.20	.23	-	.28	.26	-	-	-	-	.17
K ₂ O	-	-	-	-	-	-	.20	.20	-	-	-	-	-	-	-	-	-
Total	100.00	99.27	99.03	100.22	99.02	99.34	99.49	99.64	99.43	99.46	99.73	99.96	98.00	98.28	99.79	98.51	99.38
Cation Content (4 oxygen atoms)																	
Si	.9895	.9936	1.0105	1.0044	1.0126	.9968	1.0102	1.0124	.9944	1.0005	.9940	.9994	1.0019	1.0073	1.0029	1.0037	.9984
Al	-	.0070	.0225	-	.0163	.0035	.0216	.0166	.0091	.0050	-	.0086	.0054	.0077	.0110	.0050	.0060
Ti	-	-	-	-	.0012	-	-	.0019	-	-	.0022	-	.0040	-	.0033	-	-
Cr	.0024	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Fe	.7051	.5610	.5027	.5390	.8093	.9100	.8191	.7881	.9101	.8097	.8250	1.0168	.8689	.7420	.7653	.9140	.8777
Mg	1.2999	1.4293	1.4312	1.4411	1.1053	1.0752	1.1078	1.1286	1.0516	1.1602	1.1504	.9385	1.0829	1.2129	1.1817	1.0496	1.0905
Mn	.0077	.0055	.0041	.0044	.0128	.0088	.0085	.0111	.0165	.0100	.0143	.0101	.0157	.0100	.0128	.0127	.0112
Ca	.0046	.0065	.0071	.0067	.0134	.0071	.0115	.0121	.0120	.0116	.0103	.0156	.0125	.0088	.0113	.0087	.0100
Na	-	-	-	-	.0121	-	-	.0106	.0123	-	.0148	.0144	-	-	-	-	.0095
K	-	-	-	-	-	-	.0012	.0012	-	-	-	-	-	-	-	-	-
Total	3.0092	3.0029	2.9781	2.9956	2.9830	3.0014	2.9755	2.9826	3.0060	2.9970	3.0110	3.0034	2.9913	2.9887	2.9683	2.9937	3.0033
$\frac{100\text{Mg}}{\text{Mg} + \text{Fe}}$ 2+	64.8	71.7	74.0	72.8	57.7	54.2	57.5	58.9	53.6	58.9	58.2	48.0	55.5	62.0	60.7	53.5	55.4
K _D	.36	.38	.35	.29	.62	.72	.74	.54	.85	.54	.55	1.06	.68	.60	.55	.85	.79

Total iron as FeO

C = Cumulate

G = Groundmass

M = Megacryst

Ropx = Rim on orthopyroxene

P = Phenocryst

Rcpix = Rim on clinopyroxene

S = Sieved grain

overlap in composition of megacryst and cumulate olivine (Fo_{74-65}) indicates that some megacrysts could result from the disruption of the associated cumulates.

The coexisting olivine phenocrysts have a considerably lower range of forsteritic component (Fo_{59-54}) than the megacryst/cumulate olivine. Strongly sieved crystals with the same composition as associated unsieved phenocrysts are also present. Drever and Johnston (1957) argue convincingly that much sieving and corrosion attributed to magmatic resorption is actually the product of skeletal growth, and the similarity in composition between sieved and unsieved phenocrysts in the Comboyne hawaiites supports this view. However, some cumulate olivines have broad, strongly sieved rims which are compositionally comparable to unsieved and sieved "phenocrysts". Textural features of olivines in the hawaiites indicate that sieving has resulted from both skeletal growth and magmatic resorption. Where sieving is the result of skeletal growth, the enclosed groundmass is similar to that throughout the rock. Where sieving has resulted from magmatic resorption, the resultant holes are infilled with secondary minerals.

It is anomalous that whereas some olivine megacrysts show only limited evidence of reaction with the melt, other grains appear to have undergone pronounced resorption. A possible explanation is that sieving of olivine megacrysts and cumulates has only occurred when they have been in contact with the melt for sufficient time for reaction to take place towards a composition in equilibrium with the evolving melt. When megacrysts have been isolated, either by enclosure within other megacryst

phases or within secondary minerals, there may have been little opportunity for reaction to occur. Simkin and Smith (1970) note that diffusion in slowly cooled olivines results in alteration of the original composition towards one in equilibrium with the host melt. Prolonged contact of olivine megacrysts with a magma during varying pressure and temperature regimes would allow such reaction to occur and the resultant resorption would further facilitate more complete reconstitution of the olivine towards a composition in equilibrium with the melt.

Groundmass olivine (Fo_{58-48}) in hawaiites shows only limited variation in composition from that of phenocrysts and has a similar composition to olivine forming rims around clinopyroxene megacrysts (Fo_{55-54}). In contrast, olivine forming granular reaction rims around orthopyroxene megacrysts has a higher magnesium content (Fo_{62-61}) than the associated phenocryst and groundmass olivines.

Olivine phenocrysts and microphenocrysts in the mugearites range in composition from Fo_{50-43} , whereas the rare olivine megacrysts in the porphyritic mugearite have the composition of Fo_{72} . Olivine (?) phenocrysts in the Comboyne benmoreites have a forsterite value of Fo_{35} and phenocrysts in the South Brother trachyte have a composition of Fo_{30} .

3.2.2.3 Discussion

Although there is no marked compositional difference between the most fayalitic olivine analysed in hawaiite (Fo_{48}) and that of phenocrysts occurring in mugearites (Fo_{50-43}), there is, apart from this one instance, no overlap in olivine composition in the different rock types of the Comboyne alkaline

series, indicating that the melts from which these olivines crystallised represented different magma batches which differentiated in isolation. It is apparent that there is a steady decrease in the forsterite molecule in the more evolved rocks reflecting the control of the host melt composition on that of the precipitating olivine (Table 3.1).

There is limited but consistent variation in CaO content between the proposed medium to high-pressure and the low-pressure olivines. Low-pressure olivines, including phenocryst, ground-mass and rim phases have CaO contents ranging from 0.23% to 0.45%, and averaging 0.36%. In contrast, CaO in olivine megacrysts/cumulates varies from 0.16% to 0.25%, and averages 0.20%. MnO is also consistently higher in the low-pressure olivines relative to megacrysts/cumulates (Table 3.1).

In a detailed study of olivines with diverse crystallisation histories, Simkin and Smith (1970) concluded that there is strong evidence that CaO is controlled to some degree by depth of crystallisation. They also concluded that MnO in olivine is largely dependent on the FeO content and that it is the bulk composition of the host rock rather than the crystallisation environment that is the controlling factor. Although the delineation of high and low-pressure olivines in the Comboyne hawaiites on CaO content is not conclusive, it concurs with observations from other alkaline associations which include both high and low-pressure phases (Simkin and Smith, 1970; Wass, 1971).

A study by Roeder and Emslie (1970) on equilibrium between olivine and basaltic melts indicates that the partitioning of magnesium and iron between olivine and liquid is independent of

temperature and depends only on the magnesium and ferrous iron in the liquid from which olivine crystallises. They determined experimentally a distribution coefficient $(K_D = \frac{X_{FeO}^{ol} \times X_{MgO}^{Liq}}{X_{FeO}^{Liq} \times X_{MgO}^{ol}})$ of 0.30 for several basaltic compositions over a temperature range of 1150-1300° C. and oxygen fugacity of $10^{-6.8}$ to 10^{-12} atm. This contrasts with values determined by Nicholls (1974) who found that olivine distribution coefficients in quartz-normative basaltic and andesitic compositions under hydrous conditions varied from 0.33 to 0.50 and averaged approximately 0.40. If it is first of all assumed that most magnesian olivine megacrysts precipitated as liquidus phases from a melt corresponding to the bulk composition of the hawaiite, then K_D values of 0.29 - 0.38 are obtained, showing greater consistency with Nicholls' work for quartz-normative, hydrous melts, and thus lending support to the validity of the assumption. However, it should be noted that if the olivines are not liquidus phases, but represent equilibrium crystallisation products along with orthopyroxene and clinopyroxene, etc. well below the liquidus, then they will be in equilibrium with a liquid with a higher Fe/Mg ratio than the bulk rock composition. In this case the above assumption would lead to erroneously high K_D values and could perhaps explain the generally higher K_D values for the Comboyne rocks compared with Roeder and Emslie's value of 0.3.

The high proportion of megacryst/cumulate phases in the hawaiites and the possible crystallisation of olivine megacrysts/cumulates from a parent melt of similar composition to that of the host rock, indicate that extrusion from depth to the surface has been rapid, allowing little time for gravity settling of crystal

phases. The occurrence of hawaiiite in a plug-like form (Mount Baldy - MU6498), and the lack of zoning and deformation lamellae in olivine megacrysts and cumulates are consistent with this interpretation. If this conclusion is correct, then the comparatively low $100 \text{ Mg/Mg} + \text{Fe}^{2+}$ ratios (44-50) of the hawaiiites (which contrast with values of 68-73 required for a liquid to be in equilibrium with mantle 'pyrolite' (Green, 1970)), indicate that the melt must have been held at depth prior to the formation of the present megacryst/cumulate assemblage, during which time fractionation of magnesium-rich phases took place.

Phenocryst and groundmass olivines are up to 25% richer in the fayalite component than the coexisting megacryst phases, and, as would be anticipated have K_D values higher than predicted by Nicholls (1974) and Roeder and Emslie (1970). This supports the suggestion that these olivines crystallised from a melt with a greater ferrous iron component than that of the bulk host rock composition.

Roeder and Emslie noted that crystallisation of olivine will only cause absolute iron enrichment in the liquid at oxygen fugacities greater than 10^{-6} atm. at 1200°C . or at temperatures above 1280°C . Under most geological conditions the separation of an iron-poor phase such as plagioclase is necessary to cause enrichment in iron. The predominance of plagioclase phenocrysts-megacrysts and the presence of plagioclase in cumulates indicates that fractionation of feldspar has been prolonged, and iron enrichment of the melt would be expected. This is consistent with trends observed in other alkaline associations where marked increases in the fayalite component in olivines are only apparent in the more evolved rock types where feldspars are prominent phenocryst phases. In contrast, iron enrichment in olivine is

restricted where the crystallisation of ferromagnesian minerals has been dominant (Johnston, 1953; Wilkinson, 1957b; Le Maitre, 1962; Wass, 1971). It is concluded from the above data that crystallisation of feldspar has resulted in iron enrichment in the melt prior to the crystallisation of phenocryst and ground-mass olivine, and that the bulk host rock composition, which includes megacryst and cumulate phases, is not representative of the melt from which these low pressure olivines crystallised.

The occurrence of cumulate phases enables comparison with Green and Ringwood's (1967) experimental study on basalts where liquidus phases were determined for varying compositions and pressures. However, as previously noted $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratios of < 50 for the Comboyne hawaiites indicate that they do not represent a primary melt, and direct comparison with Green and Ringwood's results should be made with caution. Their results indicate that olivine is the liquidus phase for olivine tholeiite, olivine basalt and alkali-olivine basalt at 9 kb, and is followed by orthopyroxene and clinopyroxene in the olivine tholeiite and olivine basalt, and by clinopyroxene alone in the alkali olivine basalt. At 13.5 kb olivine is no longer on the liquidus in the three basalts studied. An experimental study on a hypersthene-normative olivine basalt by Green and Hibberson (1970) showed olivine to be the liquidus phase at 9 kb under dry conditions, whereas at 13.5 kb it was replaced by orthopyroxene and clinopyroxene. Under 'wet' conditions olivine appeared as the liquidus phase at 13.5 kb. Irving (1971) found that olivine is the liquidus phase in a nepheline mugearite up to at least 9 kb, and also noted that an increase in water content increases the pressure at which olivine is stable up to 15.9 kb. Similarly, Nicholls

and Ringwood (1973) found that the addition of water extended the maximum pressure of olivine stability in olivine tholeiites and SiO_2 -saturated tholeiites.

There is considerable evidence to indicate that the Comboyne alkaline rocks crystallised from volatile-poor melts, and that the partial pressure of oxygen was low (e.g. secondary alteration of olivine; clinopyroxene crystallisation trends; early crystallisation of plagioclase). It is concluded, therefore, that experimental data available at this time indicate that the olivine megacrysts/cumulates associated with orthopyroxene and clinopyroxene, crystallised from the host hawaiite in a pressure regime of less than 9 kb. Results from experimental runs on the Mt. Baldy hawaiite undertaken in the present study confirm this conclusion (Chapter 5).

3.2.3 Tholeiitic Rocks

3.2.3.1 Textural Relationships

Olivine in the Comboyne tholeiitic andesite (MUC577) occurs in gabbroic aggregates with plagioclase and clinopyroxene, in cumulates with orthopyroxene (which is rimmed with clinopyroxene), and as phenocrysts and microphenocrysts. The olivine phenocrysts have undergone alteration to grain boundaries resulting in a broad rim of iddingsite and haematite. This has caused a rounding of the olivine crystals, although the original crystals were euhedral. Olivine occurs interstitially in both the gabbroic aggregates and the cumulates (Plate 8a,b).

3.2.3.2 Analytical Results

The composition of olivine in the tholeiitic andesite ranges from Fo_{76-65} (Table 3.2). Olivines occurring in the gabbroic

TABLE 3.2

MICROPROBE ANALYSES OF OLIVINES IN SUB-ALKALINE ROCKS

Rock Type:	Tholeiitic Andesite		
Sample No:	6577	6577	6577
Phase:	M	A	P
SiO ₂	39.50	36.96	37.85
Al ₂ O ₃	.26	.47	.32
FeO	21.67	27.87	29.99
MnO	.13	.23	.27
MgO	38.01	33.04	31.88
CaO	.29	.36	.33
Na ₂ O	-	-	.16
Total	99.86	98.93	100.80

Cation Content (4 oxygen atoms)

Si	1.0217	.9997	1.0169
Al	.0078	.0150	.0101
Fe	.4688	.6304	.6740
Mn	.0029	.0053	.0062
Mg	1.4652	1.3320	1.2568
Ca	.0079	.0103	.0094
Na	-	-	.0083
Total	2.9743	2.9927	2.9817

$\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{2+}}$	75.8	67.9	65.1
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K _D	.33	.49	.55
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Total iron as FeO

M = Megacryst

A = Gabbroic aggregate

P = Phenocryst

inclusions and some of the phenocrysts have a similar composition (Fo_{68-65}), indicating that these phenocrysts either formed at the same time as the olivine in the inclusions, or result from the disruption of these gabbroic inclusions. The remainder of the olivine crystals are considerably richer in magnesium (Fo_{76}).

3.2.3.3 Discussion

The olivine in gabbroic inclusions and the (?)phenocrysts with lower forsterite content have a higher FeO:MgO ratio than is predicted by Roeder and Emslie's (1970) and Nicholls' (1974) data (Table 3.2) and represent crystallisation products well below the liquidus. The more Mg-rich (?)megacrysts would appear to have crystallised from a melt with the same bulk composition as the host rock. The presence of two generations of olivine "phenocrysts" suggests that there has been two periods of olivine crystallisation, or alternatively, the gabbroic inclusions are xenoliths accidentally picked up by the melt prior to eruption.

3.3 Pyroxenes

3.3.1 Introduction

Clinopyroxene phenocrysts are present in all the Comboyne alkaline rocks and also occur rarely in the aphyric types. Clinopyroxene is a major groundmass phase in the series hawaiite through to anorthoclase trachyte, and aegirine is present in some of the comendites both as a groundmass and, more rarely, as phenocryst phases. Megacrysts and cumulates of clinopyroxene and orthopyroxene are common in the hawaiites and are present in one mugearite.

The Comboyne tholeiitic andesite contains abundant megacrysts

and cumulates of orthopyroxene which are commonly rimmed by clinopyroxene. It also contains phenocrysts of clinopyroxene and both Ca-poor and Ca-rich monoclinic pyroxenes in the groundmass.

Pyroxenes are the major mafic phases in the Lorne Basin calc-alkaline rocks, although secondary alteration to amphibole and chlorite is common.

3.3.2 Alkaline Rocks

3.3.2.1 Textural Relationships

Chemical and textural data indicate that the majority of large pyroxene crystals in the hawaiites are of megacryst and cumulate origin, and that clinopyroxene phenocrysts are comparatively rare. Both olivine and orthopyroxene occur within clinopyroxene megacrysts indicating that crystallisation of clinopyroxene continued after that of the two magnesium-rich minerals ceased. Cumulates of clinopyroxene, orthopyroxene, olivine, plagioclase and ilmenite indicate that at some stage these phases crystallised simultaneously.

The Comboyne hawaiites contain six morphologically distinguishable pyroxenes. These are:

1. Euhedral to subhedral clinopyroxene megacrysts range up to 8.0 mm in length, and are black and glassy in the hand specimen. In thin section they are a light buff colour and have a sieved rim approximately 0.5 mm wide (Plate 5b). This rim in turn is sometimes enclosed within a granular reaction rim, 0.25 mm wide, consisting of olivine and clinopyroxene. These megacrysts contain rounded inclusions of orthopyroxene. Cleavage and zoning

are absent and conchoidal fracturing is common.

2. Cumulate clinopyroxene occurs with orthopyroxene poikilitically enclosed in large plagioclase crystals (Plate 3a,b). The proportion of plagioclase varies and some cumulates consist solely of clinopyroxene and orthopyroxene with minor olivine. The clinopyroxene in these cumulates exhibits a narrow reaction rim where in contact with the melt, and sometimes good cleavage. Rounded inclusions of opaques are similar to those noted as being characteristic of megacrysts by Irving (1974).

3. Orthopyroxene megacrysts range upto 1 mm across, are slightly pleochroic green to pink in thin section, and black and glassy in hand specimen. Cleavage is occasionally developed, but conchoidal fracturing is more common. The orthopyroxenes have a broad reaction rim of granular olivine and clinopyroxene where in contact with the melt, which, in some instances has completely replaced the original orthopyroxene (Plate 1a,b). This granular aggregate has, in turn, been partly altered to a felting of haematite. Similar orthopyroxene occurs in cumulates with olivine, plagioclase and ilmenite (Plate 2a), and it also occurs as subrounded inclusions in clinopyroxene megacrysts (Plate 5b) and cumulates as described in (2) above. The cumulate orthopyroxene only has reaction rims where it is in contact with the host melt.

4. Subhedral and anhedral clinopyroxenes less than 1.0 mm across are a clear green-buff colour in thin section, with sieving throughout or confined to the rim (Plate 6a).

These pyroxenes exhibit little zoning, and occur as discrete

grains and in crystal aggregates with plagioclase, ilmenite and striated apatite.

5. Phenocrysts of euhedral clinopyroxene, green-buff in thin section, range up to 0.75 mm. These have a narrow, strongly zoned rim, show minor corrosion and have inclusions of acicular apatite.

6. Groundmass clinopyroxene is a green-buff colour in thin section and granular to prismatic in form (up to 0.1 mm across), and has inclusions of acicular apatite. Similar pyroxene occurs in reaction rims around orthopyroxene and clinopyroxene megacrysts.

The rare clinopyroxene phenocrysts in mugearites have broad twin lamellae and slight corrosion to the rim. One specimen has megacrysts and cumulates of orthopyroxene and clinopyroxene exhibiting similar reaction rims as those in the hawaiites (Plate 5a). Groundmass clinopyroxene is pale green in thin section, granular in the aphyric specimens, and granular to prismatic in the porphyritic specimen.

Benmoreites contain green (in thin section) subhedral micro-phenocrysts of clinopyroxene, and the groundmass clinopyroxene is generally interstitial to the more abundant feldspar. Inclusions of clear acicular apatite are common.

Clinopyroxene, pale green in thin section, occurs in the mafic and anorthoclase trachytes. Rare prismatic twinned phenocrysts up to 1.2 mm in length have inclusions of acicular apatite and Fe-Ti oxides and exhibit some sieving. The groundmass clinopyroxene is optically similar and tends to be granular in the mafic trachytes and to form euhedral to ragged elongate crystals in the anorthoclase trachytes.

Additionally, the Mt. Kophi anorthoclase trachyte (MU6627) contains an euhedral, buff in thin section, clinopyroxene, 1.00 mm across, which has a slightly sieved rim.

Pale green clinopyroxene occurs as microphenocrysts in the alkaline pitchstone (MU6636), and aegirine-augite and aegirine, deep green in thin section, are present as euhedral and subhedral phenocrysts and as irregular "mossy" plates in the comendites. Pleochroism is χ = emerald green, γ = yellow-green, Z = pale brownish-yellow, and $\chi \wedge c$ is $1^\circ - 8^\circ$.

Aegirine-augite/aegirine and sodic amphiboles are the major ferromagnesian minerals in the comendites. Aegirine-augite/aegirine is comparatively rare in the porphyritic and non-porphyritic flows, but is more common in the coarser-grained plug-forming comendites, and in the comendite forming the summit of Mt. Killabakh (MU6640), prismatic aegirine is the major mafic phase.

3.3.2.2 Analytical Results

Clinopyroxene and orthopyroxene analyses are listed in Tables 3.3 and 3.4, and plotted on a Ca - Mg - (Fe + Mn) diagram in Fig. 3.1.

Hawaiite clinopyroxenes plotted on this diagram fall into two areas. The comparatively low calcium content of one group is similar to that reported for high-pressure clinopyroxenes (Binns et al., 1970; Wass, 1971), and textural evidence (Pages 36 and 37) supports the suggestion that a substantial proportion of the large pyroxene crystals in the hawaiites crystallised at elevated pressures. These pyroxenes also have relatively high Al_2O_3 values (4.37-6.06%) compared with the normal phenocryst and groundmass pyroxenes (1.55-3.35%, Table 3.3). The second group, which includes phenocryst, groundmass and strongly sieved and

TABLE 3.3

MICROPROBE ANALYSES OF CLINOPYROXENES IN ALKALINE ROCKS

Sample No.	Hawaiite										Mugearita									
	6498	6498	6498	6498	6498	6597	6570	6571	6597	6597	6572	6498	6498	6597	6597	6578	6578	6567	6567	6607
Phase:	C	C	C	M	M	M	P	P	S	G	G	G	Rpx	Rpx	Rpx	C	M	S	G	C
SiO ₂	50.94	49.10	49.00	50.04	51.62	51.38	50.35	50.90	51.26	50.99	49.31	50.83	50.73	50.30	52.34	49.50	49.56	50.32	51.21	50.32
TiO ₂	.68	1.54	1.41	.66	.82	.82	.64	.78	.98	1.12	2.04	1.52	1.29	1.01	.75	1.62	.45	1.52	1.00	1.70
Al ₂ O ₃	5.29	6.06	5.04	4.37	4.66	4.53	2.26	2.05	1.55	1.65	3.35	2.29	2.59	4.51	1.92	2.74	6.46	1.71	2.03	2.10
Cr ₂ O ₃	.26	-	-	-	.33	.40	-	-	.21	-	.15	-	-	.38	.42	.15	.40	.18	.17	.12
Fe ₂ O ₃	.76	2.56	1.33	3.05	-	.97	.96	.99	2.52	1.41	1.09	1.31	.07	1.30	.26	5.70	2.18	1.50	.12	.91
FeO	7.67	8.24	11.50	10.06	9.03	6.58	14.75	13.01	8.72	12.13	11.48	9.84	11.14	7.50	9.58	5.61	6.62	13.26	12.75	12.69
MnO	-	.10	.16	.16	.07	-	.41	.18	.18	.35	.22	.19	-	-	-	-	.28	.30	.25	
MgO	16.01	14.00	13.11	13.42	15.35	16.91	10.18	11.09	13.88	12.46	11.65	12.76	13.36	14.80	15.63	13.39	14.54	10.62	11.87	10.52
CaO	17.45	17.41	17.02	17.37	17.39	17.94	19.50	18.63	20.09	19.40	20.13	20.57	20.33	19.81	18.69	20.32	18.94	19.48	20.04	20.54
Na ₂ O	.73	.98	.64	.87	.74	.56	.66	.73	.61	.55	.58	.61	.49	.39	.41	.57	.76	.66	.45	.74
K ₂ O	-	-	-	-	-	-	-	-	-	-	-	.08	-	-	-	-	.09	.24	.07	.12
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
Cation Content (6 oxygen atoms)																				
Si	1.8747	1.8346	1.8383	1.8780	1.9005	1.8650	1.9449	1.9298	1.9329	1.9366	1.8745	1.9172	1.9092	1.8595	1.9434	1.8908	1.8224	1.9388	1.9401	1.9205
Al ^{IV}	.1253	.1654	.1617	.1220	.1028	.1140	.0551	.0702	.0671	.0634	.1255	.0828	.0908	.1305	.0516	.1092	.1778	.0712	.0599	.0795
Al ^{VI}	.1041	.1016	.0964	.0714	.0994	.0862	.0471	.0571	.0016	.0106	.0247	.0191	.0240	.0670	.0273	.0125	.1023	.0064	.0305	.0149
Ti	.0242	.0433	.0359	.0187	.0226	.0170	.0134	.0221	.0279	.0320	.0582	.0432	.0365	.0282	.0208	.0459	.0407	.0445	.0284	.0466
Cr	.0076	-	-	-	.0092	.0117	-	-	.0061	-	.0045	-	-	.0110	.0123	.0044	.0117	.0055	.0150	-
Fe ³⁺	.0209	.0714	.0376	.0860	-	.0265	.0277	.0282	.0711	.0402	.0311	.0371	.0021	.0362	.0072	.1614	.0601	.0432	.0034	.0626
Fe ²⁺	.2354	.2558	.3607	.3151	.2778	.2015	.4701	.4117	.2732	.3843	.3637	.3094	.3492	.2324	.2973	.1766	.2027	.4238	.4037	.4051
Mn	-	.0030	.0048	.0051	.0019	-	.0134	.0055	.0058	.0112	.0069	.0061	-	-	-	-	.0090	.0095	.0081	
Mg	.8769	.7794	.7330	.7506	.8428	.9244	.5795	.6714	.7799	.7059	.6601	.7173	.7492	.8196	.7434	.7558	.7969	.6183	.6700	.5902
Ca	.6877	.6966	.6840	.6984	.6854	.7035	.8013	.7586	.8117	.7900	.8195	.8312	.8167	.7866	.8644	.8209	.7464	.8900	.8123	.8399
Na	.0517	.0706	.0463	.0633	.0513	.0397	.0492	.0524	.0446	.0402	.0427	.0442	.0360	.0284	.0297	.0415	.0542	.0491	.0323	.0533
K	-	-	-	-	-	-	-	-	-	-	-	.0038	-	-	-	-	.0037	.0118	.0032	.0056
Total	4.0276	4.0217	4.0330	4.0026	3.9913	4.0106	4.0367	4.0370	4.0219	4.0144	4.0114	4.0114	4.0137	3.9752	4.0024	4.0190	4.0187	4.0116	4.0001	4.0090
100Mg / (Mg + Fe ²⁺) ¹	77.4	70.4	64.8	65.2	75.2	80.2	53.8	68.3	69.4	62.5	62.6	67.4	68.1	77.9	70.9	69.1	75.2	57.0	62.2	59.6
1. Fe ²⁺ is total iron as Fe ²⁺																				
End members calculated by Kushiro's (1962) method																				
NaFe ³⁺ SiO ₆	.41	6.49	3.25	5.58	-	2.44	2.44	2.44	4.06	1.83	2.64	3.25	-	2.44	.81	3.65	5.20	3.65	.41	2.44
NaAlSi ₃ O ₆	4.43	-	.80	-	4.83	1.21	2.01	2.41	-	-	.80	.80	3.22	-	2.01	-	-	1.41	2.41	2.81
CaTiAl ₂ O ₆	2.62	4.52	4.28	1.90	2.38	1.90	1.90	2.38	2.86	3.33	6.19	4.52	3.81	3.09	-	4.78	1.19	4.52	3.09	5.09
CaFe ³⁺ AlSiO ₆	-	-	-	2.47	-	-	-	-	2.96	-	-	-	-	.99	-	1.98	-	-	-	-
CaAl ₂ SiO ₆	6.76	8.72	8.07	6.54	5.67	7.63	1.96	2.62	-	.44	1.31	.44	.22	6.98	1.53	.87	13.09	-	.44	-
CaSiO ₃	31.36	29.16	28.93	30.44	31.83	32.30	38.45	36.13	28.80	38.34	37.99	40.19	40.08	33.32	36.83	39.60	30.79	39.10	39.85	40.19
MgSiO ₃	39.86	34.84	32.63	33.44	38.26	42.07	25.30	29.62	34.54	31.03	29.02	31.03	33.24	36.85	38.96	33.44	35.34	26.91	29.52	26.21
FeSiO ₃	14.16	15.17	21.11	18.47	16.76	12.14	27.84	24.28	16.36	22.96	21.51	18.08	20.45	13.72	17.55	16.36	11.87	24.94	24.01	24.15
Norms (weight percent)																				
Q	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	2.59	-	-	-	-
Or	-	-	-	-	-	-	-	-	-	-	-	.47	-	-	-	-	.47	1.42	.41	.71
Ab	6.17	8.29	5.41	7.36	6.26	4.74	5.58	6.17	5.16	4.65	4.91	5.16	4.14	3.30	3.47	4.82	6.19	5.59	3.81	6.12
Ne	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	.04	-	-	.08
An	11.16	12.14	13.06	8.02	9.39	10.12	3.21	4.50	1.49	2.03	6.54	3.27	4.87	10.56	3.40	4.92	13.66	1.00	3.31	2.66
Ol	60.26	59.36	57.93	63.09	61.80	62.74	77.60	72.24	78.80	76.82	75.90	79.80	77.79	70.07	71.90	75.63	62.24	78.46	78.78	82.00
Hy	10.19	4.61	9.79	9.99	12.62	11.84	5.27	8.53	7.55	9.58	1.20	2.54	2.15	5.72	15.63	.85	-	5.01	7.13	-
Ol	9.08	8.99	9.22	5.86	7.88	7.39	5.51	5.67	1.19	2.81	5.72	3.97	8.36	6.00	3.21	-	10.57	3.16	4.27	4.46
Ut	1.10	3.71	1.93	4.42	-	1.41	1.39	1.44	3.65	2.04	1.58	1.90	.10	1.88	.36	8.29	3.09	2.17	.17	1.32
Ca	0.38	-	-	-	.49	.59	-	-	.31	-	.22	-	-	.56	.62	.22	.57	.27	.25	.18
Il	1.67	2.92	2.68	1.25	1.56	1.18	1.22	1.48	1.86	2.13	3.87	2.89	2.45	1.92	1.42	3.08	.84	2.92	1.90	3.23

Fe₂O₃ calculated by the method of Ghent and Coleman (1973)

M = Microcryst

G = Groundmass

C = Cumulate

S = Sieved

P = Phenocryst

R = Rim around opx or cpx

MICROPROBE ANALYSES OF CLINOPYROXENES IN ALKALINE ROCKS

1. Fe^{2+} is total iron as Fe^{2+} .

End members calculated by Kushiro's (1962) method

$\text{NaFe}^3\text{SiO}_6$	2.44	4.06	.81	-	4.87	5.68	5.68	6.90	2.44	3.25	3.25
$\text{NaAlSi}_3\text{O}_8$	1.21	-	3.22	2.81	.40	-	.40	2.41	3.62	1.21	3.22
$\text{CaTiAl}_2\text{O}_6$	2.06	3.09	2.62	1.43	1.43	1.19	1.19	3.81	1.43	1.19	1.42
$\text{CaFe}^3\text{AlSiO}_6$	-	-	-	-	-	2.47	-	-	-	-	-
$\text{CaAl}_2\text{SiO}_6$.87	.65	.22	1.31	1.96	.65	1.96	4.58	.44	1.31	-
CaSiO_3	41.47	41.24	40.43	38.22	39.38	38.45	39.03	36.01	39.56	41.12	38.10
MgSiO_3	21.19	19.38	22.69	32.53	15.06	15.66	14.86	33.84	15.36	23.90	.90
FeSiO_3	29.82	30.74	29.29	22.17	36.15	35.36	35.62	14.78	36.28	27.71	52.91

Norms (weight percent)

Q	-	-	-	-	-	-	-	-	-	-	-	6.85	2.46	3.85	2.53
Or	-	.77	1.06	-	1.06	1.06	1.30	.41	1.36	1.06	.71	2.66	.47	4.53	.77
Ab	4.66	4.82	5.24	3.72	5.92	6.00	6.23	5.58	6.68	4.74	3.55	10.71	3.82	-	1.90
Na	-	-	-	-	-	-	-	3.12	-	-	-	-	-	-	-
An	2.77	1.51	2.14	3.26	.42	.02	-	5.81	.55	.50	-	-	-	-	-
Ac	-	-	-	-	-	-	.69	-	-	-	-	3.62	14.00	44.26	63.48
Ks	-	-	-	-	-	-	-	-	-	-	-	-	-	1.54	-
Ne	-	-	-	-	-	-	-	-	-	-	-	.05	6.94	2.36	2.05
Di	84.25	85.13	81.50	74.87	83.32	82.14	82.78	72.43	82.68	83.68	82.47	29.69	39.60	1.42	5.97
Hy	1.47	.76	2.81	14.81	2.63	4.34	2.60	-	1.76	4.78	5.90	25.65	5.05	21.29	.71
Ol	3.73	2.70	5.03	2.47	2.80	1.44	2.49	6.81	4.68	2.51	2.54	-	-	-	-
Wt	1.29	2.33	.42	-	2.75	4.32	2.94	3.84	1.42	1.75	-	-	-	-	-
Cs	-	-	-	-	.25	-	.25	.32	-	.21	.29	.27	-	-	-
Il	1.84	1.98	1.69	0.85	.85	.68	.72	2.37	.85	.92	.87	.84	1.10	1.84	1.27
Ap	-	-	-	-	-	-	-	-	-	-	-	2.41	.86	-	-

Fe_2O_3 calculated by the method of Ghent and Coleman (1973)

C = Cumulate S = Slaved

P = Phenocryst E = Even grained

G = Groundmass

TABLE 3.4

MICROPROBE ANALYSES OF ORTHOPYROXENES IN ALKALINE ROCKS

Rock Type:	Hawaiiite					Mugearite
Sample No:	6498	6498	6498	6597	6597	6578
Phase	C	C	M	C	M	M
SiO ₂	51.03	52.25	51.35	53.06	52.54	55.46
TiO ₂	.50	.41	.42	.36	.32	.17
Al ₂ O ₃	2.54	3.77	4.10	3.01	3.54	1.60
Cr ₂ O ₃	-	-	-	-	-	.18
FeO	23.07	15.87	15.56	13.20	13.63	11.18
MnO	.21	.11	-	.13	.16	-
MgO	20.71	24.96	24.43	26.96	26.16	29.20
CaO	1.82	1.98	1.79	1.99	1.90	2.08
Na ₂ O	-	.09	-	-	.20	-
K ₂ O	-	-	-	-	-	-
Total	99.88	99.35	97.65	98.71	98.45	99.87

Cation Content (6 oxygen atoms)

Si	1.9183	1.9050	1.9029	1.9244	1.9158	1.9654
Al ^{IV}	.0817	.0950	.0971	.0756	.0842	.0346
Al ^{VI}	.0326	.0671	.0818	.0529	.0677	.0323
Ti	.0142	.0113	.0118	.0098	.0087	.0045
Cr	-	-	-	-	-	.0051
Mg	1.1605	1.3566	1.3492	1.4573	1.4215	1.3420
Fe	.7252	.4840	.4822	.4002	.4156	.3313
Mn	.0068	.0016	-	.0039	.0050	-
Ca	.0735	.0770	.0709	.0773	.0741	.0790
Na	-	.0060	-	-	.0139	-
K	-	-	-	-	-	-
Total	4.0128	4.0044	3.9959	4.0014	4.0064	3.9942

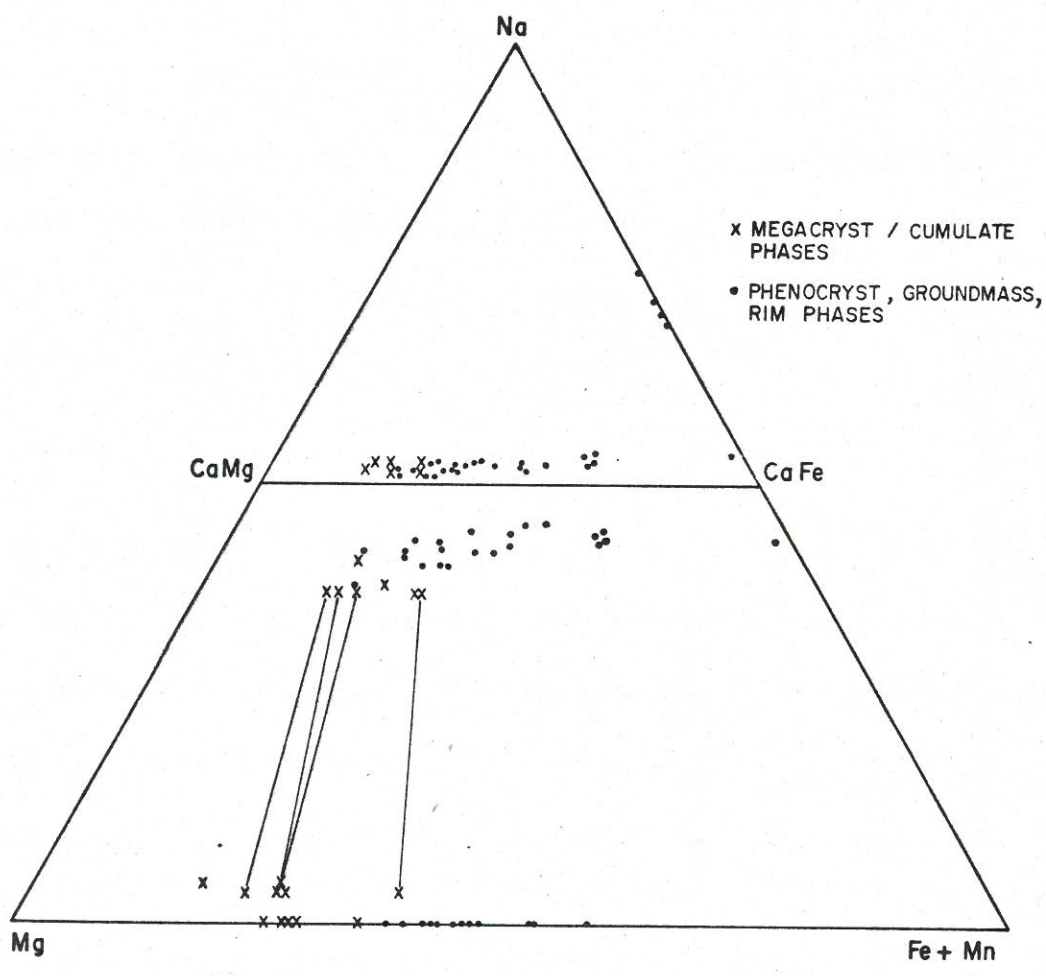
$\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}}$ 2+	61.5	73.7	73.7	78.5	77.4	82.3
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Total iron as FeO

C = Cumulate

M = Megacryst

Fig. 3.1 Trends on CaMg - CaFe - Mg - (Fe + Mn) and Na - CaMg - CaFe diagrams for pyroxenes in the Comboyne alkaline series hawaiiite - comendite, with tie lines between coexisting clinopyroxene - orthopyroxene pairs. Coexisting olivines are plotted on the Mg - (Fe + Mn) join.



reaction rim phases, has CaO and Al_2O_3 values consistent with a crystallisation or reaction origin in a near-surface environment. The anomaly of strongly sieved crystals having the same composition as euhedral phenocryst phases has already been discussed in relation to olivines. In clinopyroxenes there is a marked correlation between size of crystal and degree of sieving, with the smaller crystals showing more complete resorption. As the sieved rims of the large clinopyroxene megacrysts have a similar composition as the associated phenocryst and groundmass phases, it appears likely that the small size of some megacrysts has allowed for the complete reaction of the crystal to a composition in equilibrium with the melt at the time of phenocryst and groundmass precipitation.

The high-level clinopyroxenes exhibit slight enrichment in CaO from hawaiite to benmoreite, followed by a decrease in trachytes (Fig 3.1). Enrichment in Na_2O with differentiation is not evident in clinopyroxenes from hawaiite through to trachyte indicating there has been little replacement of $CaFe^{2+}$ by $NaFe^{3+}$. There is slight enrichment in Na_2O in the trachytes, followed by a marked increase in the acmite molecule in the comendites (Fig. 3.1). Although there is a strong similarity in bulk rock composition between the alkaline pitchstone and the comendites, the pyroxenes in these rocks are ferro-hedenbergite and aegirine-augite/aegirine respectively. A similar trend has been noted in comendites from Mayor Island, New Zealand by Nicholls and Carmichael (1969). In these, soda-hedenbergite occurs as micro-phenocrysts in glassy rocks, and aegirine-rich pyroxenes are found in crystalline comendites, usually as a groundmass phase. Nicholls and Carmichael suggest that the precipitation of an acmite pyroxene instead of soda-hedenbergite in liquids of the

same composition may require a higher oxygen fugacity at any given temperature.

Fig. 3.1 shows that although pyroxenes show only limited variation in Mg:Fe ratio in any one rock type, this ratio varies widely throughout the series, with clinopyroxenes ranging in composition from augite to ferro-hedenbergite. There is no strong evidence of groundmass pyroxene being enriched in iron relative to phenocryst phases as has been noted in other alkaline associations (Uchimizu, 1966; Wass, 1971), and in one anorthoclase trachyte there appears to be a decrease in Fe:Mg ratio between phenocryst and groundmass pyroxenes. Overall, pyroxenes show a continuous enrichment in Fe relative to Mg in the series hawaiite through to comendite, with marked increases from benmoreite to trachyte and from trachyte to alkaline pitchstone and comendite. The variation in the bulk host rock composition between benmoreite and trachyte does not appear sufficiently great to account for these compositional differences in the pyroxenes. However, this trend is probably related to a decrease in Fe-Ti oxide phases and the rarity of olivine in the trachytes. A greater proportion of Fe^{2+} , partitioned in the benmoreites between Fe-Ti oxides, olivine and clinopyroxene (with olivine undergoing Fe-enrichment more rapidly than pyroxene), is now available for incorporation in pyroxene in the trachytes.

When analysed pyroxenes are arranged in order of inferred differentiation based on SiO_2 enrichment in the host rock (excluding those pyroxenes with chemical and textural characteristics indicating a high-pressure origin), Al_2O_3 shows a tendency to decrease with fractionation, and TiO_2 content is fairly constant from hawaiite to benmoreite and then decreases in the

more evolved rocks. Al_2O_3 and TiO_2 values for the clinopyroxenes in the Comboyne alkaline series are similar to other near-saturated alkaline series such as Dogo Island (Uchimizu, 1966), Iki Island (Aoki, 1959, 1964), Nandewar Mountain (Abbott, 1969), Aden and Little Aden (Cox et al., 1970), and contrasts with undersaturated alkaline series where Al_2O_3 and TiO_2 values are higher and trend towards enrichment with concomitant depletion in SiO_2 with increased fractionation (Morotu Island, Yagi, 1953; Gough Island, Le Maitre, 1962; Shorkin Sag, Nash and Wilkinson, 1970; Lake Rudolf, Brown and Carmichael, 1971; Southern Highlands of New South Wales, Wass, 1971).

As mentioned previously, clinopyroxene megacrysts/cumulates can be identified by high Al_2O_3 (contained in the $\text{CaAl}_2\text{SiO}_6$ end-member, without correspondingly high TiO_2), and CaO values less than those typical of low-pressure pyroxenes. There is also a tendency for Na_2O to be higher in the megacryst/cumulate phases than in the associated phenocryst and groundmass pyroxenes, reflecting an increase in the jadeite component.

The orthopyroxene megacrysts/cumulates in the Comboyne hawaiites and mugearite are calcic bronzites and in the hawaiites have higher Al_2O_3 values than commonly found in orthopyroxenes which have crystallised at low pressures (Table 3.4). The Al_2O_3 content in the hawaiite orthopyroxenes varies from 2.54% to 4.10% which is comparable with values reported for orthopyroxenes in other alkaline rocks (Kuno, 1964; Binns et al., 1970; Irving, 1974). The $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratio in the orthopyroxenes shows considerable variation in any one specimen, the maximum range observed being 73.7 to 61.5. Orthopyroxene megacrysts in the mugearite are enriched in MgO and depleted in Al_2O_3 and TiO_2 relative to those in the hawaiites.

3.3.2.3 Discussion

Chemical trends in pyroxenes are dependent on the composition of the melt and point to the particular basaltic lineage of the host magma (Coombs, 1963; Muir and Tilley, 1964), and although Wilkinson (1957a) noted that pyroxenes in alkali olivine basalt associations show only minor variation in composition with fractionation when compared with tholeiitic pyroxenes, it is now apparent that this conclusion was influenced by the limited compositional range of the rocks studied. Yagi (1953) and, more recently, Aoki (1964), Nash and Wilkinson (1970), Barberi et al. (1971) and Gibb (1973), have shown that clinopyroxenes in alkali basalt associations have a wide range in composition and three distinct fractionation paths have been noted:

1. Diopside - augite - soda augite - aegirine-augite - aegirine trend, regarded as typical of strongly alkaline associations (Nash and Wilkinson, 1970).
2. Diopsidic augite - augite - ferroaugite - hedenbergite trend, recognised as being typical of mildly alkaline basaltic associations (Abbott, 1969; Barberi et al., 1971; Gibb, 1973). Pyroxenes in the Comboyne alkaline series follow this trend.
3. Diopsidic augite - augite - ferropigeonite trend, considered by Aoki (1964) to crystallise under conditions of high temperature and low oxygen partial pressure. Gibb (1973) suggests this trend is similar to quenched trends in tholeiitic associations as described by Smith and Lindsley (1971).

It is generally recognised (Le Maitre, 1962; Uchimizu, 1966; Abbott, 1969; Barberi et al. (1971) that high oxygen fugacity

results in an increase in the acmite component in pyroxenes resulting in the diopside - soda augite - aegirine-augite - aegirine fractionation path of strongly alkaline melts. Aoki (1964) and Yagi (1966) report that the crystallisation of highly sodic pyroxenes is favoured by high or increasing $f(O_2)$. Recent experimental work (Gilbert, 1969; Bailey, 1969) however, indicates that acmite can crystallise at low $f(O_2)$ under anhydrous conditions. This is substantiated by the association of aenigmatite with acmite in peralkaline rocks. The majority of alkaline rocks exhibiting the diopside -soda augite - aegirine fractionation path belong to the shallow intrusive teschenite-syenite series which is typically volatile-rich. The contrasting augite - hedenbergite trend is commonly found in volcanic, mildly alkaline series which are generally lower in volatiles.

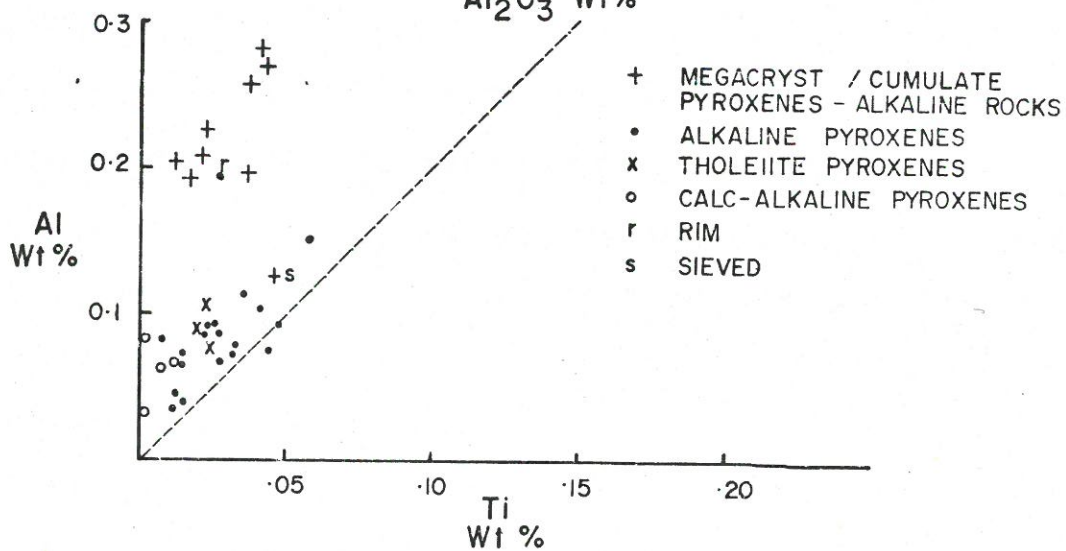
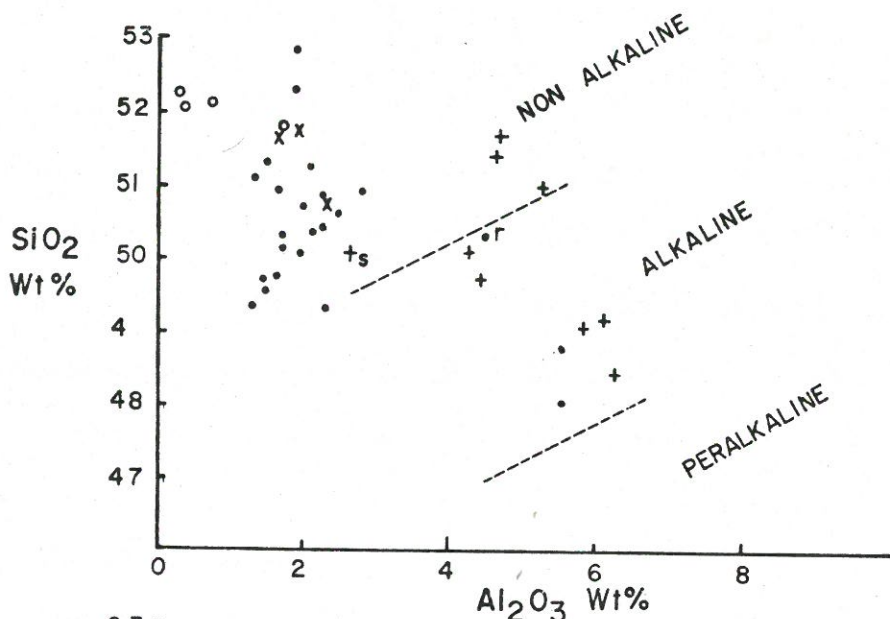
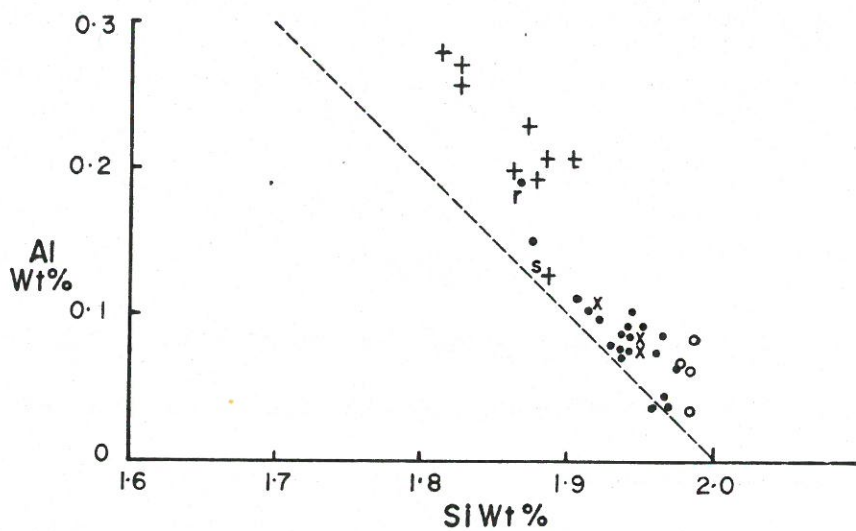
Kushiro (1960) has recognised that the proportion of Al in pyroxenes is a function of silica content in the melt and that pyroxenes in tholeiitic magmas oversaturated in SiO_2 have a higher proportion, in general, a stoichiometric amount, of Si relative to Al in the tetrahedral position than pyroxenes in undersaturated alkali basalt magmas. He demonstrated in plots of Al against Si and SiO_2 against Al_2O_3 that pyroxenes from tholeiitic, alkaline and peralkaline rocks fall into different fields. Experimental work by Gupta et al. (1973) has confirmed that as silica concentration increases in the diopside - $CaTiAl_2O_6$ - SiO_2 system there is a decrease in the solubility of Al and Ti in diopsidic pyroxenes and that the concentration of silica controls the stability of coexisting Ti-bearing phases.

Figures 3.2 a and b show that plots of Al against Si and SiO_2 against Al_2O_3 in pyroxenes from the Comboyne rocks, with the exception of megacryst and cumulate phases, fall into fields

Fig. 3.2a Plots of Al against Si for clinopyroxenes in the alkaline, tholeiitic and calc-alkaline rocks. Straight line is $Si + Al = 2$. Megacryst/cumulate clinopyroxenes fall away from this line suggesting these pyroxenes are rich in the Ca-Tschermak molecule. (Kushiro, 1960)

Fig. 3.2b Plots of SiO_2 against Al_2O_3 for clinopyroxenes in the alkaline, tholeiitic and calc-alkaline rocks showing non-alkaline, alkaline and peralkaline fields designated by LeBas (1962). Megacryst/cumulate pyroxenes from alkaline rocks are concentrated in the alkaline field, but the coexisting phenocryst and groundmass pyroxenes plot in the non-alkaline field reflecting the near-saturated nature of the Comboyne alkaline rocks.

Fig. 3.2c Al plotted against Ti for alkaline, tholeiitic and calc-alkaline clinopyroxenes. Straight line is $2Al = Ti$. The megacryst/cumulate clinopyroxenes plot away from this line indicating that Al is incorporated in the Ca-Tschermak molecule. (Kushiro, 1960).



designated by Kushiro (1960) and LeBas (1962) as typical of tholeiitic pyroxenes. This results directly from the silica-saturated rather than the tholeiitic nature of the Comboyne rocks, and indicates that these diagrams can only be used to indicate degree of silica-saturation or undersaturation and not the lineage of the host magma. A similar conclusion was made by Barberi et al. (1971) and Gibb (1973). However, Al and Ti are incorporated in the $\text{CaTiAl}_2\text{O}_6$ pyroxene end member in low-pressure silica-undersaturated melts, and not the $\text{CaAl}_2\text{SiO}_6$ end member as suggested by Barberi et al.

The normative components of pyroxenes can also be used to distinguish between alkaline and tholeiitic rocks and this is discussed more fully on page 50.

LeBas (1962) has noted that clinopyroxenes in alkaline rocks generally have TiO_2 contents greater than 1%, whereas in sub-alkaline rocks they have less than 1% TiO_2 . Again it would appear that Ti content is more directly dependent on degree of silica-saturation and oxygen fugacity than on rock type.

Verhoogen (1962) suggested the preferential substitution of Ti in silicate phases when both the partial pressure of oxygen and temperature are high, and Carmichael and Nicholls (1967) have noted that the oxygen fugacity of volcanic tholeiites may be an order of magnitude lower than in alkaline basalts. Thus, even if a tholeiitic melt was not saturated in silica, the early precipitation of Ti-oxides (and in particular ilmenite which is typically the first oxide to crystallise in tholeiitic melts (Wass, 1971)), would restrict the availability of Ti for incorporation in clinopyroxene.

It is, therefore, not surprising that titanaugites ($\text{TiO}_2 > 2\%$ - Yagi and Onuma, 1967) are absent in the near-saturated, low $f(\text{O}_2)$ Comboyne alkaline rocks. However, titanaugite does occur

in alkali olivine basalt from the Nandewar Mountain series, and the strong similarity between this and the Comboyne association suggests that if rocks less evolved (along a silica-enrichment trend) than hawaiites were located in the Comboyne area, titan-augite could occur. Silica-saturation and low $f(O_2)$ in the Comboyne magma probably also accounts for the lack of Ti-enrichment in late-formed pyroxenes as predicted by experimental work of Yagi and Onuma (1967). They found that Ti-concentration in alkali rocks increases under low-pressure fractionation, and where fractionation trends towards high alkalis there is enrichment in the aegirine component at the expense of Ti. Experimental work (Yoder and Tilley, 1962; Nesbitt and Hamilton, 1970) has also shown that in "dry" basaltic melts plagioclase crystallises ahead of pyroxene and this limits the availability of Al (and therefore Ti) for incorporation in the pyroxene structure.

When Al is plotted against Ti (Fig. 3.2c), it is apparent that the Comboyne pyroxenes fall into two distinct groups. Those considered to have crystallised at low-pressures plot near the $2Al - Ti$ line and contrast with megacryst and cumulate pyroxenes which have a greater proportion of Al relative to Ti. This also indicates that Al is included in the $CaAl_2SiO_6$ end member rather than $CaTiAl_2O_6$ and concurs with the suggestion that these pyroxenes have crystallised at elevated pressures.

The pyroxene fractionation path of the Comboyne alkaline rocks is subparallel to the Di - Hc join (Fig. 3.1) and contrasts with that of tholeiitic associations (as typified by the Skaergaard intrusion) where the early depletion in Ca is followed by an increase in Ca when the coexisting Ca-poor pyroxene ceases to crystallise. It also contrasts with strongly alkaline associations where $CaFe^{2+}$ is replaced by $NaFe^{3+}$ during fractionation (Fig. 3.3).

Fig. 3.3 Comparison of clinopyroxene compositional trends in alkaline and tholeiitic provinces. Y = Sakhalin, Japan; G = Gough Island; B = Black Jack; S = Shiant Isles; Sk = Skaergaard (all from Uchimizu, 1966). A = Japanese alkaline basalt series (Aoki, 1964), N = Nandewar (Abbott, 1969). Solid circles represent clinopyroxenes from the Comboyne alkaline series.

Thus the absence of Na-enrichment in pyroxenes in the series hawaiite through to trachyte, and the trend towards iron-enrichment without accompanying depletion in Ca, is intermediate to fractionation trends exhibited by undersaturated alkaline and saturated tholeiitic basaltic magmas.

The strong correlation between the composition of the melt (in particular the degree of silica-saturation and under-saturation) and the pyroxene differentiation trend, shows that melt composition markedly controls that of the precipitating pyroxenes. It has been suggested (Kuno, 1965) that high $f(O_2)$ results in the formation of magnetite which, in turn, causes silica-enrichment in the melt. However, in mildly alkaline associations where differentiation is accompanied by silica-enrichment, the oxygen fugacity is low, and the near-saturated nature of the melt would appear to be governed by the initial composition of the parent magma rather than by strong SiO_2 enrichment resulting from magnetite fractionation. This is supported by the fact that magnetite phenocrysts are not abundant in the more basic members of these mildly alkaline associations (Gibb, 1973; Upton and Wadsworth, 1972; Ridley, 1971b). and thus silica-enrichment in the intermediate rock types cannot be attributed to this mechanism.

It is now recognised (Aoki, 1964; Aoki and Kushiro, 1968; Ito and Kennedy, 1968) that the incorporation of the Ca-Tschermak molecule in pyroxenes is pressure dependent, and that pyroxenes formed at high pressures contain substantially greater proportions of $CaAl_2SiO_6$ component than do low-pressure pyroxenes. Thus, in contrast to Al-rich clinopyroxenes that have crystallised at low pressure, Al is in 6-fold coordination and is not accompanied by Ti-enrichment and Si-depletion (Fig. 3.2c). The

Ca-Tschermak molecule values in the Comboyne hawaiite megacryst and cumulate clinopyroxenes vary from 5.67% to 8.72% and contrast with values ranging from 0.0 to 2.62% for the associated phenocryst and groundmass phases (Table 3.3). The megacryst/cumulate Ca-Tschermak values are not as high as values given for high-pressure clinopyroxenes in strongly undersaturated alkaline rocks, but are significantly higher than values typical of clinopyroxenes considered to have crystallised at low pressures and this suggests their formation at elevated pressures.

Textural characteristics as well as chemical composition of the orthopyroxenes in the Comboyne hawaiites indicate they crystallised at elevated pressures and in near-chemical equilibrium with the coexisting clinopyroxenes and cumulates (Fig. 3.1). Al_2O_3 values vary from 2.54-4.10% which, although not as high as some values reported from other associations where orthopyroxene megacrysts have been identified (Best, 1970; Binns et al. 1970; Kuno, 1964; Wass, 1971; Irving, 1974), or for synthetic orthopyroxenes crystallised at high pressures (Green and Ringwood, 1967; Bultitude and Green, 1968), are still higher than values normally associated with orthopyroxenes from sub-alkaline associations (0.4-3.1%, Kuno, 1954).

High Al_2O_3 values in orthopyroxenes which have crystallised at elevated pressures has led to the suggestion that high pressure increases the solubility of Al_2O_3 in orthopyroxene. Conflicting with this observation are the experimental results of Anastasiou and Seifert (1972) which indicate that the solid solubility of Al_2O_3 in orthorhombic enstatite is mainly a function of temperature and not strongly pressure dependent. Nicholls and Lorenz (1973) however, found experimentally that Al-enrichment in orthopyroxenes, controlled by orthopyroxene-liquid partition, is restricted by the

co-precipitation of clinopyroxene and is the reverse of that reported by Anastasiou and Siefert for subsolidus and near-solidus orthopyroxenes in equilibrium with Al-rich crystalline phases in the $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2\text{-H}_2\text{O}$ system. It is also significant that in a study on pyroxene compositions in ultramafic rocks, Wilshire and Jackson (1974) found wide variations in compositions within single xenoliths and concluded that pyroxenes are subject to bulk composition changes resulting from partial equilibration by igneous and metamorphic processes. The above data suggest that it may not be possible to positively identify high pressure orthopyroxenes on compositional trends (in particular, Al-content), although the common association of orthopyroxenes with high Al_2O_3 contents with other high pressure phases suggests that high Al_2O_3 in orthorhombic pyroxene is indicative of crystallisation at elevated pressures.

When orthopyroxene crystals are in contact with the host melt they have undergone reaction to form olivine and clinopyroxene. The secondary clinopyroxenes are enriched in SiO_2 and MgO and depleted in CaO and TiO_2 compared with reaction rims around clinopyroxene megacrysts and normal phenocryst and groundmass phases.

The importance of the composition of the parent melt in controlling the megacryst assemblage is shown by the occurrence of orthopyroxene megacrysts and, generally, the absence of high pressure spinel in mildly alkaline rocks which are hypersthene-normative (Binns et al., 1970). In contrast, experimental work under hydrous conditions at high pressures (Bultitude and Green, 1968) indicate that orthopyroxene is a major liquidus phase in synthetic olivine-nephelinite at 18 - 22.5 kb, suggesting that orthopyroxenes reported from undersaturated basalts (Southern Highlands of N.S.W.,

Wass, 1971; Western Victoria, Irving, 1974; Arizona, Best, 1970) would appear to have crystallised under hydrous conditions.

As is normal in coexisting pyroxenes, the Fe:Mg ratio is higher in clinopyroxene megacrysts and cumulates than the associated orthopyroxenes (Tarney, 1969; Clarke and Biggar, 1972), and tie-lines between coexisting pyroxenes (Fig.3.1) follow the usual trends. The strong chemical correlation between the composition of the host melt and that of the megacryst/cumulate assemblage (particularly when compared with trends in other alkaline associations) suggests that the megacrysts/cumulates are cognate, and this is further substantiated by experimental evidence (Chapter 5).

The proposed alkaline affinities of the Comboyne series are discussed in Chapter 4. However, the transitional nature of this association makes it particularly desirable to calculate the normative components of the pyroxenes, as these have been shown by Coombs (1963) and Muir and Tilley (1964) to reflect the basaltic affinities of the host magma. It is critical to know accurately the Fe_2O_3 :FeO ratio of the clinopyroxenes, and as electron microprobe analysis only determines iron as total FeO it has been necessary to arbitrarily determine the Fe_2O_3 :FeO ratios following the method outlined by Ghent and Coleman (1973).

The pyroxene normative components thus determined (Table 3.3) confirm the transitional nature of the Comboyne alkaline association. Normative nepheline occurs in only three pyroxenes (0.04 and 0.08% in mugearites, and 3.12% in a probable xenocryst in anorthoclase trachyte), and normative quartz is absent in all cases. Normative olivine and hypersthene vary from 9.22% to 0.99% and 15.63% to 0.00% respectively in the series hawaiite to trachyte. Normative hypersthene is highest in the hawaiites and decreases towards

trachytes. As observed in other associations which include sub-calcic clinopyroxene megacrysts (Walcha, Kyogle (Binns et al., 1970), Southern Highlands of N.S.W. (Wass, 1971)), the Comboyne megacryst and cumulate clinopyroxenes have significantly higher normative hypersthene than the coexisting phenocryst and ground-mass phases.

Normative clinopyroxene values listed by Coombs (1963) as typical of tholeiitic associations (Skaergaard and Stillwater) differ from the Comboyne series in having only limited normative olivine with higher values of normative hypersthene and occasional normative quartz. In contrast, pyroxenes listed from alkaline basaltic rocks have no normative quartz, rare normative hypersthene, and contain both normative olivine and nepheline. As noted previously, the Comboyne clinopyroxenes are intermediate to these.

3.3.3 Tholeiitic Rocks

3.3.3.1 Textural Relationships

Megacrysts and cumulates of colourless, non-pleochroic orthopyroxene are prominent in the Comboyne tholeiitic andesite and are associated with minor olivine. These orthopyroxenes show evidence of reaction at grain boundaries, and in most instances are enclosed in a rim of finely twinned clinopyroxene (Plate 7a,b). More rarely, clinopyroxene appears to be replacing the entire orthopyroxene grain. Subhedral clinopyroxene also occurs in gabbroic inclusions containing plagioclase and interstitial olivine, Fe-Ti oxides and glass (Plate 8a,b). Clinopyroxene is a rare phenocryst phase as well. Anhedral grains of augite and pigeonite, slightly buff in thin section, along with plagioclase and Fe-Ti oxides, are the main groundmass phases. Pigeonite has not been recognised optically and was identified with the microprobe.

3.3.3.2 Analytical Results

The orthopyroxene in the tholeiitic andesite is bronzite and is enriched in Mg relative to the coexisting clinopyroxenes (Fig. 3.4). The clinopyroxene in the gabbroic inclusions has similar Al_2O_3 and TiO_2 values to the pyroxene rimming the bronzite, and has slightly higher FeO and MgO and lower CaO values (Tables 3.5 and 3.6). The groundmass augite is enriched in Al_2O_3 and CaO and slightly depleted in SiO_2 relative to the associated clinopyroxene occurring in the gabbroic inclusions. $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratios vary from 79 in the orthopyroxene megacrysts, 68 in the clinopyroxene in the gabbroic inclusions, to 66 and 63 in the groundmass calcic and sub-calcic clinopyroxenes respectively.

3.3.3.3 Discussion

The bronzite megacrysts are texturally similar to those in the "augite-hypersthene-olivine-andesite" from the Iki Islands (Aoki, 1959), bronzites in picritic dykes in the Lewisian gneisses (Tarney, 1969) and in the olivine-rich lavas of Nuanetsi (Cox and Jamieson, 1974). Orthopyroxene megacrysts in tholeiitic rocks are also described from South Africa and Rhodesia (Jamieson, 1966), South West Germany (Nicholls and Lorenz, 1973), the Tweed Shield, N.S.W. (Duggan and Wilkinson, 1973), and Sutherland (1974) reports orthopyroxene megacrysts in addition to lherzolite inclusions in a tholeiitic basalt from the Tasmanian volcanic province.

As has already been noted, the bronzite is enriched in MgO and Al_2O_3 relative to the clinopyroxene rim phase and clinopyroxene in the gabbroic inclusions and the groundmass (Tables 3.5, 3.6). The higher Al_2O_3 values in the bronzite contrast with the common trend for coexisting pyroxenes where clinopyroxene has higher

TABLE 3.5

MICROPROBE ANALYSES OF CLINOPYROXENES IN SUB-ALKALINE ROCKS

Rock Type:	1	1	1	1	2	3	4	5
Sample No:	6577	6577	6577	6577	6700	6675	6692	6684
Phase:	C	G	G	Ropx	E	P	P	E
SiO ₂	51.63	50.77	52.56	51.78	52.92	52.32	51.81	52.12
TiO ₂	.86	.77	.37	.79	.15	.23	.19	.36
Al ₂ O ₃	1.73	2.35	.82	1.82	.74	1.48	1.78	1.53
Cr ₂ O ₃	.14	.61	-	.22	-	.14	-	.19
FeO	13.02	12.19	20.82	11.71	8.77	13.70	14.41	12.32
MnO	.22	.17	.35	.21	.24	.50	.73	.36
MgO	15.84	13.01	19.43	14.65	14.00	11.48	10.25	12.82
CaO	15.73	18.89	4.54	17.67	22.78	19.62	19.12	19.51
Na ₂ O	.29	.36	-	.36	-	.46	.68	.22
K ₂ O	-	-	-	.03	-	.07	.15	-
Total	99.46	99.12	98.89	99.24	99.60	100.00	99.12	99.43
Cation Content (6 oxygen atoms)								
Si	1.9411	1.9287	1.9873	1.9494	1.9827	1.9833	1.9890	1.9720
Al ^{IV}	.0589	.0713	.0127	.0506	.0133	.0167	.0110	.0280
Al ^{VI}	.0180	.0341	.0242	.0301	.0196	.0494	.0694	.0403
Ti	.0242	.0221	.0107	.0225	.0042	.0066	.0055	.0102
Cr	.0039	.0182	-	.0063	-	.0064	-	.0057
Fe	.4094	.3873	.6583	.3699	.2749	.4344	.4625	.3900
Mn	.0070	.0056	.0114	.0066	.0077	.0132	.0237	.0114
Mg	.8869	.7357	1.0952	.8217	.7800	.6485	.5864	.7234
Ca	.6339	.7688	.1838	.7124	.9143	.7970	.7866	.7908
Na	.0208	.0262	-	.0258	-	.0338	.0496	.0160
K	-	-	-	-	-	.0032	.0072	-
Total	4.0041	3.9990	3.9835	3.9953	3.9967	3.9925	3.9909	3.9878

$\frac{100\text{Mg}}{\text{Mg} + \text{Fe}^{2+}}$	68.4	65.6	62.5	69.0	73.9	59.9	55.9	65.0
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Total iron as FeO

1. Tholeiitic Andesite
2. Granodiorite
3. Microgranite
4. Rhyolitic Pitchstone
5. Quartz Diorite

- C = Cumulate
 G = Groundmass
 Ropx = Rim on orthopyroxene
 E = Even grained
 P = Phenocryst

TABLE 3.6

MICROPROBE ANALYSES OF ORTHOPYROXENES IN SUB-ALKALINE ROCKS

Rock Type:	1	2	3	3
Sample No:	6577	6676	6690	6692
Phase:	M	P	P	P
SiO ₂	54.30	53.31	51.34	51.07
TiO ₂	.22	.21	.18	.13
Al ₂ O ₃	2.22	1.64	.93	.78
Cr ₂ O ₃	.07	-	-	-
FeO	13.12	22.41	29.51	30.34
MnO	.10	.72	1.21	1.10
MgO	26.79	19.48	14.36	14.10
CaO	2.31	2.09	1.41	1.23
Na ₂ O	-	.14	.19	.18
K ₂ O	-	-	.16	-
Total	99.13	100.00	99.29	98.93

Cation Content (6 oxygen atoms)

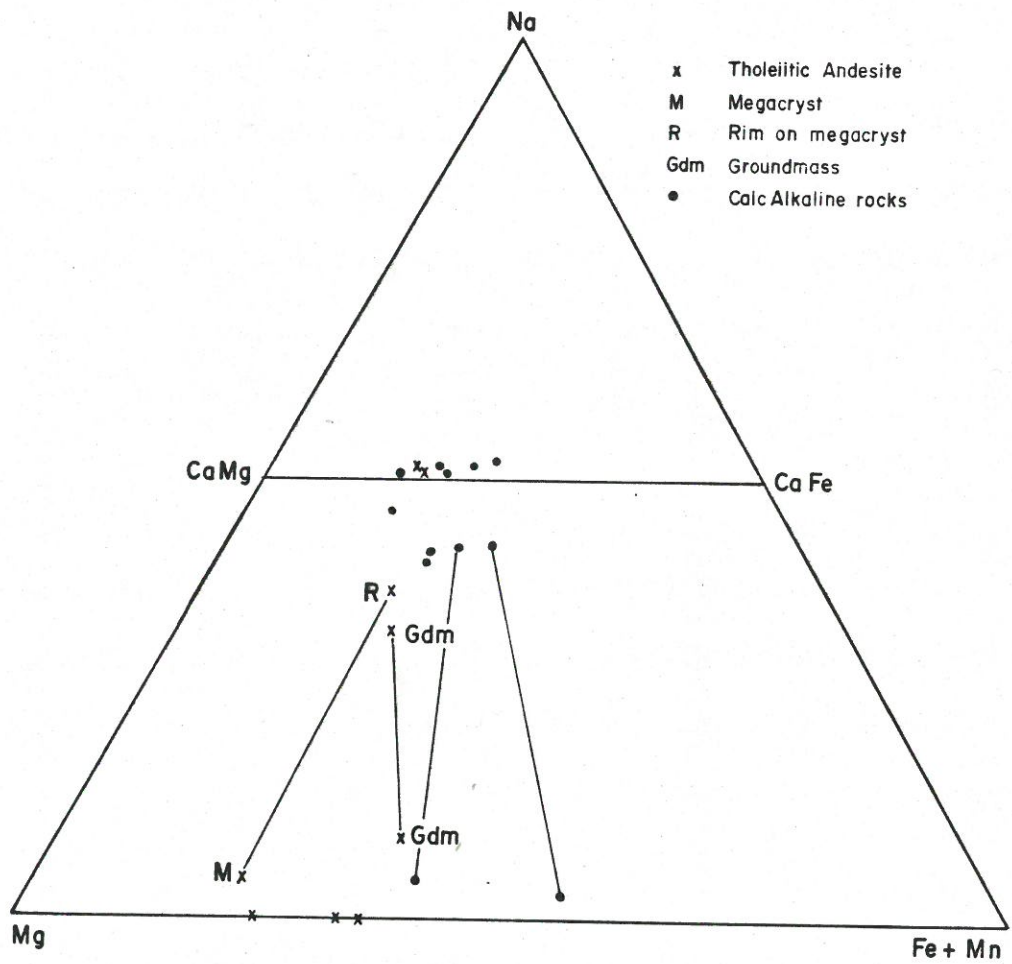
Si	1.9578	1.9918	2.0064	2.0087
Al ^{IV}	.0422	.0082	-	-
Al ^{VI}	.0521	.0639	.0428	.0362
Ti	.0058	.0059	.0053	.0040
Cr	.0020	-	-	-
Fe	.3956	.7003	.9647	.9981
Mn	.0031	.0228	.0401	.0364
Mg	1.4399	1.0843	.8366	.8267
Ca	.0894	.0836	.0590	.0520
Na	-	.0096	.0142	.0130
K	-	-	.0082	-
Total	3.9879	3.9704	3.9773	3.9751

$\frac{100\text{Mg}}{\text{Mg} + \text{Fe}^{2+}}$	78.5	60.8	46.4	45.3
---	------	------	------	------

Total iron as FeO

- | | |
|-------------------------|----------------|
| 1. Tholeiitic Andesite | M = Megacryst |
| 2. Microgranite | P = Phenocryst |
| 3. Rhyolitic Pitchstone | |

Fig. 3.4 Plots of clinopyroxenes and orthopyroxenes from tholeiitic and calc-alkaline rocks. Tie lines join coexisting clinopyroxenes and orthopyroxenes, and coexisting olivines are plotted on the Mg - (Fe + Mn) join.



Al_2O_3 than orthopyroxene, and concur with textural evidence indicating that crystallization of bronzite preceded that of augite. That the bronzite and coexisting clinopyroxenes do not represent an equilibrium assemblage is further indicated by distribution coefficients for Mg and Fe between pyroxene pairs which range from 1.62 to 1.90 and contrast with values of 0.65 - 0.86 determined for equilibrium pyroxene pairs in igneous rocks by Kretz (1961). The greatest degree of disequilibrium is between the bronzite megacrysts and the associated groundmass calcic clinopyroxene (1.90). $K_D(\text{Mg:Fe})$ between the bronzite and the clinopyroxene rim phase is 1.62 which is within the range for similar pairs from the Nuanetsi lavas (Cox and Jamieson, 1974). The distribution coefficient for Mg:Fe between the Ca-poor and Ca-rich groundmass clinopyroxenes is 0.89 which is only slightly higher than the range quoted above for equilibrium orthopyroxene-clinopyroxene pairs. Experimental studies by Green and Ringwood (1967) on basaltic compositions indicate that orthopyroxene is always more Mg-rich than coexisting olivine. Thus $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratios for the bronzite megacrysts of 78.5 indicate that they crystallised in equilibrium with the olivine megacrysts which have $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratios of 75.8. $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratios for clinopyroxenes and olivines in the gabbroic inclusions are 68.4 and 67.9 respectively, suggesting that these also represent an equilibrium assemblage which crystallised subsequent to the more Mg-rich bronzite and olivine.

As noted previously, Mg:Fe ratios indicate that the groundmass pigeonite-augite pair crystallised in equilibrium. In contrast, there appears to be some disequilibrium between pigeonite and the augite rim phase and the augite in the gabbroic inclusions. This suggests that crystallisation of the rim phase preceded that of the groundmass pyroxenes. Tarney (1969)

considered that pyroxene overgrowth relationships in dykes within the Lewisian gneisses had been facilitated by a high degree of mutual solid solution which reduced the crystallographic misfit along the c-axis. Evidence of solid solution between the Comboyne tholeiitic andesite bronzite crystals and the augite rim is apparent not only in the chemical composition of the two pyroxenes (Tables 3.5, 3.6), but also in the presence of incipient exsolution in the orthopyroxene which, in some instances, has resulted in the partial replacement of orthopyroxene by clinopyroxene.

Both textural and chemical data indicate that there are at least three generations of pyroxene crystallisation in the Comboyne tholeiitic andesite. Firstly, orthopyroxene and olivine megacrysts and cumulates crystallised at elevated pressures, and these were followed by the crystallisation of the gabbroic inclusions, and finally by the groundmass assemblage. The suggestion that orthopyroxene-olivine crystallised at elevated pressures is supported by experimental work undertaken by Cox and Jamieson (1974) on the Nuanetsi lavas. Their results indicate that equilibration of olivine and orthopyroxene in these olivine-rich melts occurred in a pressure range of 6-12 kb. Similarly, Nicholls and Lorenz (1973) report orthopyroxene and olivine crystallising together at 6 kb from a tholeiitic melt. It is concluded that the orthopyroxene-olivine megacrysts in the Comboyne tholeiitic andesite probably crystallised at a pressure of approximately 6 kb.

3.3.4 Calc-Alkaline Rocks

3.3.4.1 Textural Relationships

Clinopyroxene and orthopyroxene occur as elongate microphenocrysts (frequently corroded) in aggregates with plagioclase in the North Brother porphyritic microgranite, the pitchstones and some of the associated rhyolites. These pyroxenes are optically similar, exhibiting high relief, curved fractures (infilled with secondary minerals), and are commonly rimmed by green amphibole. Pleochroism in the orthopyroxenes is indistinct. In the Diamond Head quartz diorite and Middle Brother granodiorites pyroxene occurs as subhedral grains which, in most instances, have been completely or partially replaced by fibrous amphibole. Plates of euhedral and subhedral clinopyroxene occur in the Lorne Forest diorites, and complete replacement by green fibrous amphibole and chlorite is common.

3.3.4.2 Analytical Results

The rocks in which augite and hypersthene coexist as composite and discrete phenocrysts are strongly fractionated and markedly similar in composition, and have differentiation indexes (Thornton and Tuttle, 1960) in excess of 85.

Clinopyroxenes analysed from the North Brother porphyritic microgranite and the Kew pitchstone are chemically similar to high-level pyroxenes from other sub-alkaline associations and plot in a restricted field on a Ca - Mg - (Fe + Mn) diagram (Fig. 3.4). In contrast, orthopyroxene in the North Brother microgranite is distinctly richer in Mg relative to Fe than that in the Kew and Moorland pitchstones (19.48% against 14.10-14.36%). Similarly, Al_2O_3 , TiO_2 and CaO are higher in the more Mg-rich orthopyroxene, with MnO and Na_2O being slightly higher in the Fe-rich ortho-

pyroxene (Table 3.6)

The Diamond Head quartz diorite also contains clinopyroxenes chemically similar to those occurring in the North Brother microgranite and the pitchstones (Table 3.5). Clinopyroxene in the Middle Brother granodiorite has considerably lower Al_2O_3 and FeO and higher MgO and CaO than clinopyroxenes in the other granitic rocks in the Lorne Basin.

3.3.4.3 Discussion

Both clinopyroxene and orthopyroxene in the North Brother microgranite and the rhyolitic pitchstones are markedly enriched in Mg relative to the strongly fractionated host rocks, and similar Mg-enrichment in orthopyroxene has been noted in rhyolitic magmas from other calc-alkaline associations (Larsen et al., 1936; Carmichael, 1963b; Haslam, 1968; Ewart, 1971).

Kuno (1954) and Atlas (1952) stated that the Ca-content of orthopyroxene is generally related to the temperature at which it crystallised. While investigating the synthetic system MgSiO_3 - $\text{CaMgSi}_2\text{O}_6$ Atlas found that in enstatite the maximum number of Ca-atoms on the basis of six oxygens is 0.115 at 1100°C and 0.030 at 700°C . On this basis the orthopyroxenes from North Brother microgranite and the pitchstones with Ca varying from 0.0520 to 0.0836 would appear to have crystallised at a temperature in excess of 1000°C . More recently, Hensen (1973) determined experimentally that the $\text{Ca}/\text{Ca} + \text{Mg}$ ratio in orthopyroxene in equilibrium with clinopyroxene increases with temperature, decreases with pressure, and also may be affected by the $\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratio. Although Hensen does not give results for orthopyroxenes formed at pressures below 13.5 kb, $\text{Ca}/\text{Ca} + \text{Mg}$ values of 0.059 - 0.066 in the North Brother microgranite and the pitchstones are consistent with trends

determined for higher pressures and suggest that Hensen's results may be extended to much lower pressures.

3.4 Feldspars

3.4.1 Introduction

Feldspars are the major phenocryst and groundmass phases in the Comboyne and Lorne Basin alkaline and sub-alkaline rocks. Analyses and structural formulae of these feldspars are listed in Tables 3.7 - 3.9, and fractionation trends are shown in Figures 3.5 and 3.7.

The alkaline affinities of the Comboyne series are reflected by the composition of the feldspars in these rocks. The most calcic plagioclase in the hawaiites is sodic labradorite and both plagioclase and alkali-feldspar occur in the groundmass. Throughout the series, alkali-feldspar becomes increasingly more prominent with differentiation. In contrast, the plagioclase in the sub-alkaline rocks is comparatively more calcic and alkali-feldspar only becomes prominent in the strongly fractionated rock types.

The occurrence of peralkaline rhyolites in the Comboyne association gives an opportunity to assess further the significance of feldspar fractionation on the origin of these rocks, a subject that has been closely studied in recent years (Carmichael and MacKenzie, 1963; Bailey and Schairer, 1964; Thompson and MacKenzie, 1967; Bailey and Macdonald, 1969; Macdonald, 1969; Macdonald et al., 1970; Bailey and Macdonald, 1970; Upton et al., 1971).

In many instances the advanced alteration of feldspar phases in the Lorne Basin sub-alkaline rocks has precluded their accurate identification and analysis. Plagioclase is the major phenocryst phase in the Middle Brother and North Brother granitic rocks, the

Kew pitchstone, and in those rhyolites where residual multiple twinning enables positive identification of the feldspars. The majority of plagioclases in these calc-alkaline rocks exhibit strong zoning and a large number of analyses would be necessary to obtain a complete insight into feldspar relationships. As the ubiquitous alteration of these feldspars makes this impractical, feldspar analyses represent the composition of crystal cores except where otherwise stated.

3.4.2 Alkaline Rocks

3.4.2.1 Textural Relationships

At least three generations of feldspar crystallisation are represented in the Comboyne hawaiites. These rocks are characterised by an abundance of large feldspar crystals exhibiting great variation in morphology, and include:

Generation I

- 1) Subhedral and anhedral plagioclase up to 1.0 cm long, which poikilitically encloses clinopyroxene and orthopyroxene and shows variable patchy and normal zoning, fine deformation twinning and corrosion (Plates 3a,b).
- 2) Subhedral plagioclase in crystal aggregates with olivine and orthopyroxene (Plate 2a).
- 3) Euhedral and subhedral plagioclase crystals with strong sieving in the core, the rim, or throughout the grain (Plate 9a).

Generation II

- 1) Subhedral and anhedral plagioclase with abundant inclusions of pyroxene, Fe-Ti oxides and clear acicular apatite (Plate 10a).

- 2) Euhedral unsieved plagioclase phenocrysts with variable normal and oscillatory zoning (Plate 10b).

Generation III

- 1) Groundmass plagioclase, typically acicular with fine multiple twinning.
- 2) Alkali-feldspar as an interstitial groundmass phase and as rims around sieved and unsieved plagioclase phenocrysts.

The majority of the mugearites are essentially non-porphyritic, although there are rare phenocrysts of both plagioclase and alkali-feldspar. Both feldspars occur as unsieved and strongly sieved crystals, and are also the major groundmass phases with alkali-feldspar being more abundant than in the hawaiites.

Benmoreites typically contain glomeroporphyritic aggregates (up to 2.5 cm across), in which the major phenocryst phase is subhedral and anhedral calcic anorthoclase (Plate 13b). Plagioclase occurs in some aggregates. In many instances the anorthoclase shows evidence of resorption and inversion to microperthite suggesting some reaction took place prior to extrusion.

Phenocrysts and crystal aggregates of anorthoclase occur in the porphyritic anorthoclase trachytes, and rare euhedral and subhedral plagioclase phenocrysts are also present. Phenocrysts of alkali-feldspar are rare in the mafic trachytes. Alkali-feldspar is the only feldspar identified in the groundmass of the benmoreites and trachytes.

With two exceptions, feldspars analysed from the comendites are in the composition range anorthoclase/sanidine, plotting close to the minima on the albite-orthoclase join in the $\text{CaAl}_2\text{Si}_2\text{O}_8$ - $\text{NaAlSi}_3\text{O}_8$ - KAlSi_3O_8 system. Alkali-feldspar phenocrysts in the porphyritic comendites tend to be euhedral and plagioclase has only

been identified as crystal aggregates in the Mt. Oliver comendite and as larger zoned phenocrysts in the Mt. Coxcomb comendite.

3.4.2.2 Analytical Results

Plagioclase composition in hawaiites ranges from An_{27-53} (Table 3.7). This includes phenocryst, cumulate, (?) megacryst and groundmass phases. Strongly sieved phases have a similar range in composition to unsieved crystals. Groundmass plagioclases vary from An_{45-41} , and coexisting alkali-feldspar varies from Or_{39-32} . Feldspar phenocrysts in the mugearites include both plagioclase (An_{57-41}) and alkali-feldspar (Or_{31-27}), with groundmass phases being An_{41} and Or_{40-38} respectively (Tables 3.7 and 3.8).

Alkali-feldspar is the only feldspar analysed in the Comboyne benmoreites. Anhedral phenocrysts have the composition of calcic anorthoclase (Or_{30}) and the groundmass feldspar is sanidine (Or_{48}).

Feldspar phenocrysts analysed in the porphyritic anorthoclase trachytes are in the composition range Or_{43-26} . Groundmass feldspars in both the mafic and anorthoclase trachytes have limited variation with compositions around Or_{35} .

Alkali-feldspar compositions in the comendites range from Or_{47-34} . The plagioclase in the Mt. Oliver comendite is almost pure albite with the composition of Ab_{96} .

Plots of whole-rock normative feldspar for the alkaline series (Fig. 3.5a) are generally in accord with the actual feldspar mineralogy. Hawaiites and mugearites fall in the two-feldspar field, and the remainder of the rock types plot in the one-feldspar field of the An - Ab - Or ternary diagram. This concurs with the presence of both plagioclase and alkali-feldspar in the more basic members of the series, and the common occurrence of only a single

TABLE 3.2

MICROPORRE ANALYSES OF PLAGIOCLASES IN ALKALINE ROCKS

Rock Type:	Hawaiiite															
Sample No:	6498	6498	6498	6597	6498	6597	6572	6570	6571	6498	6498	6570	6498	6570	6570	6498
Phase:	C	M	M	C	P	P	P	P	P	S	S	S	G	G	R	R
SiO ₂	55.66	58.70	57.18	55.22	57.71	55.76	54.28	55.65	57.16	57.56	61.16	55.01	57.90	55.87	55.99	55.67
TiO ₂	.06	-	-	-	-	.10	.15	.06	-	-	-	-	.15	.16	.12	-
Al ₂ O ₃	27.29	25.37	26.74	28.12	25.48	26.66	28.80	27.26	25.13	25.32	23.84	26.69	25.49	25.74	27.41	26.79
FeO	.46	.39	.20	.40	.36	.29	.49	.44	.54	.53	.39	.50	.65	.89	.54	.40
MgO	.23	.15	.15	.36	.07	.17	.22	.13	.07	.13	.37	.11	.10	.20	.10	.09
CaO	9.29	6.92	8.27	10.09	8.22	8.61	10.71	9.63	8.22	7.82	4.91	9.54	7.62	8.36	9.86	8.89
Na ₂ O	5.83	6.93	6.26	5.63	6.44	6.11	5.51	5.55	6.18	6.39	7.67	5.56	6.38	6.02	5.37	5.77
K ₂ O	.71	1.17	.86	.39	.74	.77	.45	.58	.83	1.12	1.79	.70	1.13	.86	.61	.83
Total	99.53	99.68	99.66	100.21	100.92	99.47	100.61	99.30	99.13	98.87	100.13	99.11	99.43	99.10	100.00	99.44
Cation Content (32 oxygen atoms)																
Si	10.0952	10.5594	10.3086	9.9534	10.3527	10.2644	9.7841	10.0835	10.3451	10.4493	10.9218	10.1370	10.4489	10.3216	10.0797	10.2312
Ti	.0003	-	-	-	-	.0141	.0203	.0081	-	-	-	-	.0198	.0205	.0155	-
Al	5.8333	5.3826	5.6832	5.9731	5.5985	5.6833	6.1187	5.8224	5.5719	5.4164	5.0175	5.7546	5.4213	5.5061	5.8132	5.7012
Fe	.0594	.0581	.0290	.0597	.0533	.0442	.0732	.0663	.0807	.0804	.0585	.0763	.0997	.1359	.0819	.0604
Mg	.0503	.0403	.0385	.0967	.0190	.0461	.0578	.0353	.0188	.0367	.0991	.0291	.0267	.0545	.0256	.0248
Na	2.0495	2.4377	2.1863	1.9670	2.2134	2.1412	1.9256	1.8974	2.1084	2.1911	2.6557	1.9053	2.1745	2.0635	1.8265	1.9661
Ca	1.8050	1.3349	1.5978	1.9480	1.5801	1.6681	2.0578	1.8685	1.5925	1.5214	.9401	1.8559	1.4741	1.6265	1.9015	1.7197
K	.1650	.2697	.1969	.0909	.1695	.1772	.1028	.1319	.1905	.2586	.4088	.1626	.2599	.1997	.1397	.1919
Total	20.0865	20.0927	20.0408	20.0837	19.9865	20.0386	20.1503	19.9134	19.9079	19.9539	20.1015	19.9208	19.9250	19.9204	19.8856	19.8953
100 An An + Ab	48.3	36.8	43.6	51.3	42.8	45.2	53.3	50.5	43.8	41.8	27.3	50.2	41.2	44.9	51.9	47.4
Mol. Percent																
An	46.1	34.3	41.0	50.1	40.8	42.7	53.1	47.8	40.9	38.8	24.3	47.3	37.8	41.5	48.9	44.1
Ab	49.3	59.0	53.0	47.6	54.5	51.7	46.6	46.9	52.3	54.1	64.9	47.0	54.0	50.9	45.4	48.8
Or	4.2	7.1	5.1	2.3	4.4	4.5	3.2	3.4	4.9	6.6	10.6	4.1	6.7	5.1	3.6	4.9

C = Cumulate

M = Megacryst

P = Phenocryst

S = Sieved grain

G = Groundmass

R = Rim

TABLE 3.7 Cont'd

MICROPROBE ANALYSES OF PLAGIOCLASES IN ALKALINE ROCKS

Rock Type: <u>Mugearite</u>						<u>Trachyte</u>
Sample No:	6578	6578	6567	6567	6607	6665
Phase:	P	P	P	S	G	P
SiO ₂	56.23	53.85	56.68	58.86	57.42	59.65
TiO ₂	-	.13	-	-	.15	.14
Al ₂ O ₃	27.50	27.95	26.27	26.43	25.13	25.00
FeO	.28	.42	.44	.67	.82	.59
MgO	.14	.18	.13	-	.16	-
CaO	9.60	10.24	9.16	8.08	7.62	6.78
Na ₂ O	5.53	5.37	5.36	6.24	6.43	7.33
K ₂ O	.59	.54	.79	1.08	.97	.51
Total	99.87	98.68	98.83	101.36	98.70	100.00
Cation Content (32 oxygen atoms)						
Si	10.1121	9.8778	10.2912	10.4152	10.4457	10.6392
Ti	-	.0176	-	-	.0203	-
Al	5.8280	6.0434	5.6220	5.5120	5.3879	5.2555
Fe	.0421	.0652	.0668	.0986	.1258	.0888
Mg	.0359	.0492	.0333	-	.0423	-
Na	1.8789	1.9082	1.8370	2.0836	2.2105	2.4677
Ca	1.8512	2.0125	1.7803	1.5314	1.4861	1.2939
K	.1354	.1257	.1804	.2434	.2246	.1172
Total	19.8836	20.0996	19.8110	19.8842	19.9432	19.8623
<u>100 An</u>	50.4	52.8	50.1	43.2	41.0	35.2
An + Ab						
Mol. Percent						
An	47.6	50.8	45.4	40.1	37.8	33.6
Ab	46.8	45.4	45.3	52.8	54.4	62.0
Or	3.5	3.2	4.7	6.4	5.7	3.0

P = Phenocryst

G = Groundmass

S = Sieved Grain

TABLE 3.8

MICROPROBE ANALYSES OF ALKALI FELDSPARS IN ALKALINE ROCKS

Rock Type:	Hawaiiite			Mugearite					Benmoreite		Mafic Trachyte	Anorthoclase Trachyte		
Sample No:	6597	6572	6571	6567	6607	6578	6573	6567	6564	6601	6501	6627	6627	6627
Phase:	G	G	R	P	S	S	G	G	P	P	G	P	P	G
SiO ₂	65.86	64.61	64.81	64.62	63.78	65.13	65.30	65.54	63.09	62.59	65.29	63.75	65.95	65.43
TiO ₂	-	.13	.11	.12	.13	.14	.16	-	.16	.31	-	.16	-	.16
Al ₂ O ₃	19.50	20.66	19.33	21.72	20.63	19.92	19.87	20.30	22.67	22.97	19.98	21.77	19.27	19.47
FeO	.39	.98	.92	.39	.51	.39	.29	.37	.20	.19	.62	.11	-	1.17
MgO	-	.17	.12	-	.09	.08	.13	-	-	-	-	-	-	.10
CaO	.72	2.02	.79	2.90	2.10	1.45	.98	1.08	3.85	3.38	1.35	2.94	.36	.89
Na ₂ O	6.91	6.76	7.50	7.36	7.23	7.45	6.60	6.63	6.64	7.12	7.07	7.46	6.44	7.16
K ₂ O	6.32	5.55	4.94	3.90	4.46	4.87	5.88	6.41	4.07	4.27	5.47	3.68	7.03	5.64
Total	99.70	100.88	98.52	101.01	96.93	99.43	99.41	100.33	100.68	100.83	99.78	99.87	99.10	100.02
Cation Content (32 oxygen atoms)														
Si	11.8386	11.5334	11.7351	11.3987	11.5035	11.6722	11.7207	11.6888	11.1962	11.1482	11.6902	11.3607	11.8863	11.7200
Ti	-	.0174	.0151	.0160	.0176	.0195	.0224	-	-	.0414	-	.0211	-	.0208
Al	4.1309	4.3463	4.1258	4.5153	4.3852	4.2069	4.2023	4.2670	4.7422	4.8220	4.2157	4.5713	4.0911	4.1107
Fe	.0585	.1461	.1394	.0574	.0761	.0594	.0439	.0539	.0301	.0286	.0928	.0160	-	.1748
Mg	-	.0439	.0331	-	.0255	.0216	.0326	-	-	-	-	-	-	.0286
Na	2.4094	2.3411	2.5627	2.4526	2.4652	2.5213	2.3070	2.2319	2.2244	2.4585	2.3099	2.5089	2.1903	2.4187
Ca	.1385	.3860	.1537	.5480	.4049	.2775	.1897	.2057	.7327	.6453	.2593	.5611	.0691	.1709
K	1.4491	1.2639	1.1414	.8779	1.0262	1.1135	1.3446	1.4578	.9218	.9692	1.2503	.8358	1.6261	1.2872
Total	20.0250	20.0781	19.9063	19.8659	19.9042	19.8919	19.8622	19.9071	19.8474	20.1132	19.8982	19.8749	19.8629	19.9317
100 Or Or + Ab	38.9	36.4	31.2	27.0	30.1	31.4	37.6	40.3	29.9	29.5	35.1	25.6	43.4	35.5
Mol. Percent														
An	3.6	10.0	3.9	14.4	10.4	7.2	4.9	5.4	19.1	16.8	6.7	14.6	1.8	4.4
Ab	58.5	57.2	63.4	62.3	61.2	63.0	57.5	56.1	56.2	60.2	59.8	63.1	54.5	60.6
Or	37.3	32.8	29.2	23.0	26.3	28.8	34.7	37.9	24.0	25.2	32.3	21.7	41.8	33.3

P = Phenocryst

G = Groundmass

R = Rim

S = Sieved Grain

TABLE 3.8 Cont'd

MICROPROBE ANALYSES OF ALKALI FELDSPARS IN ALKALINE ROCKS

Rock Type:	Comendite								
Sample No:	6569	6646	6652	6652	6641	6647	6647	6636	6636
Phase:	P	P	P	P	E	P	G	P	R
SiO ₂	67.18	66.72	66.27	68.00	66.58	66.50	66.76	66.04	65.73
TiO ₂	-	-	-	-	-	-	-	-	-
Al ₂ O ₃	18.99	19.29	18.63	19.08	18.75	18.54	18.77	19.03	18.75
FeO	.38	.33	.38	.37	.65	.34	.31	-	.14
MgO	-	-	.12	-	-	-	.09	-	-
CaO	-	.12	-	.08	-	-	-	.31	.28
Na ₂ O	7.62	6.79	7.21	11.02	6.88	6.14	6.30	6.55	6.19
K ₂ O	5.59	6.99	6.36	.68	6.64	7.80	7.70	7.25	7.34
Total	99.76	100.24	98.97	99.23	99.50	99.32	99.93	99.18	98.43
Cation Content (32 oxygen atoms)									
Si	11.9589	11.8948	11.9447	11.9542	11.9872	11.9937	11.9639	11.9059	11.9399
Ti	-	-	-	-	-	-	-	-	-
Al	3.9849	4.0538	3.9580	3.9523	3.9771	3.9402	3.9633	4.0427	4.0133
Fe	.0563	.0496	.0578	.0543	.0973	.0513	.0468	-	.0223
Mg	-	-	.0323	-	-	-	.0230	-	-
Na	2.5628	2.2870	2.4529	3.6569	2.4015	2.0920	2.1335	2.2293	2.1253
Ca	-	.0234	-	.0154	-	-	-	.0602	.0550
K	1.2688	1.5896	1.4601	.1416	1.5240	1.7935	1.7604	1.6679	1.7006
Total	19.8317	19.8982	19.9058	19.7847	19.9871	19.8707	19.8909	19.9060	19.8564
100 Or Or + Ab	33.9	41.8	38.1	4.1	40.3	47.0	46.1	43.6	45.3
Mol. Percent									
An	.0	.6	.0	.4	.0	.0	.0	1.5	1.4
Ab	64.5	57.4	61.0	93.2	58.2	51.9	53.3	55.4	52.4
Or	33.0	41.3	37.6	4.0	39.2	46.1	45.5	42.8	43.4

P = Phenocryst

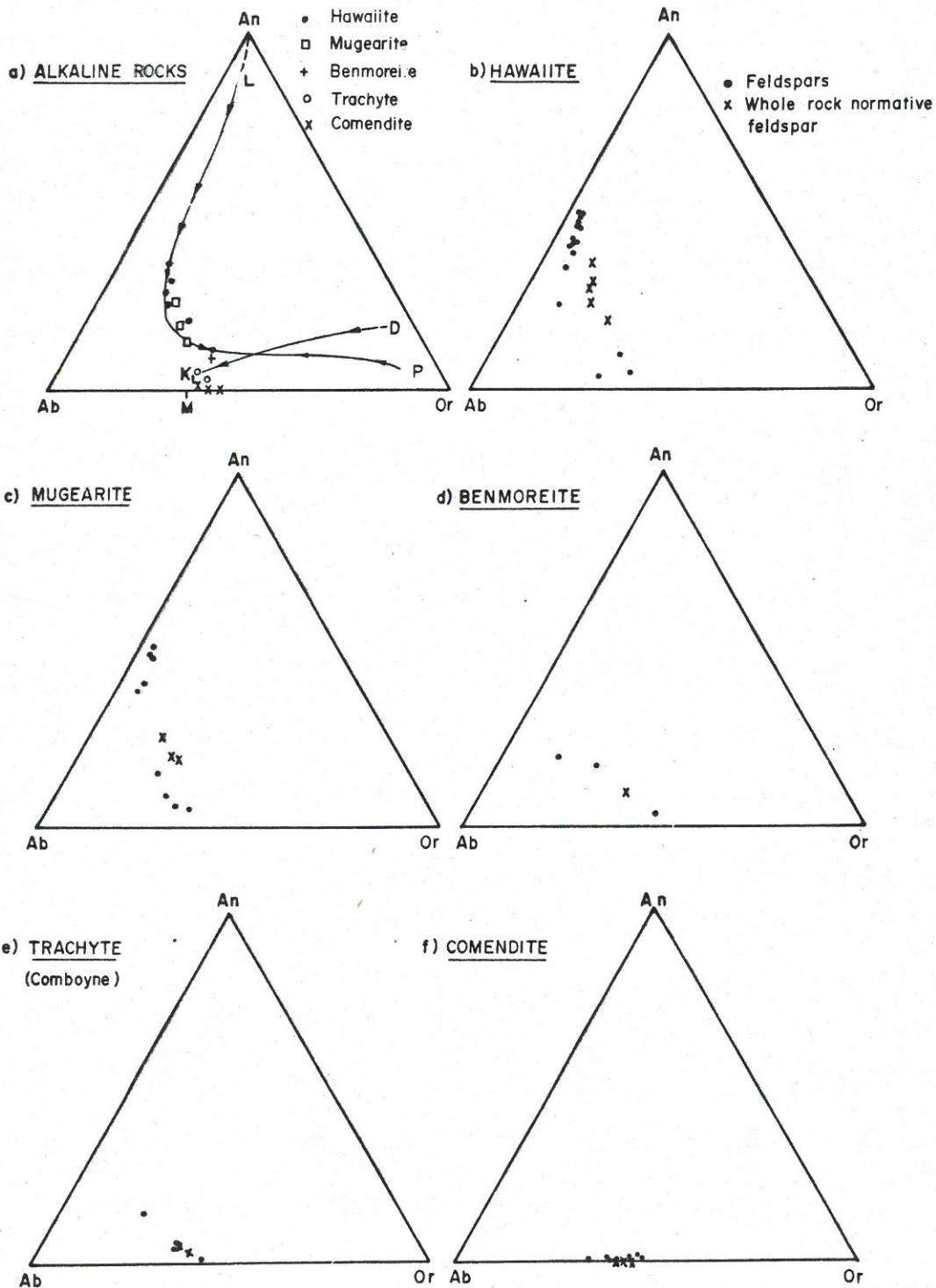
G = Groundmass

R = Rim

E = Even grained

Fig. 3.5a Whole rock normative feldspar plots for Comboyne alkaline rocks. L - P = approximate limit of solid solution under conditions expected in trachytic magmas (Tuttle and Bowen, 1958). $K_L - D$ = boundary between fields of potassium feldspar and plagioclase feldspar and represents the compositions of two feldspars whose compositions lie along the curve L - P. M = minimum of binary system Ab - Or (from Tuttle and Bowen, 1958).

Figs. 3.5b-f Plots of analysed feldspars and whole rock normative feldspar values for Comboyne hawaiites, mugearites, benmoreites, trachytes and comendites.



feldspar in the benmoreites, trachytes and comendites.

3.4.2.3 Discussion

The hawaiite feldspar assemblage is noteworthy for its textural variants with overlapping compositional ranges for the different textural types. It is, therefore, impossible to distinguish megacryst feldspar on chemical criteria alone, and it is only when feldspar occurs in cumulates with olivine, orthopyroxene and clinopyroxene, all exhibiting chemical compositions compatible with high-pressure origin, that valid evidence is provided for a deep-level origin of the feldspar.

Certainly, the great variation in feldspar texture and chemical composition is indicative of a complicated crystallisation history and the well developed zoning apparent in many crystals suggests equilibrium between crystal and melt was not maintained. Both plagioclase and alkali-feldspar were precipitated during late stages of crystallization, as shown by the occurrence of both as rims around phenocrysts and in the groundmass. However, the interstitial nature of the groundmass alkali-feldspar suggests that it may have been the last feldspar to crystallise. This is consistent with non-equilibrium crystallisation where early formed plagioclase is inhibited from reacting with the liquid, thus causing the liquid to leave the field boundary $K_L - D$ (Fig. 3.5a) and move towards the minimum (M) in the Ab - Or system, and resulting in the final crystallisation of a single feldspar (Tuttle and Bowen, 1958).

Phenocryst and groundmass plagioclases in mugearites have a similar compositional range to comparable feldspars in hawaiites. However, the more evolved nature of the mugearites is

reflected in the occurrence of alkali-feldspar as a phenocryst phase. It is interesting to note that when bulk normative compositions of the Comboyne alkaline rocks are plotted on a ternary An - Ab - Or diagram (Fig. 3.5a), megacryst-bearing hawaiites and mugearites are enriched in the alkali-feldspar component relative to those rocks in which megacrysts are sparse. This probably results from the abundance of plagioclase in the megacryst-rich specimens, as the plagioclase has a higher normative albite component and lower anorthite component than the bulk host rock composition.

Overall, the normative An - Ab - Or values for the alkaline series exhibit a smooth fractionation path and the feldspar phenocrysts define a crystallisation sequence consistent with that determined experimentally by Tuttle and Bowen (1958). The bulk composition of the porphyritic anorthoclase trachytes falls in the one-feldspar field in ternary An - Ab - Or diagrams, and the rare phenocrysts of plagioclase possible result from rapid cooling which has restricted equilibrium crystallisation.

Normative and actual feldspar plots of the peralkaline rocks show only minimal variation. Alkali-feldspars range from Or₄₇₋₃₄ and there is a concentration of normative values at Or₄₂₋₄₀ indicating this as the alkali-feldspar minimum.

Differentiation trends in peralkaline rocks can be examined using two projections devised by Bailey and Macdonald (1969). In a ternary diagram depicting molecular SiO₂, Al₂O₃ and Na₂O + K₂O, Bailey and Macdonald have shown that in any one series feldspar fractionation alone will result in a single trend radiating away from the feldspar point (F) (Fig. 3.6a). With the advent of crystallisation of another phase, for example, quartz, the trend away from the feldspar point will be interrupted

Fig. 3.6a Plots of Comboyne anorthoclase trachytes and comendites for molecular SiO_2 , Al_2O_3 and $(\text{Na}_2\text{O} + \text{K}_2\text{O})$ diagram (Bailey and Macdonald, 1969). K-K' and N-N' represent the quartz - potassium feldspar and quartz - albite cotectics respectively from the systems $\text{K}_2\text{O} - \text{Al}_2\text{O}_3 - \text{SiO}_2$ and $\text{Na}_2\text{O} - \text{Al}_2\text{O}_3 - \text{SiO}_2$ at 1 atmos. (Schairer and Bowen, 1955, 1956).

Fig. 3.6b Analyses of Comboyne comendites tied to their feldspar phenocrysts (in terms of Or/Or + Ab mol. per cent) plotted in a quadrilateral representing a plane in the volume $\text{Na}_2\text{O} - \text{K}_2\text{O} - \text{Al}_2\text{O}_3 - \text{SiO}_2$ (Bailey and Macdonald, 1969).

and will diverge from the original linear path.

When the Comboyne anorthoclase trachytes and comendites are plotted on this ternary diagram it is apparent that the comendites are only mildly peralkaline, and that although there is a distinct trend from the feldspar point (F) through anorthoclase trachytes to comendites, there is a pronounced clustering of plots in the vicinity of the quartz-albite cotectic in the system $\text{Na}_2\text{O} - \text{Al}_2\text{O}_3 - \text{SiO}_2$ at 1 atm (Schairer and Bowen, 1955 and 1956). It can also be seen that a number of specimens which have a typically peralkaline mineralogy (riebeckite/arfvedsonite, \pm aegirine, \pm aenigmatite), fall into the peraluminous field. These comendites diverge from the trend which follows a line away from the feldspar point, through the anorthoclase trachytes, shown by the remainder of the peralkaline rocks. There are two possible explanations for this divergence. Firstly, a clustering of plots with no definite correlation to feldspar fractionation may be anticipated for rocks originating from the partial melting of non-trachytic crustal material and Bailey and Macdonald (1970) used this criterion to distinguish 'continental' (anatectic) from 'oceanic' (fractionate) peralkaline associations. Secondly, secondary alteration could have resulted in the loss of alkalis. As the Comboyne comendites do show some evidence of following a crystallisation trend line and, additionally, are closely associated in the field with a mildly alkaline rock series which is considered to be parental, the interpretation favoured in this instance is that alkalis have been lost to varying degrees through magmatic or post-magmatic processes. The presence in rocks that plot in the peraluminous field of one or more sodic ferromagnesian minerals considered to typify peralkaline rocks, attests to the peralkaline affinities of the host melt and supports

the suggestion that alkali-loss has occurred in at least some of these comendites. Upton et al. (1971), in considering alkaline complexes in South Greenland have suggested that the build-up of alkalis in rocks is roughly proportional to their cooling rate. No indication of this process has been observed in the Comboyne association.

The linear trend away from the feldspar point (F) on the $\text{SiO}_2 - \text{Al}_2\text{O}_3 - (\text{Na}_2\text{O} + \text{K}_2\text{O})$ diagram shown by the anorthoclase trachytes and the comendites strongly suggests the comendites formed as the result of the extension of the fractionation trend through the anorthoclase trachytes by the removal of feldspar in the composition range Or_{35-25} (Upton et al., 1971). This is consistent with the dominance of anorthoclase (Or_{42-25}) as a phenocryst phase in these trachytes which have a bulk rock normative value of Or_{36} .

The second diagram devised by Bailey and Macdonald (1969) allows the alkali ratio to be shown diagrammatically (Fig. 3.6 b). In this projection, molecular Al_2O_3 (as percentage molecular $\text{SiO}_2 + \text{Al}_2\text{O}_3 + \text{Na}_2\text{O} + \text{K}_2\text{O}$) is plotted against $100\text{Na}_2\text{O}/(\text{Na}_2\text{O} + \text{K}_2\text{O})$ and when the host rock composition is linked to that of the feldspar phenocrysts on the Ab - Or join, the effect of the removal of these phases on the residual liquids can be determined. It is generally observed in peraluminous rocks that host rock and feldspars are tied together so that both become enriched in either Na_2O or K_2O with fractionation. In contrast, feldspars in per-alkaline rocks are generally enriched in K_2O relative to the melt (Bailey and Schairer, 1964; Thompson and MacKenzie, 1967; Bailey and Macdonald, 1969) so that fractionation results in further enrichment of Na_2O in the melt. The tie-lines between the

Comboyne comendite compositions and their alkali-feldspar phenocrysts indicate that there is only minor K-enrichment in the feldspars relative to the bulk composition of the host rock, suggesting that these comendites were never strongly peralkaline. It has to be remembered that sodic ferromagnesian minerals precipitate with feldspar in peralkaline rocks and if a significant quantity of these sodic ferromagnesian minerals co-precipitated with feldspar, then this would limit the amount of Na_2O available in the melt for feldspar crystallisation relative to that indicated by the bulk composition of the host rock. The plotting of K_2O against Na_2O (Fig. 4.1b) shows that the residual melt is depleted in both Na_2O and K_2O during the final stages of fractionation. During much of the crystallisation history of the Comboyne series there is strong enrichment in K_2O relative to Na_2O . However, at the stage of the crystallisation of the anorthoclase trachytes both Na_2O and K_2O show similar increases in values, and then later the precipitation of feldspar and sodic ferromagnesian minerals in the comendites rapidly depletes the melt in both these oxides.

3.4.3 Tholeiitic Rocks

3.4.3.1 Textural Relationships

The Comboyne tholeiitic andesite contains several generations of plagioclase (?)phenocrysts exhibiting contrasting textures. Euhedral and subhedral crystals show either no sieving or pronounced sieving to core and/or rim. Rare anhedral plates of feldspar have undergone complete disruption by sieving and appear to be reconstituted to granular plagioclase with fine twin lamellae (Plate 11a). Plagioclase is the major phase in the gabbroic aggregates in the tholeiitic andesite. It is euhedral, up to 1.00 mm long, and has

fine multiple twinning. Zoning is variable, ranging from weak normal to strong oscillatory and patchy zoning in adjacent grains. Groundmass plagioclase is acicular and finely twinned.

3.4.3.2 Analytical Results

The most calcic plagioclase analysed in the Comboyne tholeiitic andesite is the groundmass phase with the composition of An_{58} . Phenocryst plagioclase has a composition of An_{51} and that occurring in the gabbroic aggregates An_{56} (Table 3.9).

3.4.3.3 Discussion

The absence of alkali-feldspar in the groundmass, along with other criteria (Chapter 4), distinguishes the tholeiitic andesite from the mildly alkaline hawaiites with which it is closely associated in the field. The presence of (?)phenocryst plagioclase and plagioclase in the gabbroic inclusions with lower anorthite contents than the associated groundmass phase, supports the suggestion that crystallisation of feldspar, in addition to olivine, orthopyroxene and clinopyroxene, initially occurred at elevated pressures (Green, 1969).

3.4.4 Calc-Alkaline Rocks

3.4.4.1 Textural Relationships

The North Brother porphyritic microgranite has phenocrysts (up to 5 mm across) and crystal aggregates of corroded and sieved plagioclase. These plagioclases are zoned and strongly altered, and are commonly rimmed by an untwinned phase which exhibits extreme alteration and appears to be alkali-feldspar. The microgranitic groundmass consists of quartz and altered feldspar in which relict multiple twinning is apparent in some instances.

TABLE 3.9

MICROPROBE ANALYSES OF PLAGIOCLASES IN SUB-ALKALINE ROCKS

Rock Type:	1	1	1	2	3	4	4	4	4	5	5
Sample No:	6577	6577	6577	6700	6672	6690	6690	6692	6692	6666	6684
Phase:	C	P	G	P	S	P	S	P	P	E	E
SiO ₂	54.04	55.92	54.31	61.23	61.74	60.17	60.08	63.50	64.25	48.14	54.26
TiO ₂	.07	.07	.13	-	-	-	-	-	-	-	-
Al ₂ O ₃	28.23	27.12	28.27	24.21	23.50	24.01	24.39	21.92	22.04	33.01	28.46
FeO	.48	.71	.82	.22	.30	.18	.20	.18	.25	.38	.51
MgO	.23	.22	.31	.19	-	-	.09	-	-	.17	.13
CaO	10.74	9.86	11.40	5.54	4.88	5.69	6.12	3.18	3.51	16.16	11.42
Na ₂ O	5.01	5.60	4.74	8.08	7.78	7.54	7.32	7.60	8.13	2.37	4.94
K ₂ O	.39	.37	.27	.61	1.21	.80	.76	2.73	1.32	0.19	.28
Total	99.19	99.87	100.25	100.08	99.41	98.39	98.96	99.11	99.50	100.42	100.00
Cation Content (32 oxygen atoms)											
Si	9.8578	10.0822	9.8049	10.8939	11.0134	10.8538	10.7874	11.3536	11.3746	8.7963	9.8087
Ti	.0093	.0089	.0176	-	-	-	-	-	-	-	-
Al	6.0680	5.7668	6.0147	5.0753	4.9423	5.1039	5.1609	4.6190	4.5978	7.1093	6.0650
Fe	.0725	.1055	.1239	.0320	.0444	.0275	.0300	.0264	.0383	.0584	.0777
Mg	.0623	.0599	.0815	.0486	-	-	.0224	-	-	.0464	.0336
Na	1.7708	1.9068	1.6156	2.7872	2.6204	2.5685	2.4784	2.5677	2.7204	.8195	1.6834
Ca	2.0978	1.9068	2.2054	1.0559	.9323	1.1000	1.1770	.6100	.6654	3.1640	2.2130
K	.0911	.0860	.0609	.1380	.2739	.1831	.1734	.6219	.2987	.0445	.0636
Total	20.0296	19.9229	19.9245	20.0309	19.8267	19.8368	19.8295	19.7986	19.6952	20.0384	19.9450
$\frac{100 \text{ An}}{\text{An} + \text{Ab}}$	55.7	50.8	58.5	28.7	26.9	30.7	32.9	19.7	20.2	80.0	57.5
Mol. Percent											
An	53.3	48.9	56.5	27.5	24.2	28.2	30.3	15.8	17.4	80.1	56.6
Ab	42.4	47.4	40.1	68.3	65.8	63.8	61.9	64.3	68.8	20.0	41.8
Or	2.3	2.2	1.6	3.6	7.1	4.7	4.5	16.1	19.0	1.1	1.7

C = Cumulate

P = Phenocryst

G = Groundmass

S = Sieved

E = Even grained

1 = Tholeiitic Andesite

2 = Granodiorite

3 = Microgranite

4 = Rhyolitic Pitchstone

5 = Diorite

The Kew pitchstone has plagioclase phenocrysts and aggregates texturally similar to those in the North Brother microgranite, whereas feldspar phenocrysts in the Moorland pitchstone are commonly strongly sieved throughout resulting in a "brain" texture (Plate 12a). Euhedral plagioclase is less common. The fragmental appearance of many of the crystal aggregates in the North Brother microgranite and the pitchstones suggests the plagioclase and the associated orthopyroxene and clinopyroxene aggregates may come from disrupted xenoliths.

Plagioclase in granodiorites and granites forming Middle Brother, Juhle Mountain and other low hills to the north and west of Kendall, exhibits varying degrees of zoning and alteration to mica and clay minerals, and it is the dominant phenocryst and microphenocryst phase. In the texturally variable granophyres, one textural type has minor alteration, and euhedral phenocrysts of prismatic plagioclase are completely enclosed in a fine granophyric intergrowth of quartz and feldspar. A second textural type is similar to many of the markedly altered granitic rocks in the Middle Brother and Juhle Mountain area, in which large irregular phenocrysts of plagioclase occur within a groundmass of quartz and feldspar exhibiting typical microgranitic texture, which in some instances is replaced by a coarse granophyric intergrowth. The plagioclase sometimes appears to be rimmed by alkali-feldspar into which the granophyric intergrowth encroaches.

Plagioclase is the major mineral in the Diamond Head and Lorne Forest diorites, and occurs as large euhedral plates exhibiting strong oscillatory, normal and patchy zoning, and extensive fracturing. Inclusions of clinopyroxene are common. Extremely turbid interstitial material can only be qualitatively

identified as alkali-feldspar, and this is substantiated in areas where it is graphically intergrown with clear quartz in the Diamond Head quartz diorite.

Phenocrysts of euhedral, but commonly corroded, feldspar in dacites and rhyolites are mostly strongly altered to white mica and clay minerals. However, relict multiple twinning can be recognised in some instances. The groundmass feldspar in a number of specimens is clear and exhibits fine multiple twinning, suggesting the original feldspar has undergone albitization. The dacites and rhyolites consist essentially of feldspar and quartz, and the normative anorthite composition of 1 - 4% in these rocks is consistent with the feldspar being albite.

3.4.4.2 Analytical Results

The Middle and North Brother granites and the Kew pitchstone are similar in bulk composition and their plagioclase phenocrysts show a restricted composition range from An_{33-27} (Table 3.9). In contrast, the Moorlands pitchstone has zoned feldspars ranging from a sodic oligoclase core to a calcic anorthoclase rim. In the coarser granitic rocks, the occurrence of multiple twinning and granophyric intergrowths of quartz and feldspar indicate that both plagioclase and alkali-feldspar are present. This is also indicated by the whole rock normative feldspar values which fall in the anorthoclase field and are distinctly more sodic and potassic than the phenocrysts.

The Lorne Forest diorite contains an extremely calcic plagioclase (An_{80}). This contrasts with a host rock normative feldspar value of An_{47} , which is similar to that of the Diamond Head quartz diorite (An_{43}), in which analysed plagioclase has a composition of An_{58} .

3.4.4.3 Discussion

Figure 3.7a shows that all bulk rock normative feldspar values fall in the one-feldspar field. However, the ubiquitous strong zoning of plagioclase, along with instances of pronounced sieving, indicate that equilibrium was not maintained during crystallisation. This is confirmed by the occurrence of two feldspars in the silicic Lorne Basin rocks. If equilibrium has been maintained, the normative feldspar plots of the host rocks in the calcic anorthoclase field indicate the final precipitation of a single alkali-feldspar (Carmichael, 1963a). Carmichael notes that for rocks of this composition, where reaction between liquid and plagioclase is impeded, the degree of chilling determines whether a porphyritic pitchstone, rhyolite or granophyre ensues and cites as examples the Iceland and Rhum granophyres. The textural relationships observed in the high level Lorne Basin rocks are parallel with this suggestion and indicate that the microgranitic texture can be considered an extension of this series.

The common occurrence of granophyric intergrowths in a wide variety of the Lorne Basin sub-alkaline rocks raises the question of whether they have a replacement or cotectic crystallisation origin. Granophyric intergrowths of feldspar and quartz occur in the Diamond Head quartz diorite, North Brother porphyritic microgranite, and in a number of texturally dissimilar granitic rocks in the Middle Brother and Juhle Mountain area, including selvages between siliceous dykes and granodiorite host rocks.

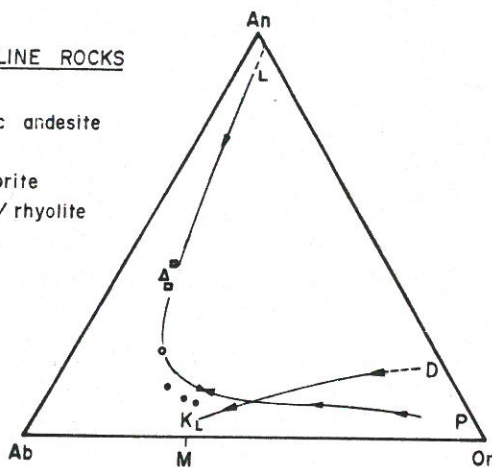
In some of these specimens there is strong evidence that the granophyric intergrowths are replacing feldspar rims and grains, as all stages of this process can be observed (Plates 12b, 13a). It is unfortunate that in these instances advanced alteration of the feldspars associated with granophyric intergrowths precludes

Fig. 3.7a Whole rock normative feldspar plots for the Lorne Basin - Comboyne Plateau sub-alkaline rocks. L - P = approximate limit of solid solution under conditions expected in trachytic magmas (Tuttle and Bowen, 1958). $K_L - O$ = boundary between fields of potassium feldspar and plagioclase feldspar and represents the composition of two feldspars whose compositions lie along the curve L - P. M = minimum of binary system Ab - Or (from Tuttle and Bowen, 1958).

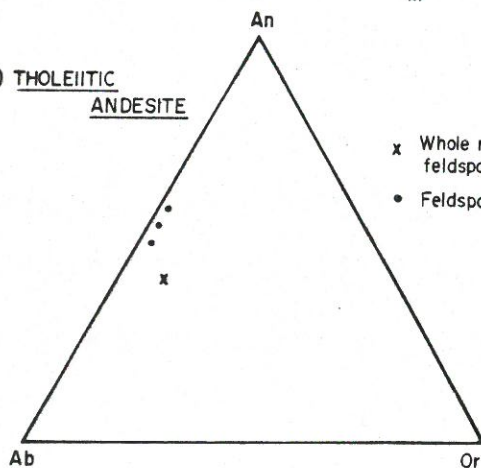
Figs. 3.7b-e Plots of analysed feldspars and whole rock normative feldspar values for tholeiitic andesite, diorites, granodiorites and granites/rhyolites.

a) SUB-ALKALINE ROCKS

- △ Tholeiitic andesite
- Diorite
- Granodiorite
- Granite / rhyolite

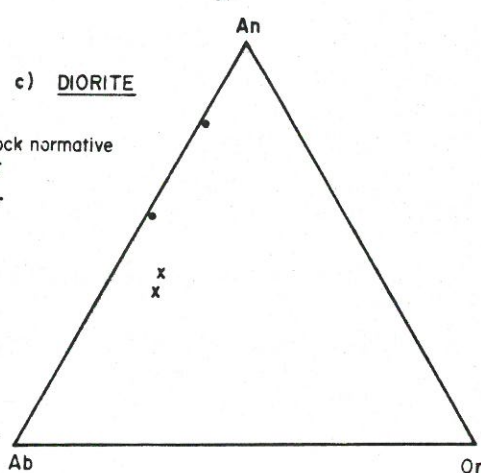


b) THOLEIITIC ANDESITE

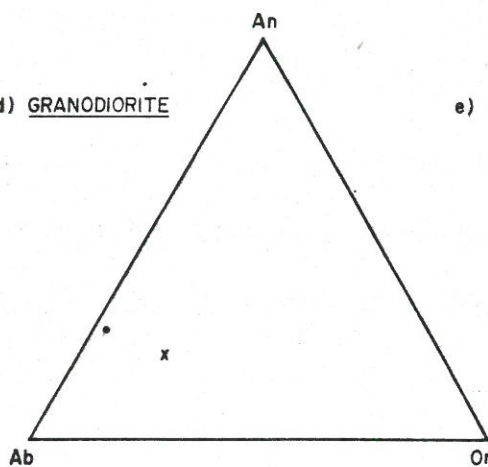


c) DIORITE

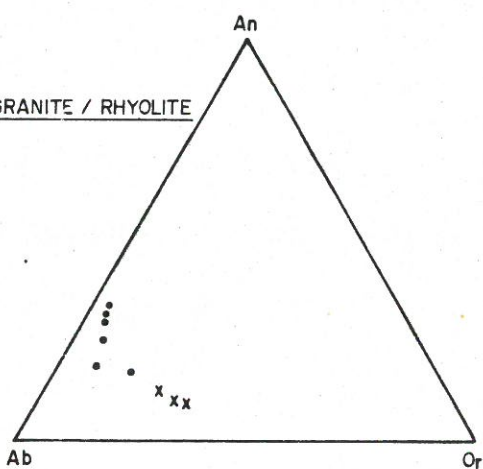
- x Whole rock normative feldspar
- Feldspar



d) GRANODIORITE



e) GRANITE / RHYOLITE



their accurate analysis. The occurrence of these intergrowths as a selvage between quartz-rich dykes in the Middle Brother area could result from one of two processes. They could have formed as the result of rapid crystallisation of a cotectic liquid, or alternatively, from the heating of the host rock by the dyke which may have caused a transformation of the original texture, assisted by hydrothermal activity. Although it is difficult to positively identify instances of cotectic crystallisation on textural evidence alone, textural relationships do indicate a replacement origin for some of the granophyric intergrowths in which quartz can be seen to be invading and replacing strongly altered feldspar.

Plagioclase in the Lorne Forest diorite is strongly enriched in calcium relative to the whole rock normative value and that for plagioclase in the Diamond Head diorite. The plagioclases in these rocks are strongly zoned and it is concluded that bulk feldspar analyses would result in compositions more closely approaching the normative whole rock value.

3.5 Opaque Minerals

3.5.1 Introduction

Fe-Ti oxides are important mineral phases in the Comboyne hawaiites, mugearites, benmoreites and trachytes, but exceedingly rare in the associated peralkaline rhyolites. In the Lorne Basin sub-alkaline rocks Fe-Ti oxides are ubiquitous but volumetrically minor. Sulphide minerals are rare in the alkaline rocks, being limited to rounded inclusions of pyrrhotite in silicate and oxide megacrysts in the hawaiites, and to minor pyrrhotite in the groundmass in hawaiites, mugearites, benmoreites and trachytes. In contrast, cubic and anhedral chalcopyrite and pyrite are

comparatively common in a number of the Lorne Basin dioritic, granitic and dacitic rocks, particularly those occurring at Diamond Head.

Although recent work on Fe-Ti oxides provides valuable data on the temperature, oxygen fugacity and, to a certain extent, composition of host melts at the time of their crystallisation (Buddington and Lindsley, 1964; Sato and Wright, 1966; Carmichael, 1967; Anderson, 1968b; Anderson and Wright, 1972; Wass, 1973; Prevot and Mergoil, 1973), studies on oxide phases in basaltic rocks are limited and there is a need for more complete chemical data in a wide variety of basaltic compositions.

In this study discrete and intergrown grains of titanomagnetite and ilmenite were analysed in order to determine the temperature and oxygen fugacity of the melts at the time coexisting Fe-Ti oxides equilibrated, and the effect of fractionation of these oxides and coexisting silicate phases on the evolution of the mildly alkaline, near-saturated to saturated Comboyne melts.

3.5.2 Terminology

The terminology adopted for the Fe-Ti oxides is as follows:

Titanomagnetite - an optically homogeneous phase within the series Fe_3O_4 - Fe_2TiO_4 (magnetite-ulvospinel) containing some Ti.

Ilmenite - a rhombohedral mineral with the composition FeTiO_3 and within the Fe_2O_3 - FeTiO_3 (haematite - ilmenite) series, with oxide stoichiometry close to R_2O_3 .

3.5.3 Recalculation of Analyses

The determination of the composition of the Fe-Ti oxides by electron microprobe methods results in problems of resolving the $\text{Fe}_2\text{O}_3:\text{FeO}$ ratio and at the same time obtaining some check on the accuracy of the analyses by deriving totals close to one hundred percent. The methods outlined by Anderson (1968b) for recalculation of titanomagnetites and ilmenites overcome both these problems and are followed in this study.

Magnetite - Ulvospinel Series

1. Al allotted to $\text{Mg}_{.7}\text{FeO}_{.3}\text{Al}_2\text{O}_4$
2. Remaining Mg allotted to Mg_2TiO_4
3. Mn calculated as Mn_2TiO_4
4. Ca and Cr calculated as CaFe_2O_4 and CrFe_2O_4
5. Fe_2TiO_4
6. Fe_3O_4
7.
$$\text{Usp} = \frac{\text{mole percent } \text{Fe}_2\text{TiO}_4}{\text{mole percent } \text{Fe}_3\text{O}_4 + \text{Fe}_2\text{TiO}_4}$$

Ilmenite - Haematite Series

1. Al and Cr allotted to Al_2O_3 and Cr_2O_3
2. Mg, Mn and Ca allotted to MgTiO_3 , MnTiO_3 and CaTiO_3
3. Remaining Ti allotted to FeTiO_3
4. Remaining Fe allotted to Fe_2O_3
5.
$$\text{Hm} = \frac{\text{mole percent } \text{Fe}_2\text{O}_3}{\text{mole percent } \text{Fe}_2\text{O}_3 + \text{FeTiO}_3}$$

All the Fe-Ti oxide analyses have relatively high values of SiO_2 , Na_2O and K_2O and these have been discounted and the respective values taken as zero. In light of analyses from other associations (Wilkinson, 1965; Carmichael, 1967; Wass, 1973), and the acceptable totals obtained after recalculation, this is considered to be justified, although some proportion of the SiO_2

value is most likely real. As the oxide values obtained by the non-dispersive technique of electron microprobe analysis are independent, the questionable results obtained for SiO_2 , Na_2O and K_2O should have no effect on the accuracy of the remaining oxides.

3.5.4 Alkaline Rocks

3.5.4.1 Textural Relationships

Titanomagnetite coexists with ilmenite as discrete and intergrown phases in the Comboyne hawaiites, mugearites, benmoreites and trachytes. Phenocrysts and microphenocrysts are not abundant and, although Fe-Ti oxides are important ground-mass phases in the series hawaiite to trachyte, there is a pronounced decrease in abundance in the anorthoclase trachytes and they are essentially absent in the comendites.

Hawaiites. At least three generations of Fe-Ti oxides are present in the Comboyne hawaiites. These are:

1. Megacrysts: Large corroded and embayed crystals of ilmenite range up to 4 mm in diameter, and are commonly associated with olivine, orthopyroxene and clinopyroxene. Globules of sulphides in these ilmenites are similar to those occurring in silicate megacrysts, but tend to be less rounded. Titanomagnetite intergrowths occur in these ilmenite megacrysts and have the following forms:
 - a) Continuous and discontinuous rims separating the ilmenite and groundmass. When lamellae of titanomagnetite are also present, these most commonly do not extend into the rim, but terminate within the ilmenite grain (Plate 14a).
 - b) Fine-scale rational and irrational lamellae.

c) Broad irrational lamellae .

Large corroded Fe-Ti oxide megacrysts have broad areas of titanomagnetite and ilmenite with curved interfaces between the two phases. Sometimes isolated granules of titanomagnetite occur within the ilmenite phase, close to and parallel to the titanomagnetite-ilmenite phase boundary (Plate 14b).

Rare corroded megacrysts of titanomagnetite, up to 2.5 mm across, occur as discrete crystals (sometimes with ilmenite rims and lamellae (Plate 16b) and are associated with corroded apatite up to 1.5 mm long (Plate 6b) and olivine. Similar apatites are observed with sieved clinopyroxene (Plate 6a), as inclusions in titanomagnetite, olivine and plagioclase, and as discrete crystals.

2. Phenocrysts: Cubic and skeletal phenocrysts of titanomagnetite up to 0.5 mm across are more abundant than the coexisting ilmenite phenocrysts which are generally smaller. The titanomagnetite phenocrysts commonly contain irrational lamellae and areas of ilmenite.

3. Groundmass Phases: Subhedral titanomagnetite is the major groundmass Fe-Ti oxide phase. It occurs as discrete grains and, less commonly, with intergrowths of ilmenite. Groundmass ilmenite is generally acicular and skeletal, with common 'swallow tail' terminations.

Mugearites. Ilmenite megacrysts have only been observed in the mugearite (MU6578) containing orthopyroxene, clinopyroxene and olivine megacrysts. This ilmenite occurs in cumulates with sieved clinopyroxene and plagioclase, is anhedral, has titanomagnetite lamellae, but is not rimmed by the spinel phase. The mugearites also contain rare subhedral and anhedral phenocrysts of

titanomagnetite (up to 0.25 mm) which have minor fine and coarse lamellae and areas of ilmenite. The broader lamellae are commonly irrational and generally terminate within the grain. The groundmass oxide phase consists predominantly of subhedral to anhedral titanomagnetite in which intergrowths of ilmenite are rare. Discrete ilmenite is not as abundant as magnetite and occurs as acicular laths and in vermicular intergrowths with pyrrhotite and silicates resulting from secondary alteration of ferromagnesian silicates.

Benmoreites. Large corroded (?) megacrysts of titanomagnetite (up to 3.00 mm across) show variable development of ilmenite intergrowths. Rare rounded crystals of ilmenite (up to 1 mm across) are similar to the ilmenite megacrysts in the hawaiites. The ilmenites contain rounded inclusions of pyrrhotite, have broad irrational lamellae and irregular rims of titanomagnetite, and occur as discrete grains and in crystal aggregates with olivine, anorthoclase, amphibole and striated apatite. Titanomagnetite is the predominant groundmass Fe-Ti oxide and commonly has very fine-scale irrational lamellae of ilmenite. Fine laths of ilmenite, generally skeletal in form, are less common. Ilmenite also occurs as secondary vermicular intergrowths similar to those occurring in the mugearites.

Trachytes. Rare phenocrysts and microphenocrysts of subhedral and anhedral titanomagnetite (up to 0.75 mm across) occur in the mafic and anorthoclase trachytes. These exhibit abundant fine-scale lamellae of ilmenite. Broader lamellae are less common. Subhedral to anhedral titanomagnetite in the groundmass shows minor development of ilmenite intergrowths and is more abundant in the mafic trachytes than in the anorthoclase trachytes. Discrete ilmenite has not been observed either as a phenocryst or groundmass

phase in the trachytes.

Comendites. Magnetite is extremely rare in the comendites and is commonly rimmed by aenigmatite.

3.5.4.2 Analytical Results

In order to determine the approach to equilibrium between the Fe-Ti oxides and the melt and, in turn, to estimate oxygen fugacity and temperature, both discrete and intergrown grains of titanomagnetite and ilmenite megacryst, phenocryst and groundmass phases were analysed where possible. Table 3.10 and Figure 3.8 show that in any one rock type there is generally little variation in composition between discrete and intergrown phenocrysts and groundmass phases, suggesting that a close approach to equilibrium has been attained. However, there is some compositional variation (Hm_{3-10}) between the large corroded megacrysts of ilmenite with titanomagnetite intergrowths found in the hawaiites, which may result from the differences in the proportion of titanomagnetite to ilmenite in these megacrysts. The apparent lack of equilibrium attained by these coexisting phases is consistent with their large size and is similar to the findings of Anderson (1968b). He reports that most titaniferous magnetites in the Tristan da Cunha alkaline basalts are strongly zoned and that intergrown ferrian ilmenite increases in haematite content away from the contact with titaniferous magnetite.

The major chemical differences between the hawaiite megacryst and coexisting phenocryst and groundmass Fe-Ti oxides are their Al_2O_3 and MgO contents. Although the Al_2O_3 and MgO values in the megacrysts are not as high as those reported by Aoki (1966), Lewis (1973), Wass (1973) for high pressure "spineliferous titanomagnetites" (up to 11% and 6% respectively), and ilmenite (up to

TABLE 3.10

MICROPROBE ANALYSIS OF TITANOMAGNETITES AND ILMENITES FROM ALKALINE ROCKS

TITANOMAGNETITES

Rock Type:	Hawaitite									Mugearite		Benmoreite		Trachyte
Sample No:	6493	6498	6597	6597	6597	6572	6572	6572	6572	6567	6567	6501	6501	6503
Phase:	M I	M I	G O	G O	G I	M I	P O	P O	P O	P I	G O	M O	G I	P O
TiO ₂	23.21	22.52	22.98	24.67	20.06	25.44	25.04	25.03	24.40	26.63	27.47	24.71	18.88	16.74
Al ₂ O ₃	2.79	4.11	.38	.51	1.06	2.75	1.51	2.10	1.52	1.50	.62	1.91	.26	.76
Cr ₂ O ₃	.32	-	.24	.28	.50	.24	1.04	1.46	.65	.29	-	-	.21	-
FeO	68.04	67.99	71.47	68.54	72.64	53.64	67.47	67.17	67.43	66.55	67.12	63.74	73.44	75.05
MnO	.37	.32	.42	.53	.45	.41	.46	.53	.41	.65	.63	.56	.54	.70
MgO	2.81	3.82	.21	.79	.58	2.70	1.43	1.96	1.50	1.25	.68	.89	.16	-
CaO	.19	-	.10	.15	.19	.18	.21	.17	.23	.09	.18	-	-	.24
Total	97.73	98.76	95.80	95.67	95.48	98.36	97.16	98.42	95.14	96.95	96.90	95.31	93.67	93.59
Mg ₇ Fe ₃ Al ₂ O ₄	2.12	3.03	.30	.46	.76	2.12	1.21	1.67	1.21	1.21	.61	1.52	.30	.61
Mg ₂ TiO ₄	4.82	6.58	.32	1.45	.80	4.66	2.41	3.37	2.57	2.09	1.12	1.23	.32	-
Mn ₂ TiO ₄	.67	.67	.67	.89	.67	.67	.67	.89	.67	1.11	1.11	.69	.69	1.11
CoFe ₂ O ₄	.65	-	.43	.65	.65	.65	.86	.65	.86	.43	.65	-	1.54	.86
FeCr ₂ O ₄	.22	-	.22	.22	.45	.22	.90	1.12	.45	.22	-	-	.22	-
Fe ₂ TiO ₄	57.47	53.22	63.28	66.63	54.33	63.95	65.96	64.40	63.95	70.43	74.24	65.41	61.43	45.84
Fe ₂ O ₃	32.65	33.89	36.58	27.55	39.83	26.63	25.93	26.63	27.32	22.23	20.38	27.79	41.91	40.39
Total	98.60	99.39	101.60	97.85	97.49	96.90	97.54	98.73	97.03	97.72	98.11	97.89	97.61	96.81
Usp.	63.77	59.72	63.37	70.75	57.70	70.60	71.78	70.75	70.07	76.01	73.46	70.50	55.13	48.60

ILMENITES

Rock Type:	Hawaiite								Mugearite			
Sample No:	6498	6597	6597	6597	6572	6572	6572	6498	6572	6567	6567	
Phase:	M I	G O	G O	G I	M I	P O	P O	G O	P I	P I	G O	
TiO ₂	51.99	49.49	49.73	49.34	51.24	50.00	50.54	50.82	50.80	52.03	50.25	
Al ₂ O ₃	.39	.26	.27	.33	.53	.32	.34	.29	.37	.29	.25	
Cr ₂ O ₃	-	-	.21	-	.13	.14	-	-	-	-	.52	
FeO	40.84	47.98	45.63	46.55	43.84	45.77	45.53	44.18	45.41	42.96	47.37	
MnO	.49	.54	.50	.57	.48	.48	.61	.53	.44	.61	.99	
MgO	4.28	.44	1.50	.98	3.16	1.77	1.61	2.51	2.21	1.60	.20	
CaO	.16	.15	.35	.35	.21	.33	.27	.24	.19	-	.25	
Total	98.15	98.66	98.20	98.12	99.59	98.81	99.00	98.57	99.42	98.54	99.74	
Al ₂ O ₃	.39	.26	.27	.33	.53	.32	.34	.29	.37	.29	.25	
Cr ₂ O ₃	-	-	.21	-	.13	.14	-	-	-	-	.52	
CaTiO ₃	.41	.41	.82	.82	.54	.82	.68	.54	.41	-	.54	
MgTiO ₃	12.74	1.32	4.45	2.89	9.38	5.29	4.81	7.45	6.61	5.05	.60	
MnTiO ₃	1.06	1.21	1.06	1.21	1.06	1.06	1.36	1.06	.91	1.36	1.56	
FeTiO ₃	81.19	92.11	86.60	83.02	83.77	86.35	86.02	85.44	86.80	91.20	92.11	
Fe ₂ O ₃	2.72	4.95	5.11	5.43	4.63	5.43	4.31	4.15	4.79	.96	4.15	
Total	98.51	100.26	98.72	98.70	100.04	99.41	99.52	98.93	99.89	98.86	100.14	
H ₂	3.24	5.10	5.56	5.81	5.24	5.92	4.67	4.63	5.23	1.04	4.31	
log ₁₀ f(O ₂)	-15.4	-12.5	-11.0	-11.8	-11.4	-10.5	-12.0	-	-	-	-11.5	
T ^o C.	825	940	1020	910	1000	1040	970	-	-	-	1010	

Total iron as FeO

End Members calculated by Anderson's (1966b) method

M = Mugearite

D = Discrete Phase

P = Phenocryst

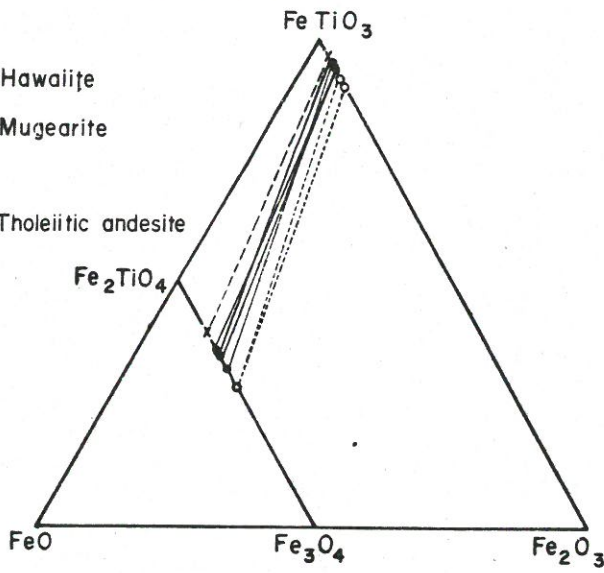
I = Intergrowth

G = Groundmass

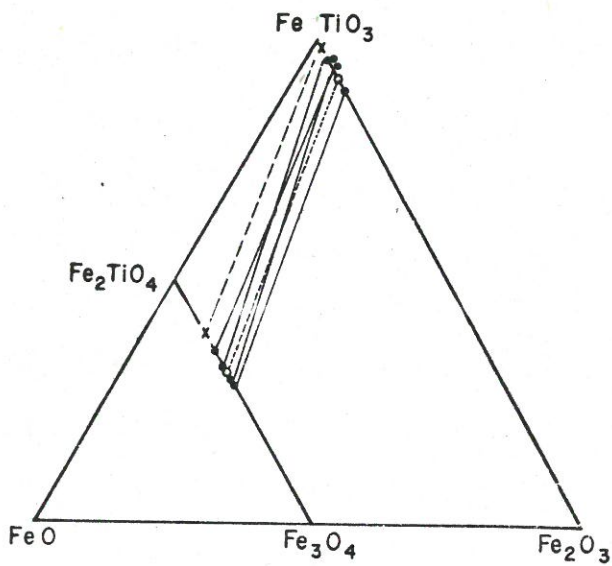
Fig. 3.8 Discrete and intergrown Fe-Ti oxide phases plotted on $\text{FeTiO}_3 - \text{FeO} - \text{Fe}_2\text{O}_3$ diagram showing degree of solid solution of these phases in magnetite - ulvospinel and haematite - ilmenite series.

Coexisting Discrete Phases

- - Hawaite
- x - Mugearite
- - Tholeiitic andesite



Intergrowths



1% and 8% respectively), they are significantly higher than the corresponding values in the phenocryst and groundmass phases (Table 3.10). The association of the Fe-Ti oxides with other megacryst phases which have textural and chemical characteristics indicative of a high-pressure origin supports the suggestion that Al_2O_3 content in these oxides may be pressure-dependent. The same may apply to MgO as Wass (1971) reports discrepancies between the actual MgO values in high-pressure Fe-Ti oxides phases from the Southern Highlands of New South Wales and those predicted by Lovering and Widdowson (1968) using solid-solid reactions. Similar differences have been found in the present study. The lower Al_2O_3 and MgO content in the Comboyne Fe-Ti megacrysts compared with oxide megacrysts from other basaltic associations possibly results from their crystallisation at a lower pressure. However, Carmichael and Nicholls (1967) and Prevot and Mergoil (1973) report a decrease in Al_2O_3 content in Fe-Ti oxides with increased evolution of the host rock which suggests that Al_2O_3 content may also be a function of temperature and/or melt composition.

Both chemical and textural criteria can be used to estimate the crystallisation interval in which oxidation of the ilmenite megacrysts occurred. In all instances Al_2O_3 is higher in the coexisting titanomagnetite than in ilmenite (2.06-2.79% and 0.39-0.59% respectively), whereas MgO is generally higher in the rhombohedral phase (2.62-3.82% compared with 3.16-4.76%). These trends are similar to the relationships observed between coexisting ilmenite and magnetite in other basaltic associations. Thus distribution relationships of Al_2O_3 and MgO between ilmenite and magnetite are consistent with their having crystallised simultaneously. However, textures in which large corroded ilmenite grains are rimmed by titanomagnetite and which contain titanomagnetite

lamellae generally terminating within the grain (Plate 14a) and compositionally identical to the rim material, suggest that both the rims and lamellae result from subsolidus reaction rather than direct precipitation from the melt, and that this reaction occurred after the onset of corrosion. The compositions of the respective ilmenite - titanomagnetite phases indicate that this subsolidus reaction was an oxidation reaction, as the proportion of Fe^{3+} in the magnetite is greater than that in the coexisting ilmenite. It would therefore appear that even in subsolidus re-equilibration Al_2O_3 is preferentially incorporated in the spinel phase. Alternatively, it is possible that the secondary titanomagnetite also formed at elevated pressures (but lower than the pressure at which ilmenite formed) and that the relatively high Al_2O_3 values in the titanomagnetite result directly from this. This is supported by the occurrence of strongly corroded titanomagnetite megacrysts which crystallised after the ilmenite megacrysts as evidenced by textural, chemical and experimental criteria.

Textural criteria include the occurrence (a) of ilmenite associated with olivine, clinopyroxene and orthopyroxene which show compositional characteristics indicating precipitation at the highest pressures represented by observed megacryst/cumulate assemblages and (b) of the corroded magnetite grains associated with phases whose compositional characteristics indicate precipitation under a lower pressure regime than that for the earlier ilmenite-bearing assemblages. In addition, apatite megacrysts are exclusively associated with magnetite-bearing assemblages. The order of crystallisation of apatite can be deduced from observations of its inclusion within titanomagnetite, olivine and plagioclase, but never within ilmenite. Chemical criteria are based on the difference in Mg:Fe ratios of the

olivines. The Mg:Fe ratios of olivines associated with magnetite and apatite are lower and hence later in the crystallisation sequence (Green and Ringwood, 1967). Experimental criteria derive from experimental runs (Chapter 5) wherein no magnetite has been identified. Runs at 6.5 kb with 2% H₂O show clinopyroxene as the liquidus phase, followed by the crystallisation of olivine, orthopyroxene, ilmenite and (?)apatite.

The distribution relationship of Al₂O₃ and MgO in the megacrysts is also found in the phenocryst and groundmass phases. There is also a marked decrease in minor element concentration between phenocryst and groundmass phases while there is little variation between these phases (and megacrysts) in the major elemental components of Fe and Ti. This also supports the suggestion that Al₂O₃ and MgO contents are to some degree temperature and/or melt composition dependent.

Haematite content in the hawaiite ilmenite megacrysts varies from 10.41% to 3.24%, whereas in the phenocryst and groundmass phases it shows only limited variation from 5.81% to 4.63%, including both discrete and intergrown types. The magnetite ulvospinel component varies from 59.72% to 70.60% in the megacrysts and from 57.10% to 71.85% in the phenocrysts and groundmass. The ulvospinel content of titanomagnetite in the mugearites is higher and the ilmenite haematite content lower than in the hawaiites, which is consistent with a decrease in oxygen fugacity from the hawaiites to mugearites. The large corroded crystals of titanomagnetite in the benmoreites have a similar ulvospinel content as the titanomagnetite occurring in the hawaiite megacrysts. In contrast, the ulvospinel content of the groundmass titanomagnetite is significantly lower (Table 3.10). This supports the suggestion that the cumulate phases in the benmoreites (olivine, anorthoclase,

amphibole, titanomagnetite, apatite) crystallised at elevated pressures, but not as high as the hawaiite megacryst/cumulate assemblage.

The occurrence of ilmenite megacrysts in the Comboyne hawaiite and the absence of high-pressure spinel phases (excluding the titanomagnetite megacrysts which are considered to have crystallised at less than 6 kb (page 78)) are consistent with observations by Binns et al. (1970) and Wass (1971) that high-pressure spinel is restricted to highly undersaturated alkaline basaltic compositions. Emslie and Lindsley (1969) showed that at high pressures the spinel field in the system Ab - An - Fo is drastically reduced by the addition of silica. Therefore, with an adequate degree of silica-saturation in natural basalts, potential spinel would react with the melt to produce aluminous orthopyroxene (Binns et al., 1970). Certainly, observations to date on natural rocks support this conclusion.

3.5.4.3 $f(O_2)$ and Temperature

Buddington and Lindsley (1964) in a series of phase equilibria studies on the system $FeO - Fe_2O_3 - TiO_2$ determined curves representing the temperature and oxygen fugacity of formation of coexisting pairs of titaniferous magnetite and ilmenite, and found that the equilibrium composition of coexisting Fe-Ti oxide pairs is dependent on oxygen fugacity and temperature. In turn, oxygen fugacity is a function of the following controls which have been fully reviewed by Carmichael and Nicholls (1967) and Wass (1973):

1. The composition, chemical relationships and the extent of melting of phases present at the site of magma generation.
2. Variations in temperature and pressure from the site of

magma generation to final crystallisation.

3. Variations in melt composition resulting from additions and losses prior to eruption.

The first Fe-Ti oxide to crystallise in the Comboyne hawaiites was ilmenite and, as already noted, textural relationships indicate it precipitated almost simultaneously with olivine, orthopyroxene and clinopyroxene.

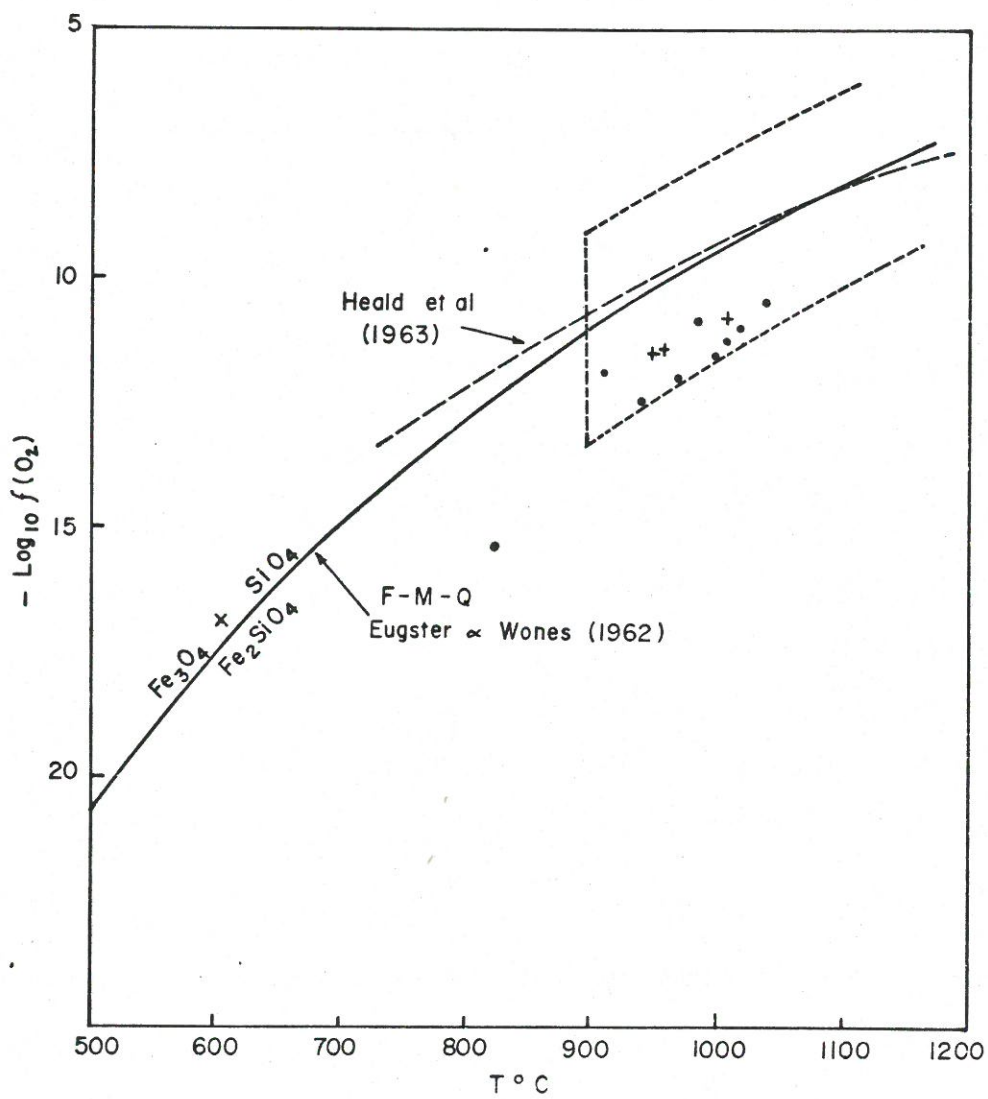
The crystallisation of megacryst ilmenite ahead of magnetite contrasts with the normally observed trend in alkali olivine basalts where titaniferous magnetite is the first oxide to crystallise. Ferrian ilmenite is generally recognised as the first oxide to form in tholeiitic basalts (Anderson, 1968a; Wass, 1973), and it would therefore appear that in this respect the mildly alkaline-transitional Comboyne series shows greater affinity to tholeiitic than alkali olivine basalts.

Despite the variation in composition between ilmenite - magnetite pairs in these megacrysts, the oxygen fugacity and temperatures (using Buddington and Lindsley's (1964) curves for coexisting oxides) determined for megacryst ilmenite - magnetite do show a range of values lying on a smooth curve with oxygen fugacity falling with temperature (Fig. 3.9).

Ilmenites similar to those occurring as megacrysts in the Comboyne rocks are described by Anderson (1968b) in trachybasalts from the Tristan da Cunha association. The tendency for these large corroded ilmenite crystals to be rimmed by titanomagnetite indicates that with a changing physico-chemical environment the ilmenite became unstable in the host melt. This interpretation is consistent with phase equilibria studies at 1 atm in air of the MgO - Iron Oxide - TiO_2 system (Woermann et al., 1969) which show there are large primary phase regions of spinel and pseudobrookite on the liquidus surface, whereas primary phase areas of the rhombo-

Fig. 3.9 Oxygen fugacities ($-\log_{10} \text{ atm.}$) plotted against temperature ($^{\circ}\text{C.}$) for Comboyne alkaline and tholeiitic rocks. Solid circles = alkaline rocks, crosses = tholeiitic andesite.

F - M - Q is the quartz - magnetite - fayalite buffer curve (Eugster and Wones, 1962). The long dashed curve is the calculated equilibrium oxygen fugacity of analysed volcanic gas (Heald et al., 1963). Area partly enclosed by small-dashed line represents the field for basalts deduced by Carmichael and Nicholls (1967).



hedral phases are restricted to only two small regions adjacent to the $\text{MgO} - \text{TiO}_2$ and Iron Oxide - TiO_2 boundary lines.

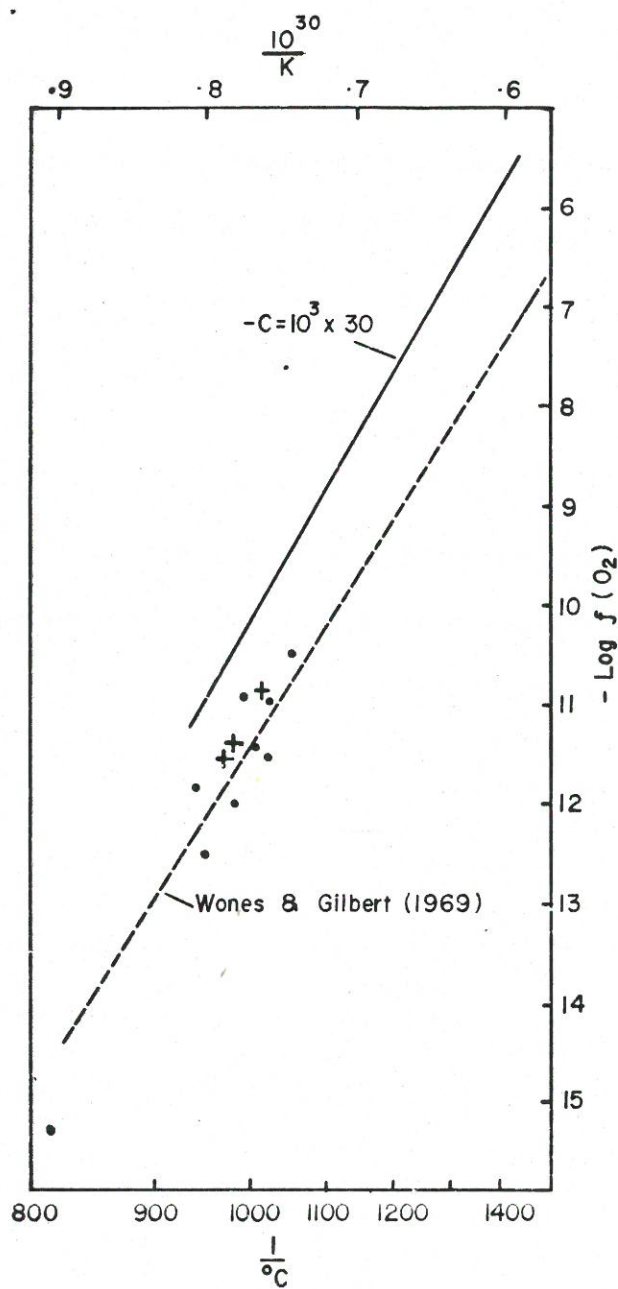
The initial precipitation of ilmenite, followed by its reaction to give ilmenite-magnetite intergrowths, indicates that subsequent to the formation of ilmenite there was either 1) an increase in temperature, 2) an increase in oxygen fugacity, or 3) a combination of these two processes. Textural relationships and experimental results (Chapter 5) indicate that at approximately 6 kb olivine and clinopyroxene were the first minerals to crystallise, with ilmenite, orthopyroxene and plagioclase appearing slightly later. Coombs (1963) considered that initial $\text{Fe}_2\text{O}_3:\text{FeO}$ ratios in alkali-basalts would be a maximum of 0.15. Allowing for the more fractionated nature of the Comboyne hawaiites, a $\text{Fe}_2\text{O}_3:\text{FeO}$ upper limit of 0.30 has been adopted in this study. $\text{Fe}_2\text{O}_3:\text{FeO}$ ratios in olivines are less than 0.01, and Anderson (1968b) quotes a value of 0.03 for orthopyroxene and 0.1 for augite. The range of $\text{Fe}_2\text{O}_3:\text{FeO}$ ratios for Fe-Ti oxides recorded by Buddington and Lindsley (1964), Fudali (1965), Sato and Wright (1966) and Carmichael (1967) varies from 0.1 to 0.25 for ilmenite and 0.1 to 0.37 for magnetite. From these values it is apparent that an excess of olivine, clinopyroxene and orthopyroxene crystallisation over that of ilmenite would result in an increase in the $\text{Fe}_2\text{O}_3:\text{FeO}$ ratio of the melt, and if the temperature remains constant oxygen fugacity would also increase. Because of the smooth curve relationships (Fig. 3.9) of oxygen fugacity and temperature, despite the variation in individual compositions, it is concluded that variation in temperature is the primary factor in the calculated increase in oxygen fugacity in the hawaiites after the initial crystallisation of ilmenite megacrysts, and that variation in melt composition resulting from crystal fractionation was

probably only a minor factor. One way of accounting for this increase in temperature is decrease in pressure as magma moved upward through the crust. Figure 3.10 shows that when $-\log_{10} f(O_2)$ is plotted against $1/C^0$ for rocks in which equilibrium between Fe-Ti oxides and melt has been approached, these plots form a smooth trend almost coinciding with the quartz-magnetite-fayalite line (Wones and Gilbert, 1969), and the theoretical slope of $C = -10^3 \times 30$ calculated from data of Shibata (1967) and Darken and Gurry (1946). The slope of this line describes the change in oxygen fugacity for a given change in temperature when the ratio of $FeO_{1.5}$ and FeO are held constant. The plots of the Comboyne rocks parallel to the $C = -10^3 \times 30$ line supports the conclusion that temperature has been the most significant factor in accounting for change in oxygen fugacity.

The suggestion that the Comboyne magmas have behaved as closed, internally buffered systems is also supported by data from other associations where similar conclusions have been made from co-equilibrated Fe-Ti oxides from basic and silicic lavas, including alkali olivine and tholeiitic basalts (Carmichael, 1967; Carmichael and Nicholls, 1967; Anderson, 1968b; Wass, 1973). Nash and Wilkinson (1970) came to the same conclusion using silicate-oxide reactions, as did Heald et al. (1963) and Sata and Wright (1966) using volcanic gases.

The oxygen fugacity determined for the Comboyne hawaiites varies from $10^{-15.4}$ to $10^{-10.5}$ atm. at temperatures ranging from $825^\circ C$ to $1040^\circ C$. These fugacity values are comparable with those reported in other igneous associations (Carmichael, 1967; Wass, 1973), and the temperatures are consistent with experimentally determined values for the appropriate phases crystallising (Chapter 5).

Fig. 3.10 Oxygen fugacities ($-\log_{10} \text{atm.}$) plotted against temperature ($10^3/T^\circ\text{K}$) for Comboyne alkaline and tholeiitic rocks. Solid circles = alkaline rocks, crosses = tholeiitic andesite. Unbroken line represents theoretical slope of $C = 10^3 \times 30$ calculated using the data of Shibata (1967) and Darken and Gurry (1946), see page 83. The dashed line is the experimentally determined curve of Wones and Gilbert (1969) for the system magnetite - quartz - fayalite. (After Wass, 1971).



Fudali (1965) in a series of experiments in which the Kilauean basalts were equilibrated at a constant temperature of 1200°C and an original known $\text{Fe}_2\text{O}_3:\text{FeO}$ ratio, found a maximum range of $f(\text{O}_2)$ of $10^{-9.1}$ to $10^{-6.5}$ atm and noted a marked tendency towards an increase in oxygen fugacity with increase in the silicic nature of the host. These results are consistent with values determined from Buddington and Lindsley's (1964) curves in which the $f(\text{O}_2)$ and temperatures for the fractionated lavas of the Kilauea volcano range from $10^{-10.2}$ to $10^{-7.6}$ atm. and 1050° to 1180°C respectively (Anderson and Wright, 1972). The tendency towards higher $f(\text{O}_2)$ in Fudali's study has been attributed by Carmichael (1967) and Wass (1973) to the oxidation of the samples used in the study resulting in the experimentally determined $f(\text{O}_2)$ being higher than if unoxidised material had been used.

Sato and Wright (1966) measured directly the $f(\text{O}_2)$ of Hawaiian tholeiitic basalts in lava ponds and report a value of $10^{-9.6}$ atm. at 1000°C ., and an extrapolation of their results gives a value of $10^{-8.4}$ at 1200°C . which compares with Fudali's experimental value of $10^{-8.2}$ atm. at the same temperature. Using Buddington and Lindsley's curves, Carmichael (1967) found that silicic rocks from a number of areas, including Thingmuli and California, range from $10^{-13.9}$ to $10^{-9.1}$ atm. at temperatures varying from 815° to 1010°C . Similarly, Wass (1973) reports $f(\text{O}_2)$ for alkaline basalts from the Southern Highlands of N.S.W., of $10^{-13.0}$ to $10^{-8.0}$ atm. at temperatures varying from 940° to 1200°C . In most of the examples cited above, lower $f(\text{O}_2)$ values correspond to lower temperatures.

The $f(\text{O}_2)$ and temperatures determined for the Comboyne alkaline rocks are similar to those reported by Carmichael (1967) and Wass (1973), and differences can be attributed to variations in silica

content, $\text{Fe}_2\text{O}_3:\text{FeO}$ ratio of the host melt, total alkalis (in particular K), and the point in the crystallisation history of the melt represented by the equilibrated Fe-Ti oxides.

Oxygen fugacity and temperature determined for equilibration of coexisting oxide phases for the Comboyne mugearites are within the range determined for the associated hawaiites.

3.5.5 Tholeiitic Rocks

3.5.5.1 Textural Relationships

Subhedral and skeletal laths of ilmenite are the most abundant groundmass oxide phases in the Comboyne tholeiitic andesite, and discrete grains of magnetite are similarly mostly subhedral and elongate. Fe-Ti oxide phenocrysts have not been observed. The ilmenite grains tend to have broad intergrowths of titanomagnetite which commonly forms discontinuous rims around the ilmenite, either as a solid rim or as dendritic branches growing out from euhedral boundaries (Plate 15a). These dendritic outgrowths indicate that the titanomagnetite has precipitated from the melt rather than forming by reaction with the ilmenite.

Dendritic and skeletal growth results in a large surface area and represents a high energy growth form. This is indicative of a crystallisation environment with high temperature gradients, supersaturation or undercooling of the melt, coupled with a limited number of nuclei, and slow diffusion rates of the critical ions (Rice et al., 1971; Wass, 1973). Spry (1969) suggested that continued crystallisation will eventually culminate in infilling between dendrites resulting in an euhedral crystal. Varying degrees of this process are observed in the Comboyne tholeiitic andesite, and the presence of an interstitial glassy residuum in a relatively coarse-grained groundmass suggests that undercooling

has been instrumental in the dendritic magnetite rimming ilmenite laths in which swallow tail terminations are common.

The Fe-Ti oxide assemblage in the tholeiitic andesite gabbroic inclusions is texturally similar to that of the host rock. Globules of pyrrhotite occur throughout the groundmass and the orthopyroxene megacrysts and cumulates contain large embayed crystals of pyrrhotite up to 0.8 mm across.

3.5.5.2 Analytical Results

Analyses of titanomagnetite and ilmenite groundmass phases (Table 3.11) in the Comboyne tholeiitic andesite indicate, respectively that these have lower ulvospinel and higher haematite contents than similar oxides in the alkaline association. Al_2O_3 values are typical of Fe-Ti oxides precipitated in a low pressure environment, and there is only minor variation in composition between discrete and intergrown phases.

3.5.5.3 $f(O_2)$ and Temperature

The oxygen fugacity and temperature determined for the Comboyne tholeiitic andesite, using ulvospinel and haematite values for coexisting titanomagnetite and ilmenite, show limited variation from $10^{-10.8}$ to $10^{-11.5}$ atm. at temperatures from 1010° to 955° C. These values are intermediate to those determined by Anderson and Wright (1972) for the Kilauea tholeiitic basalts ($10^{-10.2}$ to $10^{-10.6}$ atm. and 1050 to 1180° C), and Carmichael (1967) for salic tholeiitic and calc-alkaline rocks ($10^{-13.9}$ to $10^{-9.1}$ atm. at 815° to 1010° C.). The $f(O_2)$ values are significantly lower than those derived by Fudali (1965) in equilibrium experiments on a range of rock types including tholeiitic basalts and andesites ($10^{-9.0}$ to $10^{-6.5}$ atm. at 1200° C.). The abundance

TABLE 3.11

MICROPROBE ANALYSES OF TITANOMAGNETITE AND ILMENITE IN

THOLEIITIC ANDESITE

TITANOMAGNETITE

Rock Type:	Tholeiitic Andesite		
Sample No:	6577	6577	6577
Phase:	G D	G D	G I
TiO ₂	19.41	19.49	21.72
Al ₂ O ₃	1.10	1.11	1.35
Cr ₂ O ₃	-	-	.15
FeO	72.45	72.78	71.77
MnO	.50	.50	.38
MgO	.37	.37	.72
CaO	.29	.29	.20
Total	94.12	94.54	96.29
Mg _{0.7} Fe _{0.3} Al ₂ O ₄	1.67	1.67	1.06
Mg ₂ TiO ₄	.48	.48	1.12
Mn ₂ TiO ₄	.89	.89	.67
CaFe ₂ O ₄	1.08	1.08	.86
FeCr ₂ O ₄	-	-	.22
Fe ₂ TiO ₄	52.77	52.99	58.58
Fe ₂ O ₃	40.52	40.75	35.89
Total	97.41	97.86	98.40
Usp.	56.57	56.53	62.01

ILMENITE

Rock Type:	Tholeiitic Andesite		
Sample No:	6577	6577	6577
Phase:	G D	G D	G I
TiO ₂	47.10	47.83	47.68
Al ₂ O ₃	.40	.33	.41
Cr ₂ O ₃	.23	.14	-
FeO	46.77	46.72	47.22
MnO	.38	.49	.40
MgO	1.30	1.08	1.24
CaO	.25	.46	.23
Total	96.43	97.05	97.18
Al ₂ O ₃	.40	.33	.41
Cr ₂ O ₃	.23	.14	-
CaTiO ₃	.68	1.09	.54
MgTiO ₃	3.85	3.25	3.73
MnTiO ₃	.60	1.06	.91
FeTiO ₃	83.16	84.53	84.37
Fe ₂ O ₃	8.30	7.51	8.15
Total	97.22	97.91	98.11
Hm	9.08	8.16	8.81
log ₁₀ f(O ₂)	-11.4	-11.5	-10.8
T° C.	960	955	1010

End members calculated by Anderson's (1968b) method

Total iron as FeO

G = Groundmass

D = Discrete phase

I = Intergrowth

of silicate phenocrysts (page 19) and the relatively coarse grain size of the groundmass suggests an extended crystallisation interval for this tholeiitic andesite, and the $f(O_2)$ and temperature determined from the groundmass Fe-Ti oxides probably represents equilibration values for late in the crystallisation interval. The occurrence of interstitial glasses, however, indicates that the residual liquid was quenched during the final stages of crystallisation.

Wass (1973) observes that both skeletal and dendritic forms indicate a lack of chemical equilibrium during crystallisation and, because of this, $f(O_2)$ and temperature cannot be determined accurately for the host rock using Buddington and Lindsley's (1964) curves for coexisting ilmenite and magnetite. The close agreement between analyses of both discrete and intergrown ilmenite and titanomagnetite in the Comboyne tholeiitic andesite suggest that in this instance equilibrium between coexisting Fe-Ti oxides and the melt was approached.

3.5.6 Calc-Alkaline Rocks

3.5.6.1 Textural Relationships

Anhedral ilmenite is the major Fe-Ti oxide phase present in the Lorne Basin calc-alkaline rocks. It is distinctly pink in colour and strongly anisotropic. Irregular patches and lamellae of bluish-white haematite are common within ilmenite crystals in the more altered rocks.

In a number of instances, including the Lorne Forest and Diamond Head diorites and the Middle Brother granodiorite, magnetite phenocrysts (up to 0.3 mm across) exhibit varying degrees of replacement by silicates, which include sphene. Initially replacement is confined to an apparent outlining of {111}

exsolution lamellae of ilmenite. With more complete replacement the former magnetite can only be identified by a lattice of (111) lamellae in a matrix of secondary minerals (Plate 15b). Equidimensional pyrite up to 0.5 mm across, totals approximately 2% in the Diamond Head dicrite, and approximately 5% in the Diamond Head dacites where pyrite is up to 1.0 cm across. Scattered globules of chalcopyrite occur in most of the Lorne Basin sub-alkaline rocks.

3.5.6.2 Analytical Results

Due to the generally oxidised nature of the Lorne Basin calc-alkaline rocks and the absence of discrete magnetite and ilmenite phases, analyses of the Fe-Ti oxides have not been attempted.

3.6 Amphibole

3.6.1 Introduction

Amphibole has only been identified in two rock types in the Comboyne alkaline series. Sparse prismatic amphibole occurs in the Comboyne benmoreites, and riebeckite/arfvedsonite is abundant in the associated comendites. In many instances, the fine grain size of the riebeckite/arfvedsonite has prevented its accurate analysis by electron microprobe.

Green-brown hornblende occurs in the groundmass and as rare microphenocrysts in the Lorne Basin calc-alkaline granitic rocks. Secondary fibrous amphibole is common in most of the calc-alkaline rocks. Trace quantities of (?)riebeckite/arfvedsonite occur in the strongly altered South Brother microgranite.

3.6.2 Alkaline Rocks

3.6.2.1 Textural Relationships.

Amphibole occurs in the Comboyne benmoreites as euhedral prismatic crystals within crystal aggregates consisting primarily of irregular corroded grains of anorthoclase, with minor olivine, titanomagnetite and striated apatite (Plate 13b). The amphibole ranges up to 0.3 mm long and exhibits strong pleochroism with X = pale yellow, Y = dark red-brown, and Z = dark brown.

Riebeckite/arfvedsonite solid solutions are commonly present as a groundmass phase only in the comendites, but fibrous and non-fibrous euhedral and subhedral phenocrysts up to 4 mm long occur in some plugs. These phenocrysts are commonly surrounded by a "mossy" rim of riebeckite/arfvedsonite similar in form and colour to that occurring in the groundmass. Riebeckite/arfvedsonite is a ubiquitous phase in all the comendites and is generally more abundant than aegirine-augite/aegirine when these two sodic ferromagnesian minerals coexist.

Pleochroism in the sodic amphiboles is variable. Phenocryst pleochroism is commonly X = deep prussian blue, Y = blue-green, Z = pale greenish-yellow. The pleochroic scheme for riebeckite/arfvedsonite forming rims on phenocrysts and occurring in the groundmass is X = prussian blue, Y = light indigo blue, Z = light greenish-brown. Ernst (1962) notes that as sodic amphiboles become richer in arfvedsonite, the colour changes from blue to green. This suggests that the amphibole phenocrysts in the Comboyne comendites have a higher arfvedsonite content than the coexisting groundmass amphibole. Pleochroism in the euhedral to subhedral prisms of riebeckite/arfvedsonite in the relatively coarse-grained Mt. Killabank comendite plug is X = deep prussian blue, Y = deep indigo blue, Z = light greenish-brown.

3.6.2.2 Analytical Results and Discussion

The analyses of amphibole listed in Table 3.12 have been calculated on a water-free basis with 23 oxygen atoms per formula unit.

Following the nomenclature of Leake (1968) the amphiboles in the Comboyne benmoreite are edenitic in composition, although in contrast to pure edenite which has no Al^{VI} component, the benmoreite amphibole contains .49 atomic proportions of Al^{VI} . The occurrence of this edenitic amphibole within crystal aggregates of anorthoclase, olivine, titanomagnetite, ilmenite and striated apatite indicates that these coexisting phases precipitated in a similar temperature-pressure regime.

Leake (1971), in a study of aluminous and edenitic hornblendes from metamorphic and igneous rocks, noted that there is no record of a naturally occurring amphibole with a composition close to edenite-ferrcedenite, and that metamorphic amphiboles approach this composition more closely than igneous amphiboles. He concluded that temperatures below that at which magmas exist appear essential for the stability of edenitic amphibole, and that there is no evidence to indicate that high pressures are required.

Although there is no other known recorded instance of edenitic amphibole occurring in volcanic rocks, Gilbert (1969) synthesized pargasite and edenite under wet conditions at 20 kb and 900° C. and concluded that the upper limits of thermal stability are lower at higher pressures. Green and Ringwood (1968) report pargasitic amphibole precipitating from a high-alumina quartz-tholeiite composition under hydrous conditions at approximately 1000° C and 10 kb, and Irving (1971) concluded that iron-rich kaersutite (considered to be coeval with titanbiotite, anorthoclase,

MICROPROBE ANALYSES OF AMPHIBOLES IN ALKALINE AND SUB-ALKALINE ROCKS

Rock Type:	Benmoreite		Comendite							Granodiorite
Sample:	6564	6564	6646	6652	6652	6647	6647	6569	6569	6700
Phase:	C	C	P	P	P	P	P	G	G	G
SiO ₂	45.78	44.66	45.20	48.94	49.75	48.09	50.55	49.89	50.19	47.19
TiO ₂	1.21	1.17	1.66	.78	.69	1.14	.95	.22	.27	1.17
Al ₂ O ₃	7.18	7.48	2.89	2.69	1.73	2.25	.81	.74	.64	5.93
FeO	15.67	18.24	31.24	34.94	34.34	32.14	31.49	33.04	32.21	20.86
MnO	.22	.29	.50	.47	.39	.39	.34	.89	1.00	.37
MgO	11.98	9.90	1.90	-	.13	.09	-	.11	.09	10.43
CaO	11.25	11.21	4.90	1.43	.39	1.52	.31	2.12	2.21	9.97
Na ₂ O	2.82	2.80	6.26	7.35	8.30	9.70	9.58	7.95	7.96	2.11
K ₂ O	1.59	1.71	1.30	1.38	1.53	1.46	1.67	1.58	1.50	.74
P ₂ O ₅	-	.15	-	-	-	-	-	.55	.97	-
Total	97.70	97.61	95.85	97.98	97.25	96.78	95.70	97.09	97.04	98.77
Cation Content (23 oxygen atoms)										
Si	7.1720	7.1028	7.7484	8.1855	8.3570	8.1339	8.5459	8.3938	8.3928	7.3962
Al ^{IV}	.8259	.8972	.3324	-	-	-	-	-	-	.6038
Al ^{VI}	.4982	.5058	.2516	.5294	.3429	.4485	.1616	.1448	.1264	.4916
Ti	.1427	.1395	.2144	.0984	.0866	.1451	.1204	.0272	.0348	.1380
Mg	2.7963	2.3474	.4849	-	.0331	.0240	-	.0267	.0224	2.4358
Fe	2.0530	2.4264	4.4790	4.8874	4.8241	4.5469	4.4525	4.6470	4.5057	2.7347
Mn	.0283	.0389	.0732	.0662	.0550	.0556	.0490	.1270	.1418	.0487
Ca	1.8878	1.9105	.9002	.2558	.0697	.2753	.0557	.3812	.3964	1.6748
Na	.8345	.8420	2.0274	2.3224	2.6326	3.0981	3.0589	2.5252	2.5142	.6411
K	.3173	.3472	.2850	.2944	.3280	.3161	.3596	.3377	.3198	.1476
P	-	.0197	-	-	-	-	-	.0787	.1367	-
Total	16.5560	16.5775	16.7965	16.6395	16.7290	17.0435	16.8036	16.6893	16.5910	16.3123
Z(Si,Al)	8.4961	8.5058	8.3324	8.7149	8.6999	8.5824	8.7075	8.5386	8.5192	8.4916
Y(Ti,Fe,Mn,Mg)	5.0203	4.9523	5.2515	5.0520	4.9988	4.7716	4.6219	4.8279	4.7047	5.3572
X(Ca,Na,K)	3.0396	3.1194	3.2126	2.8726	3.0303	3.6895	3.4742	3.2441	3.2304	2.4635
$\frac{100\text{Mg}}{\text{Mg} + \text{Fe}^{2+}}$	57.7	49.2	ND	ND	ND	ND	ND	ND	ND	47.1

C = Cumulate
 P = Phenocryst
 G = Groundmass
 ND = Not determined

and apatite megacrysts), from the Victorian Newer Volcanics, formed at pressures of less than 12 kb.

Kaersutite megacrysts have been identified in a number of alkaline basalt associations (Wilkinson, 1962; White, 1966; Mason, 1966, 1968a, 1968b; Aoki, 1967, 1970; Dickey, 1968; Le Maître, 1969; Bäst, 1970; Varne, 1970; Wilshire et al. 1971; Wass, 1971; Irving, 1974; Kesson and Price, 1972) and these have been interpreted as high pressure crystallisation products. From a synthesis of data (Lovering and White, 1969; Kushiro, 1970), it appears that pargasitic amphibole is stable in alkali basalts under wet conditions at temperatures between 960 - 1015° C. at pressures of 9 - 15 kb. In high-alumina quartz tholeiites, experimental results (Green and Ringwood, 1968) indicate pargasite crystallises over a pressure range of 9 to 10 kb at 920 to 1040° C. Nesbitt and Hamilton (1970) report the crystallisation of pargasite from an alkali olivine basalt at 2 kb.

The common occurrence of kaersutite megacrysts in under-saturated rocks, and its apparent scarcity in silica-saturated basalts, suggests that the host rock composition as well as crystallisation environment are the primary control on its formation. Similarly, the occurrence of edenitic amphibole in the silica-saturated Comboyne benmoreites instead of the more Al, Ti-rich and Si-poor kaersutitic amphibole is probably a direct response to melt composition and the pressure and temperature at the time of its crystallisation. These edenitic amphiboles contrast with the kaersutitic amphiboles produced in hydrous experimental runs on the Mt. Baldy hawaiiite at 5 and 10 kb. TiO_2 is significantly lower in the benmoreites (1.09 - 1.52%) than in the Mt. Baldy hawaiiite (2.36%), and the availability of Ti may be instrumental in determining the composition of the

amphibole precipitated. The reduced availability of TiO_2 for incorporation in the benmoreite amphibole would in turn restrict the amount of Al_2O_3 in the lattice.

The occurrence of amphibole in crystal aggregates with a similar mineral assemblage as cumulates considered to have crystallised at high pressures in other associations (Wass, 1971; Irving, 1974), supports the conclusion that these aggregates are cognate and crystallised at elevated pressures. Le Maitre (1969) concluded that kaersutite from Tristan da Cunha crystallised between 5 and 25 km (2-8 kb). Wass (1971) suggested that the association of kaersutite with pyroxene megacrysts with the highest CaO and TiO_2 values, and with olivine, titanomagnetite and ilmenite, indicates that they crystallised at lower pressures than sub-calcic pyroxenes, Ti-poor spinels and orthopyroxenes. Aoki (1970) also noted that the petrography and chemistry of inclusions in alkali basalts from Iki Island, Japan, and recent experimental work on natural rock systems, suggested that the Iki Island kaersutite-bearing inclusions formed under hydrous conditions at approximately 5 - 10 kb. Similarly, Irving (1971) concluded from experimental results for a nepheline mugearite that the presence of feldspar restricts the pressures of formation of the megacryst assemblage consisting of iron-rich kaersutite, titanbiotite, anorthoclase and apatite, to less than 12 kb. From the above data, and from experimental work undertaken in the present study (Chapter 5), it is suggested that the amphiboles in the Comboyne benmoreites crystallised at pressures of less than 6 kb.

The common occurrence of riebeckite/arfvedsonite and aegirine-augite/aegirine as interstitial groundmass phases and their less common occurrence as phenocrysts, indicates that these minerals, along with the more rare aenigmatite, were late-crystallisation

phases in the Comboyne comendites. Textural evidence relating to order of crystallisation of these sodic ferromagnesian minerals tends to be ambiguous. However, riebeckite/arfvedsonite phenocrysts are larger and more abundant than aegirine phenocrysts, and the latter tend to be partially rimmed by riebeckite/arfvedsonite, suggesting that crystallisation of amphibole continued after that of pyroxene. A late-stage vein in the Mt Killabuck comendite contains aegirine as the only ferromagnesian mineral. In this rock euhedral phenocrysts of aegirine are partially enclosed within riebeckite/arfvedsonite. Sodic ferromagnesian assemblages observed in the Comboyne comendites include:

1. Riebeckite/arfvedsonite - aegirine - aenigmatite
2. Riebeckite/arfvedsonite - aegirine
3. Riebeckite/arfvedsonite

Jacobsen et al. (1958) proposed that fluctuations in the chemical composition of the magma could result in variations in order of crystallisation of aegirine and riebeckite. Bailey (1969) suggested variation in oxygen fugacity as the controlling factor, and Nash and Wilkinson (1970) suggested that acmite and arfvedsonite may simply be in equilibrium over a finite, but perhaps limited, oxygen fugacity-temperature interval in natural rocks. Ernst (1962) has shown experimentally that riebeckite/arfvedsonite solid-solutions can exist stably only if fluid pressure is of the order of 1000 bars or more and oxygen fugacity is low. Magmatic riebeckite/arfvedsonite typically is confined to reduced silicic and/or alkali syenites and granites with high alkali and volatile content, low CaO, and enriched in FeO relative to MgO.

Figure 3.11 indicates that a decrease in oxygen fugacity will move a liquid from fields precipitating one or more of the minerals haematite, acmite, magnetite, fayalite and aenigmatite, to one

precipitating riebeckite/arfvedsonite. A single sodic ferromagnesian mineral, riebeckite/arfvedsonite, commonly occurs in the coarser-grained, more slowly cooled plug-forming comendites suggesting equilibrium crystallisation has been approached. Late-stage aegirine in veins traversing the Mt. Killabuck comendite indicate that an increase in volatiles probably resulted in an increase in oxygen fugacity causing the liquid to enter the aegirine stability field (Fig. 3.11).

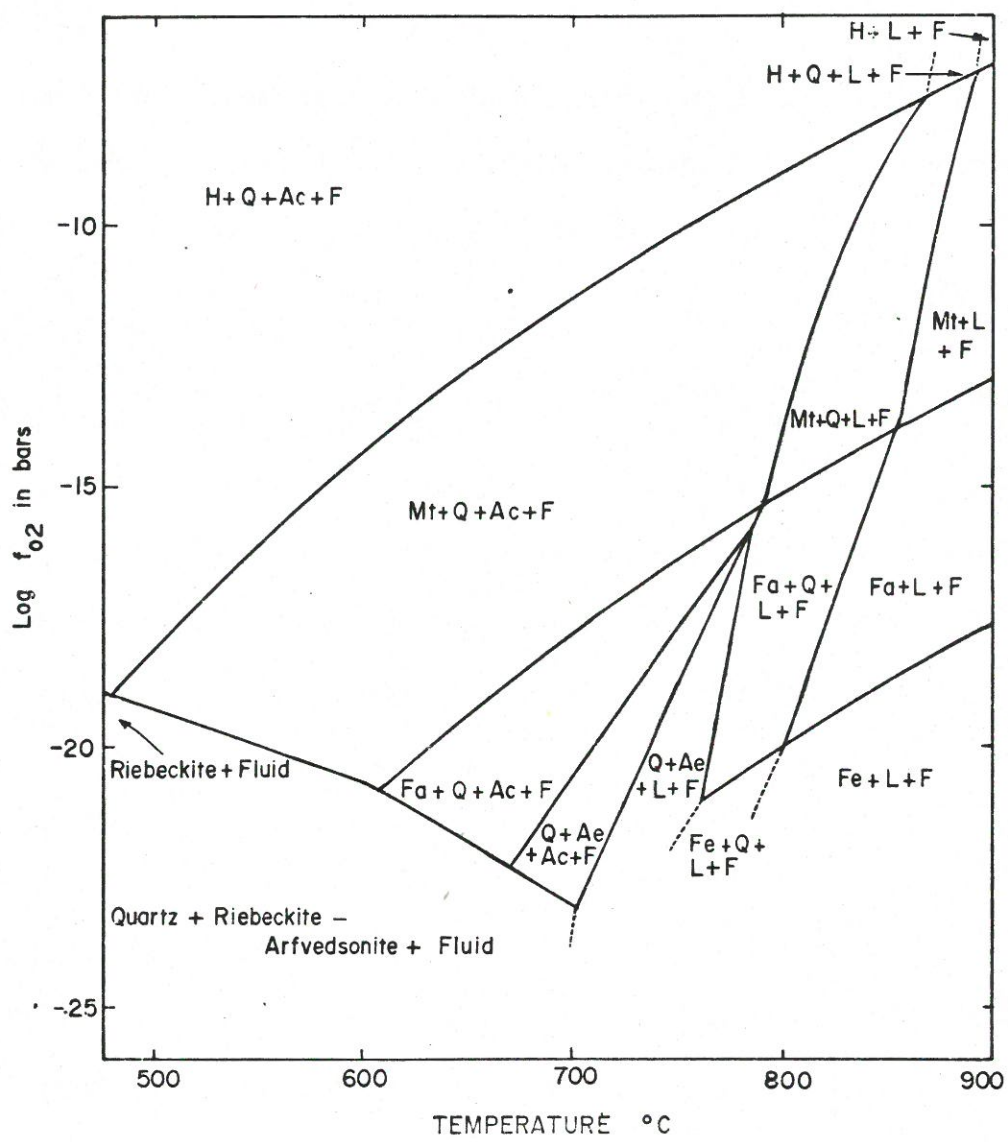
It is concluded that the different sodic ferromagnesian assemblages in the Comboyne comendites can be attributed primarily to fluctuations in oxygen fugacity during crystallisation and variations in rate of crystallisation (flow or plug).

As has also been noted by Nash et al. (1969) in the Mt. Suswa, Kenya, association, the Comboyne riebeckite/arfvedsonite solid solutions are strongly enriched in MgO and CaO relative to the host rock. Highest MgO and CaO values occur in phenocrysts from plugs and in groundmass phases in flows which have slightly lower SiO_2 and higher MgO and CaO values than the comendites forming the associated plugs.

Riebeckite/arfvedsonite coexisting with aenigmatite in the Comboyne comendites tends to show depletion in TiO_2 and Al_2O_3 and enrichment in MnO, CaO and Na_2O relative to that occurring in plugs and not coexisting with aenigmatite. Riebeckite/arfvedsonite phenocrysts in the Mt. Coxcomb comendite (which contains groundmass aenigmatite) have comparatively high Al_2O_3 , TiO_2 (Table 3.12) which supports textural evidence indicating they crystallised before aenigmatite.

Fig. 3.11 Isobaric $f(\text{O}_2)$ - T diagram for the bulk composition $\text{Na}_2\text{O} \cdot 0.5\text{FeO}_x \cdot 0.8\text{SiO}_2$ + excess water at 500 bars Pfluid (after Ernst, 1962), showing stability fields for the sodic-ferromagnesian minerals riebeckite/arfvedsonite, acmite and aenigmatite.

Fa = fayalite; Ae = aenigmatite;
Mt = magnetite; H = haematite; Q = quartz; Ac = acmite; Fe = iron; F = fluid; L = liquid.



3.6.3 Calc-Alkaline Rocks

3.6.3.1 Textural Relationships

Amphibole in the Lorne Basin calc-alkaline rocks is mostly secondary and with chlorite, biotite and Fe-Ti oxides, commonly replaces pyroxene. However, microphenocryst and groundmass green-brown hornblende occurs in some of the granitic rocks and this hornblende appears to be primary in origin.

Green fibrous amphibole replaces pyroxene in porphyritic microgranites and granodiorites, and pyroxenes in the North Brother porphyritic microgranite are commonly rimmed by clear green amphibole which exhibits only minor pleochroism. This amphibole is, in turn, partially replaced by brown biotite.

Two texturally distinct amphiboles occur in the Middle Brother granodiorite. One is fibrous and irregular in outline and ranges up to 0.5 mm long. The second type is clear and prismatic, and has a maximum length of approximately 0.35 mm. Both these amphiboles are pleochroic with X = straw yellow, Y = brown, Z = greenish-brown. The fibrous type appears to replace pyroxene and rare cores of pyroxene remain. However, the clear nature and euhedral outline of the prismatic amphibole suggest it has crystallised directly from the melt.

Rare "mossy" crystals of riebeckite/arfvedsonite have also been tentatively identified in the South Brother microgranite. The fine grain-size and the strong alteration of this rock restricts the positive identification of the majority of mineral types present.

3.6.3.2 Analytical Results and Discussion

The only amphibole analysed in the Lorne Basin sub-alkaline rocks occurs as brown prismatic crystals in the Middle Brother granodiorite. An analysis of this amphibole (Table 3.12) indicates

that it is sub-aluminous with the composition of a ferro-actinolitic hornblende (Leake, 1968). Similar Al-poor amphiboles in the Ben Nevis complex are considered by Haslam (1968) to have formed as the result of the breakdown of pyroxene at near solidus temperatures and unusually low pressures. Like the Ben Nevis complex, the Lorne Basin plutonic rocks are associated with volcanics and the fine grain size and outcrop pattern of the plutonic rocks indicate crystallisation occurred in a near-surface environment. Eggler (1972) notes that amphibole stability is a function of pressure, temperature and the fugacity of H_2O in the melt, and that, additionally, these intensive parameters must also be related to reactions involving amphibole. He showed experimentally that decrease in water vapour pressure limits the stability field of amphibole and that pyroxene will be the stable phase in anhydrous melts.

The high level of emplacement of the Lorne Basin calc-alkaline rocks proves the anhydrous character of the parental melts, and the initial crystallisation of pyroxene in these rocks is consistent with the above data. The subsequent replacement of pyroxene by fibrous amphibole indicates that a late-stage increase in volatile activity increased the water vapour pressure to a point where amphibole was stable. The occurrence of clear, euhedral grains of amphibole suggests that concomitant with the replacement of pyroxene, minor late-stage crystallisation of primary magmatic amphibole occurred.

Although riebeckite/arfvedsonite is generally not found in calc-alkaline rocks, experimental results (Ernst, 1962) show that hydroxyl-bearing pure riebeckite is stable at low temperatures and high oxidation states in authigenic low-grade metamorphic and pegmatitic environments. Ernst (1962) refers to riebeckite in

fissure-fillings and in pegmatite associated with the Quincy granite as being "obviously magmatic" and notes that Palache and Warren (1911) noted optically similar material from the Quincy granite proper which Ernst considers to be late-stage overgrowths and replacements of the blue-green or brown hornblende. Ernst also notes that near-pure riebeckite has been analysed from felsite dykes in Shetland and suggests that this fine-grained amphibole may have crystallised at subsolidus temperatures.

It is suggested that the (?)riebeckite in the South Brother microgranite probably crystallised as the result of auto-metamorphic processes at subsolidus temperatures during late-stage hydrothermal activity.

3.7 Aenigmatite

3.7.1 Textural Relationships

Aenigmatite is not common in the Comboyne comendites, and has only been positively identified in flows and a single plug (Mt. Coxcomb). In all instances it coexists with riebeckite/arfvedsonite and aegirine-augite/aegirine and exhibits the same "mossy" texture as the associated groundmass pyroxene and amphibole. Where these three sodic ferromagnesian minerals coexist they are present in approximately equal proportions. Aenigmatite pleochroism is x = light yellow-brown, y = red-brown, z = deep brown-black, and birefringence appears fairly high but is masked by the strong colour. The "mossy plates" range up to 0.3 mm across and show uniform extinction on rotation indicating the same crystal orientation throughout. Rare subrounded grains of magnetite are rimmed by aenigmatite.

3.7.2 Analytical Results and Discussion

The Comboyne aenigmatites are similar in composition to groundmass aenigmatites analysed by Abbott (1967a) from the Nandewar Mountain trachyte, although Al_2O_3 and K_2O tends to be higher and MnO lower. Nicholls and Carmichael (1969) note that aenigmatites generally show little variation in composition, although those from crystalline comendites have slightly less TiO_2 than those found in obsidians. In the present study this tendency is not apparent when aenigmatites from plugs and flows are compared. However, it is noted that in both instances aenigmatite occurs as a groundmass phase.

The extremely fine grain-size of aenigmatite in the Comboyne comendites has caused difficulty in successfully analysing this mineral by microprobe methods. Kelsey and McKie (1964), after determining the unit cell of aenigmatite on a basis of 40 anions, concluded that Si is confined to 4-fold coordination, Al^{3+} and Fe^{3+} can occupy either 4-fold or 6-fold sites, and that there is no more than twelve 4-fold sites in the unit cell. The aenigmatite from Mt. Coxcomb has 12.1717 atoms of Si in the structural formula indicating that the fine grain-size may have resulted in this analysis not being completely accurate (Table 3.13).

Aenigmatite has only been identified in peralkaline rocks in which there is an excess of $\text{Na}_2\text{O} + \text{K}_2\text{O}$ over Al_2O_3 , and conditions of low oxygen fugacity are considered necessary for its formation (Abbott, 1967a). Ernst (1962) synthesized Ti-free aenigmatite with $p\text{H}_2\text{O}$ less than 1000 bars and $f(\text{O}_2)$ defined by the magnetite-wustite and wustite-iron buffers, and both Kelsey and McKie (1964) and Abbott (1967a) noted that aenigmatite is stable only at very low $f(\text{O}_2)$. Abbott, however, suggested that Ti-bearing natural aenigmatite would be stable over a wider range of relatively low

TABLE 3.13

MICROPROBE ANALYSES OF AENIGMATITE AND APATITEIN ALKALINE ROCKS

Rock Type:	<u>Comendite</u>		<u>Hawaiite</u>
Sample No:	6569	6646	6498
Phase:	Aenigmatite G	Aenigmatite G	Apatite M
SiO ₂	40.96	42.35	-
TiO ₂	7.13	7.57	-
Al ₂ O ₃	.90	1.84	-
FeO	41.10	38.03	-
MnO	.76	.68	-
MgO	-	.15	.16
CaO	.43	.21	52.48
Na ₂ O	7.26	7.02	-
K ₂ O	.17	.34	-
P ₂ O ₅	-	-	39.40
Total	98.71	98.19	92.04

Cation Content (40 oxygen atoms)

Si	11.9721	12.1717
Al	.3091	.6217
Ti	1.5668	1.6355
Mg	-	.0639
Fe	10.0470	9.1409
Mn	.1879	.1658
Ca	.1352	.0639
Na	4.1136	3.9134
K	.0632	.1264
P		
Total	28.3949	27.9032

100Ti 13.49 15.08

Ti + Fe²⁺

Total iron as FeO

oxygen fugacities and Nicholls and Carmichael (1969) estimated the free energy of formation for aenigmatite and from this calculated that it may be stable at oxygen fugacities well above the NNO buffer. Lindsley (1971) showed experimentally that the stability curve of aenigmatite lies between the NNO and FMQ buffers at 750° C and 500 bars. Lindsley and Haggerty (1971) report the replacement of ilmenomagnetite by aenigmatite with the selective replacement of (111) ilmenite lamellae indicating that it occurred after oxidation resulted in the formation of these lamellae. This is consistent with experimental results (Lindsley, 1971) which indicate the maximum thermal stability of aenigmatite is less than 900° C.

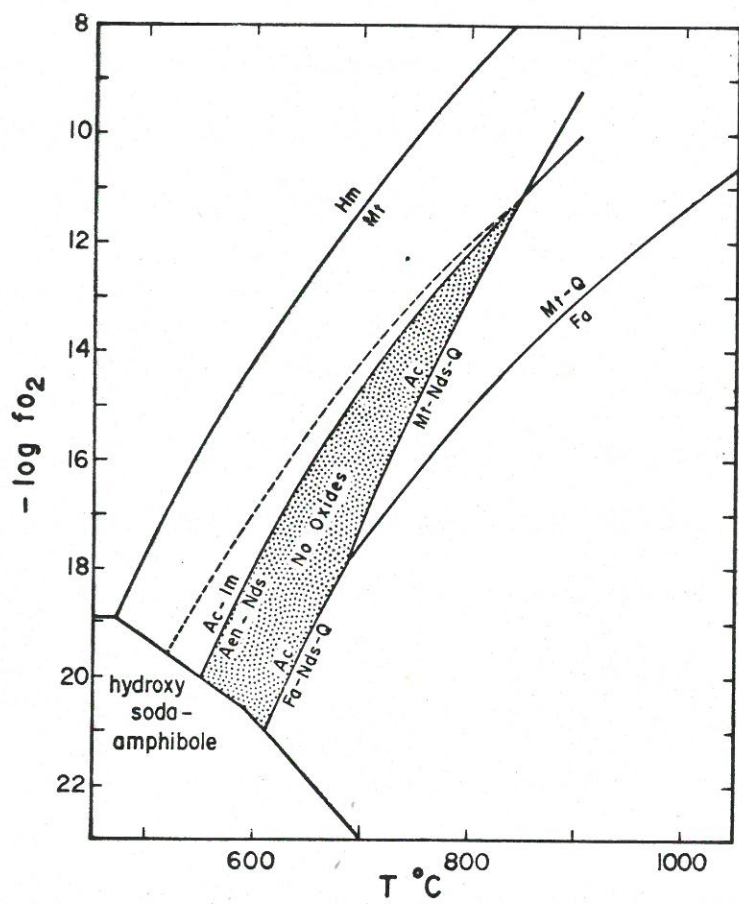
Abbott (1967a) describes titanomagnetites rimmed by aenigmatite and the same relationship has been observed in a Comboyne comendite flow. Both these occurrences support Nicholls and Carmichael's (1969) suggestion that there is a field in $f(\text{O}_2) - T$ space where Fe-Ti oxides are absent and where acmite and aenigmatite coexist, and as crystallisation proceeds the residual liquids go from this field into that where arfvedsonite is stable (Figure 3.12 - also Ernst, 1962).

This crystallisation path is comparable with those observed in the Comboyne peralkaline rocks. Fe-Ti oxides occur only as extremely rare microphenocrysts and are rimmed by aenigmatite. In most instances, riebeckite/arfvedsonite is the major sodic ferromagnesian phase and it is the only mafic mineral in a number of comendite plugs where its pleochroism suggests it has a higher riebeckite component than those sodic amphiboles coexisting with aegirine (Ernst, 1962).

It is concluded that the Comboyne comendites started to crystallise at temperatures in excess of 800° C. and $f(\text{O}_2)$ of

Fig. 3.12 No-oxide field delineated by Nicholls and Carmichael (1969). Nicholls and Carmichael note that this field will shift in $f(O_2)$, T space and be suppressed or extended as the result of variations in the degree of solid solution in the phases involved.

Hm = haematite; Mt = magnetite; Ac = acmite;
Il = ilmenite; Aen = aenigmatite; Fa = fayalite; Q = quartz; Nds = sodium disilicate.



approximately 10^{-12} atm. With a decrease in temperature and $f(O_2)$, the melt entered the oxide-absent field delineated by Nicholls and Carmichael (1969) (Fig. 3.12) resulting in the crystallisation of aenigmatite, aegirine and riebeckite/arfvedsonite.

3.8 Apatite

3.8.1 Textural Relationships

Euhedral, but slightly rounded, prismatic apatite crystals up to 2.5 mm long are common in the Comboyne hawaiites and benmoreites, and occur less commonly in the mugearites and mafic trachytes. They are present as discrete crystals, and as inclusions in titanomagnetite, olivine and plagioclase crystals, and as crystal aggregates with clinopyroxene and titanomagnetite in the hawaiites, and with anorthoclase, olivine, titanomagnetite and edenitic amphibole in benmoreites. The central area of these large apatite crystals characteristically has acicular inclusions which show high reflectivity and which intersect at an angle of 60° (Plates 6 & 16a). Clear acicular apatite is a common ground-mass phase in both alkaline and sub-alkaline rocks in the Comboyne and Lorne Basin area.

3.8.2 Analytical Results and Discussion

A partial analysis of a striated apatite megacryst associated with strongly sieved clinopyroxene in the Mt. Baldy hawaiite (Fig. 6a) shows it has 52.48% CaO, 39.40% P_2O_5 and 0.16% MgO (Table 3.13). An attempt was also made to analyse the acicular inclusions in these apatites and although the fine grain size prevented accurate analysis, the results indicate that these inclusions are probably spinel with a FeO:MgO ratio of slightly less than 2:1, and with minor SiO_2 , Al_2O_3 , MnO and CaO (Plate 16a).

Apatite (?)phenocrysts/megacrysts described from other alkaline associations are listed in Table 3.14. This table also lists the approximate value of P_2O_5 in the host rock. In most instances this value is less than 1.5%, and is not high compared with similar rocks in which large apatite crystals are not observed. The common occurrence of apatite in close association with other phases considered to have crystallised at elevated pressures (olivine, clinopyroxene, amphibole, anorthoclase, plagioclase, titanomagnetite) suggests that the apatite formed in a similar pressure-temperature regime. Textural evidence also supports this conclusion. These apatites exhibit rounding of crystal faces indicating that corrosion has occurred subsequent to their initial crystallisation. Additionally, the presence of crystallographically orientated rods of (?)spinel suggests that this phase was exsolved as the result of a changing physico-chemical environment. The identification of apatite in the hawaiite experimental runs undertaken in this study (Chapter 5) also supports the suggestion that the large striated apatites in the Comboyne alkaline rocks crystallised at elevated pressures.

TABLE 3.14

APATITE (?) PHENOCRYSTS/MEGACRYSTS DESCRIBED FROM

SELECTED BASALTIC ASSOCIATIONS

Location	Rock Type	Approx. P_{205} in Host Rock	Reference
Comboyne	Hawaiite	.39 - 1.16%	This study
	Mugearite	.31 - 1.31%	" "
	Benmoreite	.47 - .88%	" "
	Mafic Trachyte	.31%	" "
Nandewar	Hawaiite	1.30%	Pers.obs. Abbott (1969)
East Anakies, Vic.	Nepheline Mugearite	.68%	Irving (1971)
Dogo Island	Felsic Rocks	1.00%	Uchimizu (1966)
Gough Island	Trachyte	.74%	Le Maitre (1962)
Iki Island	Trachyandesite	.75%	Aoki (1970)
Kenya Rift	Phonolite	.07%	Lippard (1973)
Marquesas Is.	Hawaiite	1.41%	Bishop & Woolley (1973)
Jebel Khariz	Trachyte	.25%	Gass & Mallick (1968)
Aden and Little Aden	Trachyandesite Trachyte	.65%	Cox et al., (1969)
Morotu	Dolerite, Monzonite, Syenite	.52%	Yagi (1953)
Southern Highlands, N.S.W.	Hawaiite	.72%	Wass (1971)
Massif Central	Alkali basalt	-	Wass (pers.comm.)
Hawaii (Kohala)	Mugearite	1.97%	Pers.obs. Macdonald and Katsura (1968)
Spring Mt., N.S.W.	Olivine analcinite	1.17%	Irving (1974) Wilkinson (1962)
Garrawilla, N.S.W.	Alkali basalt	-	Wass (pers.comm.)
Keepit Dam, N.S.W.	Alkali basalt	-	Wass (pers.comm.)
Bombo Quarry, Kiama, N.S.W.	Alkali basalt	-	Wass (pers.comm.)
Delegate, N.S.W.	Olivine nephelinite	1.04%	Irving (1974) Lovering & White (1969)
Kakanui, N.Z.	Melanephelinite	-	Dickey (1968)
Tristan da Cunha	Trachybasalt	.52%	Baker et al. (1964)
Siberia and Dish Hill, California	Basanite	1.30%	Wilshire & Trask (1971)
Comores Archipelago	Hawaiite Mugearite	.89% .88%	Flower (1973) " "
Reunion Is.	Benmoreite	.50%	Upton & Wadsworth (1972)
St. Helena	Trachybasalt	.80%	Baker (1969)
	Trachyandesite	.70%	" "

TABLE 4.1

MAJOR AND TRACE ELEMENT ANALYSES OF ALKALINE ROCKS

Rock Type:	Hawaiite										Mugearite				
Sample No:	6571	6570	6572	6596	6565	6500	6499	6597	6562	6498	6607	6625	6608	6567	6578
SiO ₂	51.65	51.61	51.43	54.36	53.07	54.51	52.86	55.01	52.55	51.13	52.18	51.66	52.82	56.52	59.54
TiO ₂	2.30	2.73	2.69	2.21	2.18	2.17	2.25	1.99	2.23	2.36	2.29	2.29	2.07	1.67	1.37
Al ₂ O ₃	16.84	16.23	16.35	16.42	16.62	16.34	15.78	16.15	15.98	16.01	15.83	15.93	16.08	15.41	16.53
Fe ₂ O ₃	2.44	2.66	2.59	2.25	2.44	2.26	2.57	2.18	2.56	2.63	2.93	2.91	2.79	2.38	1.67
FeO	8.14	8.90	8.69	7.48	8.11	7.54	8.53	7.26	8.51	8.76	9.75	9.70	9.30	7.92	5.58
MnO	.17	.18	.16	.13	.17	.17	.18	.13	.18	.16	.19	.18	.20	.17	.17
MgO	4.55	3.89	3.82	4.25	3.88	3.95	3.90	4.15	3.81	4.17	2.43	2.73	2.35	2.19	2.16
CaO	7.82	7.27	7.45	6.50	6.68	6.18	6.72	6.01	6.59	6.83	5.57	5.76	5.21	4.57	4.65
Na ₂ O	4.04	4.15	4.40	4.47	4.34	4.54	4.42	4.45	4.50	4.63	4.73	4.76	4.88	4.95	4.81
K ₂ O	1.41	1.78	1.83	1.46	1.90	1.88	2.07	2.29	2.22	2.16	2.80	2.77	3.07	3.47	3.21
P ₂ O ₅	.65	.60	.59	.47	.61	.46	.69	.39	.87	1.15	1.30	1.31	1.23	.75	.31
100 Mg Mg + Fe	49.89	43.78	43.92	50.31	46.01	48.30	44.83	50.47	44.39	45.89	30.77	33.40	31.06	33.01	40.82
Q	-	-	-	1.91	-	1.11	-	.73	-	-	-	-	-	.30	5.36
Or	8.33	10.52	10.80	8.61	11.23	11.11	12.25	13.53	13.12	12.76	16.53	16.36	18.12	20.51	18.98
Ab	34.15	35.07	37.24	37.80	36.70	38.39	37.35	37.63	38.06	39.16	40.02	40.20	41.27	41.83	40.66
An	23.67	20.42	19.46	20.45	20.26	18.65	17.12	17.33	16.85	16.53	13.67	13.97	12.93	9.60	14.04
Ol	8.93	9.72	11.32	7.14	7.35	7.40	9.68	8.16	8.42	8.08	4.65	5.09	4.13	6.89	5.87
Hy	13.18	11.07	3.63	15.56	14.66	14.88	10.90	14.80	8.32	.74	5.85	1.74	3.64	13.00	9.36
Ol	2.34	2.78	7.32	-	.72	-	3.08	-	5.27	11.74	8.08	11.34	9.67	-	-
Wt	3.54	3.85	3.75	3.25	3.54	3.28	3.74	3.16	3.71	3.81	3.60	3.57	3.42	2.92	2.41
Il	4.37	5.19	5.11	4.19	4.13	4.12	4.28	3.78	4.24	4.48	4.35	4.35	3.94	3.16	2.60
Ap	1.51	1.39	1.36	1.09	1.41	1.07	1.60	.90	2.02	2.69	3.01	3.03	2.84	1.75	.72
100 An An + Ab	40.94	38.81	34.32	35.11	35.56	32.71	31.44	31.54	30.69	29.68	25.45	25.79	23.87	18.67	25.67
F.I.	42.48	45.59	48.04	48.32	47.93	50.61	49.70	51.89	51.18	51.91	56.57	56.59	59.42	62.67	65.00
Ba	309	369	369	315	427	336	581	382	410	560	628	642	776	526	304
Rb	24	24	28	26	32	36	41	43	35	30	42	40	39	56	84
Sr	615	577	582	436	499	431	536	435	527	633	520	536	493	313	366
Y	24	26	26	29	29	32	33	28	32	29	45	48	49	38	40
U	2	<1	2	<1	<1	<1	2	2	<1	<1	<1	4	2	3	3
Th	5	5	5	2	8	4	10	9	10	7	9	8	8	11	15
Zr	204	284	282	207	270	259	313	334	335	242	465	460	545	376	454
V	90	100	100	ND	ND	76	ND	63	ND	71	51	ND	ND	73	46
Cr	71	42	43	ND	ND	71	ND	86	ND	40	2	ND	ND	37	40
Ni	39	23	25	ND	ND	29	ND	38	ND	22	.1	ND	ND	19	32
Cu	37	37	39	ND	ND	33	ND	36	ND	38	39	ND	ND	33	28
Zn	71	81	80	ND	ND	86	ND	81	ND	80	107	ND	ND	90	91
Ga	24	26	27	26	27	28	27	23	28	26	28	32	30	31	33
Pb	5	3	6	6	4	5	9	7	4	3	5	3	6	6	8
K/Rb	488	609	528	466	485	434	419	.442	527	592	542	563	645	513	310
Rb/Sr	.04	.04	.05	.06	.06	.08	.08	.10	.07	.05	.08	.07	.08	.18	.23
Th/U	2.50	-	2.50	-	-	-	5.00	4.50	-	-	-	2.00	4.00	2.00	5.00
Sr/Ca	.011	.011	.011	.009	.011	.010	.011	.010	.011	.013	.013	.013	.013	.010	.011
Ba/Sr	.50	.64	.64	.72	.86	.78	1.08	.88	.78	.88	1.21	1.20	1.57	1.68	.83

TABLE 4.1 Cont'd

Rock Type:	Dacite								Trachyte		Anorthoclase Trachyte			
Sample No:	6575	6566	6564	6563	6505	6576	6507	6501	6667	6627	6631	6630	6635	6520
SiO ₂	57.04	56.31	57.22	57.14	57.60	57.05	58.24	59.24	62.60	63.24	64.19	65.63	65.32	67.00
TiO ₂	1.52	1.42	1.30	1.27	1.23	1.52	1.09	.62	1.06	.49	.46	.37	.30	.32
Al ₂ O ₃	15.53	15.35	15.68	15.59	15.95	15.53	16.72	15.43	15.67	17.45	17.60	15.98	16.43	17.19
Fe ₂ O ₃	3.52	3.70	3.51	3.65	3.42	3.52	2.85	3.32	2.52	1.61	1.31	1.71	1.48	.74
FeO	7.04	7.57	7.05	7.30	6.83	7.04	5.67	6.65	5.04	3.22	2.62	3.42	2.93	1.49
MnO	.16	.14	.24	.18	.17	.16	.17	.26	.35	.11	.07	.16	.10	.03
MgO	1.69	1.44	1.14	1.12	1.09	1.69	1.51	.60	1.02	.47	.65	.16	.14	.10
CaO	3.95	3.77	3.54	3.29	3.13	3.96	3.25	2.97	2.89	1.93	1.39	1.17	1.32	.69
Na ₂ O	5.03	4.79	4.85	4.96	5.15	5.03	5.31	5.42	6.33	6.18	6.23	5.69	6.22	6.46
K ₂ O	3.70	4.54	4.69	4.72	4.72	3.70	4.74	5.20	2.06	5.16	5.25	5.56	5.70	5.93
P ₂ O ₅	.81	.88	.78	.78	.71	.81	.47	.31	.46	.15	.15	.10	.06	.04
100Mg Mg + Fe ²⁺	29.98	25.30	22.39	21.48	22.31	29.95	32.22	13.86	26.52	20.71	30.61	7.75	7.90	10.78
Q	2.57	.71	1.34	.86	.56	3.29	-	-	9.41	2.24	3.20	7.54	4.07	5.08
Or	21.65	26.80	27.71	27.91	27.90	24.02	28.01	30.69	12.17	30.46	31.04	32.85	33.67	35.03
Ab	42.52	40.50	41.05	41.95	43.57	45.17	44.90	45.81	53.54	52.25	52.67	48.14	52.59	54.63
An	8.89	7.00	7.15	6.23	6.45	6.96	7.80	2.44	8.27	4.67	4.78	1.63	.10	.42
Ol	4.64	5.06	4.58	4.37	3.82	5.31	4.42	8.92	2.60	3.35	.94	3.10	5.30	2.38
Hy	9.68	9.69	8.81	9.28	8.77	7.09	6.13	4.32	7.28	3.43	4.25	3.33	1.42	.69
Ol	-	-	-	-	-	-	1.48	1.11	-	-	-	-	-	-
Kt	5.10	5.49	5.09	5.29	4.97	4.44	4.10	4.81	3.65	2.33	1.89	2.47	2.14	1.07
Il	2.52	2.70	2.47	2.41	2.33	2.39	2.07	1.17	2.01	.93	.87	.71	.57	.61
Ap	1.85	2.05	1.80	1.81	1.64	1.33	1.09	.73	1.07	.35	.35	.23	.14	.09
100 An An + Ab	17.28	14.74	14.83	13.12	12.89	13.36	14.81	5.05	13.38	8.21	8.32	3.27	.20	.76
F.I.	66.94	68.01	70.10	70.72	72.03	72.48	72.91	76.50	75.12	84.95	86.91	88.53	90.33	94.74
Ba	559	1019	1096	1038	1082	544	766	189	598	455	509	417	97	97
Rb	57	54	54	54	52	67	72	81	27	92	117	116	130	135
Sr	297	179	187	178	196	253	200	21	154	175	198	54	14	18
Y	40	43	39	38	36	40	34	47	68	35	25	75	58	38
U	4	<1	1	<1	3	3	<1	4	1	2	5	3	3	4
Th	9	8	10	6	8	13	9	10	9	13	14	15	12	18
Zr	401	556	611	614	613	456	653	613	618	657	809	870	1052	1191
V	11	ND	4	ND	ND	ND	ND	4	<1	<1	<1	ND	ND	ND
Cr	<1	ND	<1	ND	ND	ND	ND	<1	<1	<1	<1	ND	ND	ND
Ni	<1	ND	<1	ND	ND	ND	ND	<1	<1	6	8	ND	ND	ND
Cu	24	ND	21	ND	ND	ND	ND	24	16	22	22	ND	ND	ND
Zn	107	ND	91	ND	ND	ND	ND	132	87	83	87	ND	ND	ND
Ga	32	33	34	34	32	33	32	39	27	35	32	41	36	39
Pb	4	6	5	6	6	6	9	7	7	8	9	10	10	9
K/Rb	535	687	704	713	738	494	537	528	618	464	371	394	363	366
Rb/Sr	.19	.30	.29	.30	.27	.26	.36	3.85	.17	.53	.59	2.15	9.29	7.50
Th/U	2.25	-	-	-	2.67	4.33	-	2.50	5.00	6.50	2.80	5.00	4.00	4.50
Sr/Ca	.011	.007	.008	.008	.009	.011	.009	.001	.008	.013	.020	.007	.002	.004
Ba/Sr	1.88	5.69	5.86	5.83	5.52	2.15	3.83	9.00	3.88	2.60	2.57	7.72	6.93	5.39

TABLE 4.1 Cont'd

Rock Type: <u>stone</u>	Comendite											
Sample No:	6636	6639	6637	6679	6638	6669	6506	6573	6643	6644	6645	6646
SiO_2	73.32	70.50	74.07	73.35	70.76	69.23	71.61	72.10	72.79	71.95	73.22	73.81
TiO_2	.19	.20	.11	.13	.17	.26	.16	.15	.15	.18	.16	.17
Al_2O_3	13.44	14.13	13.95	11.39	14.84	14.26	12.14	12.28	13.55	11.16	12.45	12.14
Fe_2O_3	.78	1.64	2.21	4.62	1.62	2.28	4.05	3.30	3.93	3.75	1.89	1.52
FeO	2.10	1.87	.94	1.28	1.91	2.11	1.56	1.64	.03	2.25	1.70	1.96
MnO	.08	.07	.05	.07	.05	.08	.06	.07	.14	.09	.06	.08
MgO	.03	.10	.14	.13	.09	.07	.04	.07	.30	.07	.15	.13
CaO	.50	.41	.08	.14	.16	.47	.26	.17	.08	.34	.39	.42
Na_2O	5.13	6.19	4.34	4.56	5.59	6.13	5.80	5.79	4.36	5.71	5.52	5.37
K_2O	4.41	4.87	4.09	4.31	4.79	5.00	4.31	4.43	4.67	4.50	4.41	4.35
P_2O_5	.02	.02	.02	.02	.02	.12	.02	.02	.02	.01	.05	.05
$\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{2+}}$	1.76	5.09	7.90	4.06	4.48	2.85	1.36	2.59	12.96	2.13	7.25	6.46
Q	24.59	15.89	32.84	30.95	18.60	14.02	22.60	22.35	28.97	24.16	24.43	26.19
Or	26.06	28.76	24.17	25.44	26.30	29.56	25.44	26.15	27.61	26.59	26.10	25.71
Ab	43.35	45.55	36.69	34.63	47.31	45.50	38.50	38.52	36.89	32.37	39.57	38.24
An	.65	-	.28	-	.66	-	-	-	.28	-	-	-
C	-	-	2.29	-	.21	-	-	-	1.22	-	-	-
Ac	-	4.75	-	3.46	-	5.49	9.31	9.22	-	10.83	5.47	4.41
Na	-	.33	-	-	-	-	-	-	-	.84	-	.50
Ol	1.47	1.68	-	.45	-	1.36	1.03	.64	-	1.45	1.41	1.54
Hy	2.34	2.61	.36	.12	2.20	2.74	1.58	2.64	.75	3.40	2.60	2.97
Wt	1.13	-	2.87	3.97	2.35	.56	1.19	.15	.11	-	-	-
Ms	-	-	.24	.68	-	-	-	-	3.84	-	-	-
Il	.36	.38	.22	.25	.32	.49	.30	.28	.28	.34	.30	.32
Ap	.05	.05	.05	.05	.05	.28	.05	.05	.05	.02	.12	.12
$\frac{100 \text{ An}}{\text{An} + \text{Ab}}$	1.48	-	.74	-	1.39	-	-	-	.74	-	-	-
F.I.	94.00	95.28	93.70	94.48	94.21	94.57	95.84	96.23	93.47	94.79	95.57	95.05
Ba	36	2	7	3	6	7	6	5	20	9	83	73
Rb	190	221	337	408	260	197	261	289	904	304	292	239
Sr	8	6	6	6	7	7	7	6	5	9	23	25
Y	86	112	103	85	69	75	152	116	124	151	93	133
U	7	6	7	12	6	6	7	5	22	12	9	10
Th	24	.31	38	53	30	22	37	36	108	42	32	27
Zr	815	1157	1522	3135	1300	1283	2350	2394	5827	2580	1303	1300
V	ND	ND	ND	ND	ND	<1	<1	ND	ND	ND	ND	<1
Cr	ND	ND	ND	ND	ND	<1	<1	ND	ND	ND	ND	<1
Ni	ND	ND	ND	ND	ND	8	16	ND	ND	ND	ND	5
Cu	ND	ND	ND	ND	ND	21	25	ND	ND	ND	ND	9
Zn	ND	ND	ND	ND	ND	276	608	ND	ND	ND	ND	327
Ga	38	50	59	77	56	59	69	76	94	75	60	58
Pb	16	19	31	27	8	13	22	1	115	28	22	21
K/Rb	183	183	98	86	152	212	127	126	43	122	126	125
Rb/Sr	23.75	36.83	56.17	68.00	37.14	28.14	40.14	48.17	180.80	33.78	12.70	11.56
Th/U	3.43	5.17	5.43	4.42	5.00	3.67	5.29	7.20	4.91	3.50	3.56	2.70
Sr/Ca	.002	.002	.010	.006	.006	.002	.004	.005	.009	.004	.008	.009
Ba/Sr	4.38	.33	1.16	.50	.86	1.00	.86	.83	4.00	1.00	3.61	2.92

TABLE 4.1 Cont'd

Rock Type:	Comendite									
Sample No:	6647	6648	6650	6659	6652	6640	6641	6653	6654	6658
SiO ₂	74.71	74.25	75.65	73.60	75.22	74.37	74.24	74.80	74.68	70.22
TiO ₂	.09	.09	.13	.06	.10	.15	.15	.09	.08	.14
Al ₂ O ₃	12.04	14.28	11.47	13.32	11.98	10.95	10.80	12.17	13.95	16.18
Fe ₂ O ₃	1.89	2.40	2.10	2.20	2.24	3.26	3.89	1.97	1.75	.98
FeO	1.54	1.18	1.35	.73	1.02	1.34	1.12	1.55	.80	.55
MnO	.05	.05	.05	.08	.04	.05	.06	.06	.04	.04
MgO	.04	.02	.05	.03	.08	.05	.04	-	.04	.05
CaO	.09	.05	.10	.08	.12	.20	.22	.09	.09	.23
Na ₂ O	5.21	4.05	4.81	5.84	5.01	4.99	5.19	4.93	4.45	6.42
K ₂ O	4.34	3.60	4.27	4.05	4.17	4.64	4.27	4.32	4.11	5.17
P ₂ O ₅	.01	.02	.02	.01	.02	.02	.02	.01	.02	.02
$\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{2+}}$	2.17	1.06	2.59	1.82	4.51	1.98	1.53	-	2.94	5.69
Q	27.92	36.81	31.57	24.71	30.38	29.34	29.44	29.37	32.87	12.79
Or	25.67	21.27	25.20	24.00	24.66	27.40	25.20	25.55	24.29	30.55
Ab	37.75	34.25	35.28	46.14	38.35	30.49	31.83	38.54	37.60	54.30
An	-	.12	-	-	-	-	-	-	.39	.05
C	-	3.68	-	-	-	-	-	-	-	-
Ac	5.46	-	4.73	2.99	3.56	9.42	10.64	2.75	2.06	-
Na	.02	-	-	-	-	.23	-	-	-	-
Di	.31	-	.32	-	.40	.76	.86	.28	-	.35
Wo	-	-	-	-	-	-	-	-	-	.22
Hy	2.68	0.18	1.93	.33	.93	2.03	1.39	1.81	.10	-
Wt	-	3.48	.67	1.70	1.48	-	.31	1.48	2.47	1.42
Hm	-	-	-	-	-	-	-	-	.05	-
Il	.17	.17	.25	.11	.19	.28	.28	.17	.15	.27
Ap	.02	.05	.05	.02	.05	.05	.05	.05	.02	.05
$\frac{100 \text{ An}}{\text{An} + \text{Ab}}$	-	.34	-	-	-	-	-	-	1.02	.10
F.I.	96.82	96.01	96.77	97.84	96.95	96.88	97.11	96.21	96.82	97.64
Ba	6	3	2	5	6	2	3	2	14	15
Rb	422	445	311	575	73	416	353	431	503	215
Sr	5	8	5	6	37	5	6	6	7	9
Y	82	149	166	71	262	95	53	155	194	66
U	8	14	11	12	29	10	7	16	20	9
Th	39	51	37	100	86	44	48	39	58	24
Zr	1519	1862	1849	2543	1680	2848	3067	1625	1009	666
V	ND	ND	ND	ND	<1	<1	ND	ND	ND	ND
Cr	ND	ND	ND	ND	<1	<1	ND	ND	ND	ND
Ni	ND	ND	ND	ND	1	16	ND	ND	ND	ND
Cu	ND	ND	ND	ND	<1	7	ND	ND	ND	ND
Zn	ND	ND	ND	ND	811	237	ND	ND	ND	ND
Ga	83	100	62	92	68	79	77	80	76	51
Pb	27	37	12	32	64	30	26	28	70	9
K/Pb	85	65	113	58	469	92	99	83	67	200
Rb/Sr	84.40	55.63	62.20	95.83	1.97	83.20	58.83	71.83	71.86	23.89
Th/U	4.88	3.64	3.36	8.33	2.97	4.40	6.86	2.44	2.90	2.67
Sr/Ca	.008	.020	.007	.010	.041	.004	.004	.010	.012	.006
Ba/Sr	1.20	.38	.40	.83	.16	.40	.50	.33	2.00	1.67

TABLE 4.2

MAJOR AND TRACE ELEMENT ANALYSES OF SUB-ALKALINE ROCKS

Rock Type:	Andesite						Diorite				Dioritic Xenoliths				Granodiorite					
Sample No:	6577	6580	6589	6584	6585	6587	6586	6587	6581	6580	6570	6569	6571	6568	6702	6700	6593	6599	6581	
SiO ₂	52.58	56.10	50.64	60.96	56.94	57.79	66.22	71.47	62.51	65.29	55.71	61.73	62.62	63.96	64.25	65.99	64.52	65.84	65.69	
TiO ₂	2.08	1.47	2.48	1.02	1.13	.83	.83	.40	.92	.80	1.81	1.37	1.29	1.18	1.10	1.11	1.05	.86	1.12	
Al ₂ O ₃	14.94	13.95	17.24	18.30	18.02	19.02	15.37	15.05	17.20	16.49	16.74	15.96	15.94	16.52	15.88	15.85	15.83	15.61	16.25	
Fe ₂ O ₃	2.71	2.85	1.74	2.12	2.39	1.99	.94	.85	1.07	.94	2.27	1.28	.94	2.37	1.49	.55	1.46	1.28	1.42	
FeO	9.05	9.50	10.27	4.05	4.37	3.65	8.38	3.07	3.87	3.08	7.05	5.31	5.26	3.39	4.22	4.25	3.67	3.71	3.23	
MnO	.19	.23	.37	.11	.12	.14	.26	.40	.12	.09	.23	.16	.20	.14	.12	.10	.12	.13	.07	
MgO	5.26	3.89	5.70	2.56	3.90	3.69	1.20	.42	3.56	2.36	3.67	2.16	1.88	1.53	1.68	1.23	1.63	1.30	1.87	
CaO	8.49	5.23	7.49	5.36	7.99	8.23	.60	.24	3.28	3.78	5.36	4.78	3.10	2.42	4.14	3.14	4.03	3.27	2.84	
Na ₂ O	3.35	5.36	3.19	3.95	3.77	3.70	4.04	4.77	5.83	4.36	6.12	4.88	5.53	5.14	4.43	4.95	4.71	4.94	4.92	
K ₂ O	.97	.71	.36	1.32	1.15	.75	1.69	3.22	1.41	2.61	.69	1.98	2.75	2.05	2.38	2.60	2.40	2.75	2.33	
P ₂ O ₅	.38	.71	.52	.25	.22	.19	.27	.11	.23	.21	.35	.39	.48	.50	.31	.23	.38	.31	.26	
$\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{2+}}$	50.90	42.21	49.74	58.30	61.40	64.30	20.36	19.59	62.10	57.69	48.12	42.04	38.93	44.54	41.53	34.00	42.84	38.42	50.77	
Q	2.92	3.30	1.73	16.44	8.38	10.37	27.08	28.35	9.33	17.26	-	11.50	9.02	15.96	17.13	16.95	16.52	15.91	18.63	
Or	5.73	4.19	2.13	7.80	6.79	4.48	11.15	19.04	8.30	15.40	4.09	11.70	16.23	16.81	14.05	15.34	14.15	16.24	13.78	
Ab	28.33	45.33	27.01	33.41	31.85	31.33	34.12	40.33	49.26	36.86	51.75	41.31	46.74	43.44	37.48	41.84	39.84	41.74	41.63	
An	22.87	11.92	31.64	24.93	28.87	33.05	1.19	.44	14.80	17.41	16.18	15.78	10.59	8.78	16.42	13.39	14.98	12.32	12.39	
C	-	-	-	1.24	-	-	6.27	3.55	.67	.13	-	-	-	1.78	-	-	-	-	1.09	
Ol	13.60	7.85	1.71	-	7.50	5.24	-	-	-	-	6.71	4.49	1.42	-	1.76	.54	2.10	1.56	-	
Hy	17.59	18.84	27.35	10.58	10.49	10.61	16.62	6.04	13.81	9.58	12.06	9.87	11.05	6.40	8.21	8.49	7.41	7.02	7.70	
Ol	-	-	-	-	-	-	-	-	-	-	1.66	-	-	-	-	-	-	-	-	
Mt	3.93	4.13	2.52	3.08	3.46	2.89	1.36	1.24	1.56	1.36	3.30	1.85	1.39	3.44	2.15	.79	2.13	1.86	2.06	
Il	3.95	2.79	4.70	1.93	2.15	1.58	1.58	.75	1.75	1.52	3.43	2.59	2.46	2.24	2.09	2.11	2.00	1.63	2.12	
Ap	.88	1.65	1.21	.59	.51	.45	.63	.26	.53	.48	.82	.91	1.10	1.15	.71	.54	.88	.72	.60	
$\frac{100 \text{ An}}{\text{An} + \text{Ab}}$	44.66	20.81	53.96	42.74	47.54	51.34	3.37	1.08	23.11	32.08	23.83	27.65	23.83	16.82	30.47	24.24	27.31	22.79	22.93	
D.I.	36.98	52.82	30.87	57.65	47.02	45.18	72.35	87.72	66.89	69.52	55.84	64.51	71.99	76.21	68.66	74.13	70.51	74.69	74.04	
Ba	145	97	101	229	185	192	236	860	752	603	69	325	269	282	205	331	294	394	412	
Rb	40	16	5	34	26	21	66	82	27	49	21	44	73	89	64	68	63	65	56	
Sr	386	170	619	403	440	457	153	204	621	461	204	327	204	330	242	258	294	266	378	
Y	21	32	38	92	21	20	29	29	22	19	39	40	52	42	49	50	42	45	31	
U	1	1	1	2	1	2	4	3	1	2	2	2	2	3	4	2	1	1	1	
Th	10	6	5	6	7	8	9	11	8	9	4	9	9	10	12	6	12	10	11	
Zr	155	98	223	167	166	154	225	232	185	210	231	318	257	168	215	271	263	370	196	
V	ND	ND	223	131	138	103	59	15	ND	ND	ND	ND	58	ND	81	ND	ND	40	ND	
Cr	ND	ND	24	37	20	47	<1	<1	ND	ND	ND	ND	2	ND	6	ND	ND	<1	ND	
Ni	ND	ND	12	12	11	20	<1	<1	ND	ND	ND	ND	<1	ND	5	ND	ND	5	ND	
Cu	ND	ND	17	35	40	23	12	13	ND	ND	ND	ND	22	ND	16	ND	ND	30	ND	
Zn	ND	ND	93	52	46	50	55	148	ND	ND	ND	ND	89	ND	86	ND	ND	75	ND	
Ga	25	18	13	20	19	18	19	16	17	16	21	20	20	22	20	21	22	21	17	
Pb	5	<1	5	6	7	6	13	41	4	7	5	7	13	10	16	15	16	15	7	
K/Rb	200	368	580	317	364	296	225	313	418	423	258	374	308	263	311	317	315	351	331	
Rb/Sr	.10	.09	.01	.08	.06	.05	.43	.40	.04	.11	.10	.13	.36	.27	.26	.26	.21	.24	.15	
Th/U	10.00	6.00	5.00	3.00	7.00	4.00	4.50	3.67	8.00	4.50	2.00	4.50	4.50	3.33	3.00	3.00	12.00	10.00	11.00	
Sr/Ca	.006	.005	.010	.011	.008	.008	.037	.126	.027	.018	.005	.009	.009	.019	.008	.012	.010	.011	.019	
Ba/Sr	.38	.57	.19	.57	.42	.42	1.54	4.22	1.21	1.31	.34	.99	1.32	.85	.85	1.28	1.00	1.48	1.09	

TABLE 4.2 Cont'd

Rock Type:	Grenodiorite			Granite												Pitchstone		Rhyolite	
Sample No:	6679	6680	6695	6696	6697	6701	6704	6705	6704	6703	6707	6702	6708	6661	6653	6690	6692	6703	6693
SiO ₂	67.63	67.09	69.03	70.04	71.23	73.25	75.10	71.56	70.59	70.74	71.73	72.10	73.96	75.66	76.11	75.23	74.66	71.51	77.23
TiO ₂	.79	.69	.67	.47	.42	.35	.26	.34	.39	.41	.37	.34	.28	.09	.10	.28	.29	.57	.11
Al ₂ O ₃	15.79	15.51	15.00	15.02	15.38	14.19	14.03	14.44	14.02	14.83	14.58	14.45	14.05	13.11	12.95	13.70	13.52	15.60	11.98
Fe ₂ O ₃	2.15	1.87	1.24	.68	1.34	.44	.93	1.05	1.24	1.31	1.25	1.13	.58	1.12	1.41	.72	.65	.83	1.17
FeO	2.17	2.45	2.21	2.57	1.40	1.35	.33	1.58	1.73	1.70	1.51	1.41	1.22	.15	.17	1.17	1.11	.35	1.03
MnO	.07	.13	.12	.08	.05	.06	.04	.08	.09	.14	.09	.08	.06	.02	.01	.05	.06	.62	.07
MgO	.74	.91	.59	.66	.41	.33	.12	.45	.45	.51	.45	.36	.27	.10	.05	.25	.34	.43	.10
CaO	1.93	2.78	2.08	1.87	1.35	1.02	.89	1.32	1.60	1.37	1.12	1.15	.34	.08	.06	1.62	1.30	.26	.09
Na ₂ O	5.58	5.44	5.72	5.08	5.37	4.57	4.76	5.32	5.24	5.34	5.23	5.03	4.64	4.57	4.53	4.44	4.43	5.82	3.46
K ₂ O	2.89	2.91	2.51	3.44	2.92	4.35	3.54	3.78	3.59	3.55	3.58	3.67	4.54	5.09	4.61	2.90	3.56	3.93	4.74
P ₂ O ₅	.26	.22	.18	.09	.10	.09	.03	.08	.26	.10	.09	.08	.06	.01	.01	.04	.06	.08	.02
$\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{2+}}$	37.86	39.86	32.16	31.42	34.34	30.37	39.47	33.75	31.73	34.71	34.78	31.23	28.27	54.35	33.33	27.56	35.29	68.59	14.88
Q	19.21	17.40	20.42	21.17	25.53	26.84	31.87	22.50	22.24	22.03	24.23	24.68	28.03	29.34	31.98	35.59	31.63	21.57	38.28
Or	17.07	17.20	14.95	20.34	17.46	25.69	20.91	22.33	21.20	20.94	21.17	22.89	26.82	30.09	27.23	14.74	21.02	23.24	27.99
Ab	47.15	45.96	48.35	42.95	45.98	38.65	40.27	44.98	44.34	45.13	44.20	42.59	39.23	38.64	38.30	37.55	37.45	49.22	29.30
An	7.85	9.34	9.15	8.02	6.11	4.55	4.20	4.37	6.25	6.04	4.98	5.20	1.28	.24	.23	7.75	6.06	.78	.32
C	.61	-	.12	-	-	.30	.83	-	.03	-	.29	-	1.05	-	.43	.86	.17	1.49	1.04
Di	-	2.47	-	.55	-	-	-	1.37	-	.11	-	-	-	.07	-	-	-	-	-
Hy	2.68	3.10	3.62	4.88	1.92	2.48	.30	2.08	2.80	2.84	2.41	2.15	2.09	.22	.12	1.84	1.98	1.24	1.12
Mt	3.12	2.71	1.80	.99	1.96	.64	.44	1.52	1.80	1.90	1.81	1.65	.83	.29	.29	1.05	.94	1.20	1.69
Hm	-	-	-	-	-	-	.63	-	-	-	-	-	-	.92	1.21	-	-	-	-
Il	1.49	1.30	1.27	.89	.81	.66	.49	.65	.74	.78	.70	.65	.53	.17	.19	.52	.55	1.07	.21
Ap	.60	.52	.42	.21	.23	.19	.07	.20	.60	.23	.21	.19	.14	.02	.02	.10	.15	.19	.05
$\frac{100 \text{ An}}{\text{An} + \text{Ab}}$	14.31	16.88	15.92	15.73	11.74	10.53	9.49	8.85	12.36	11.80	10.11	10.88	3.18	.62	.60	17.10	13.93	1.55	1.06
O.I.	83.43	80.55	83.62	84.46	88.97	91.18	93.05	89.81	87.78	88.10	89.60	90.16	94.08	98.07	97.51	87.88	90.15	94.03	95.57
Ba	430	305	363	424	319	549	389	384	386	436	408	406	366	220	331	448	423	410	191
Rb	59	77	58	87	109	108	86	103	92	93	95	104	114	107	89	184	112	104	132
Sr	202	153	249	152	208	165	94	133	126	135	110	115	56	8	11	332	121	169	40
Y	54	50	74	39	59	33	47	36	36	37	31	35	31	84	104	30	28	40	53
U	4	3	2	3	5	3	4	4	2	2	3	4	2	3	4	4	3	4	3
Th	11	10	10	10	14	11	12	12	11	12	12	104	15	11	11	12	13	13	16
Zr	461	485	361	383	562	304	363	306	319	353	319	312	278	667	756	228	213	482	284
V	ND	ND	ND	ND	ND	7	ND	ND	ND	ND	ND	ND	<1	<1	ND	ND	ND	ND	ND
Cr	ND	ND	ND	ND	ND	<1	ND	ND	ND	ND	ND	ND	<1	<1	ND	ND	ND	ND	ND
Ni	ND	ND	ND	ND	ND	1	ND	ND	ND	ND	ND	ND	<1	1	ND	ND	ND	ND	ND
Cu	ND	ND	ND	ND	ND	18	ND	ND	ND	ND	ND	13	13	ND	ND	ND	ND	ND	ND
Zn	ND	ND	ND	ND	ND	46	ND	ND	ND	ND	ND	37	39	ND	ND	ND	ND	ND	ND
Ga	23	21	20	20	31	18	19	19	19	20	17	19	18	29	25	17	14	18	18
Pb	13	21	17	6	12	20	15	12	13	13	12	10	7	13	8	14	12	13	19
K/Rb	404	311	359	329	222	334	344	297	323	318	313	338	332	468	431	106	255	311	299
Rb/Sr	.29	.50	.23	.57	.52	.65	.91	.77	.73	.69	.86	.90	2.04	13.38	8.09	.55	.93	.62	3.30
Th/U	2.75	3.33	5.00	3.33	2.80	3.67	3.00	3.00	6.50	6.50	4.00	3.25	7.50	3.67	2.75	3.00	3.25	3.25	5.33
Sr/Ca	.015	.003	.017	.011	.021	.023	.015	.014	.011	.014	.014	.014	.023	.009	.028	.030	.013	.039	.067
Ba/Sr	2.13	1.99	1.46	2.79	1.53	3.33	4.14	2.89	3.06	3.23	3.71	3.53	6.54	27.50	30.09	1.35	3.50	2.43	4.77

4.2 Major Element Analyses

4.2.1 Alkaline Rocks

4.2.1.1 Classification

Although the Comboyne rocks consist primarily of intermediate and silicic types, and rocks more basic than hawaiite have not been found, the series exhibits a wide variation in chemical composition and includes hawaiite, mugearite, benmoreite, mafic trachyte, anorthoclase trachyte and comendite.

Classification of the series is based on the scheme proposed by Coombs and Wilkinson (1969) in which Differentiation Index (Thornton and Tuttle, 1960) is plotted against normative plagioclase anorthite value. Figure 4.1a shows that the Comboyne rocks follow a similar trend to that exhibited by the Hawaiian alkaline series. The Nandewar rocks, which are closely comparable to those at Comboyne, show some scattering between the Gough Island and Hawaiian trends. Plots of the Comboyne rocks on this diagram, identify the series as sodic rather than potassic. The Hebridean alkaline rocks are slightly more sodic than the Comboyne, Hawaiian and Nandewar rocks, and these contrast with the more potassic Tristan da Cunha and Gough Island associations. Sodic and potassic associations are also delineated on a $\text{Na}_2\text{O}/\text{K}_2\text{O}$ versus SiO_2 diagram (Fig. 4.2), and again the Comboyne series falls in the sodic field but is less sodic than the Skye Eocene lavas (Thompson et al., 1972), the Atlantic islands of Madeira, Bouvet and Ascension (Schmincke, 1973), and the Ethiopian Danakil Depression rocks (Barberi et al., 1970). On this diagram the mildly alkaline associations of Comboyne, Nandewar (Abbott, 1969) and Aden and Little Aden (Cox et al., 1970) exhibit similar values and trends.

Fig. 4.1a Plots of Comboyne alkaline rocks on Differentiation Index versus Normative Plagioclase composition diagram, and classification adopted for these rocks (Coombs and Wilkinson, 1969). Comparative generalised trend lines for Hebridean, Hawaiian, Gough Island and Tristan da Cunha series taken from Coombs and Wilkinson, 1969. Solid circles = Comboyne alkaline rocks, Crosses = Nandewar Mountain alkaline series (Abbott, 1969).

Fig. 4.1b Plots of K_2O versus Na_2O . Solid circles = Comboyne hawaiites, mugearites, benmoreites and trachytes. Open circles = Comboyne comendites. Crosses = Nandewar Mountain alkaline series (Abbott, 1969). Generalised trend lines for Yellowstone, Gough Island, Tristan da Cunha, Hawaiian and Hebridean provinces taken from Coombs and Wilkinson (1969).

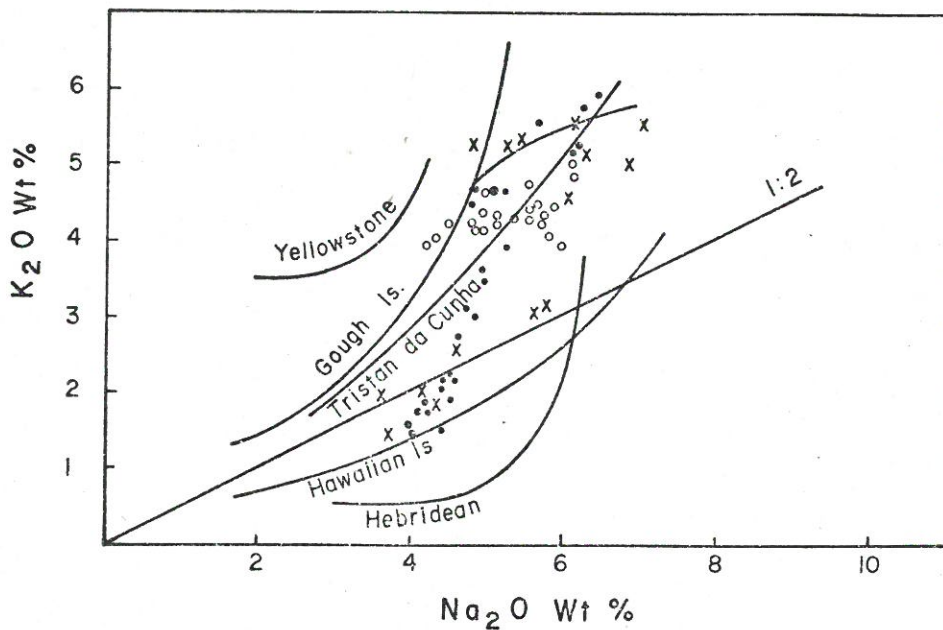
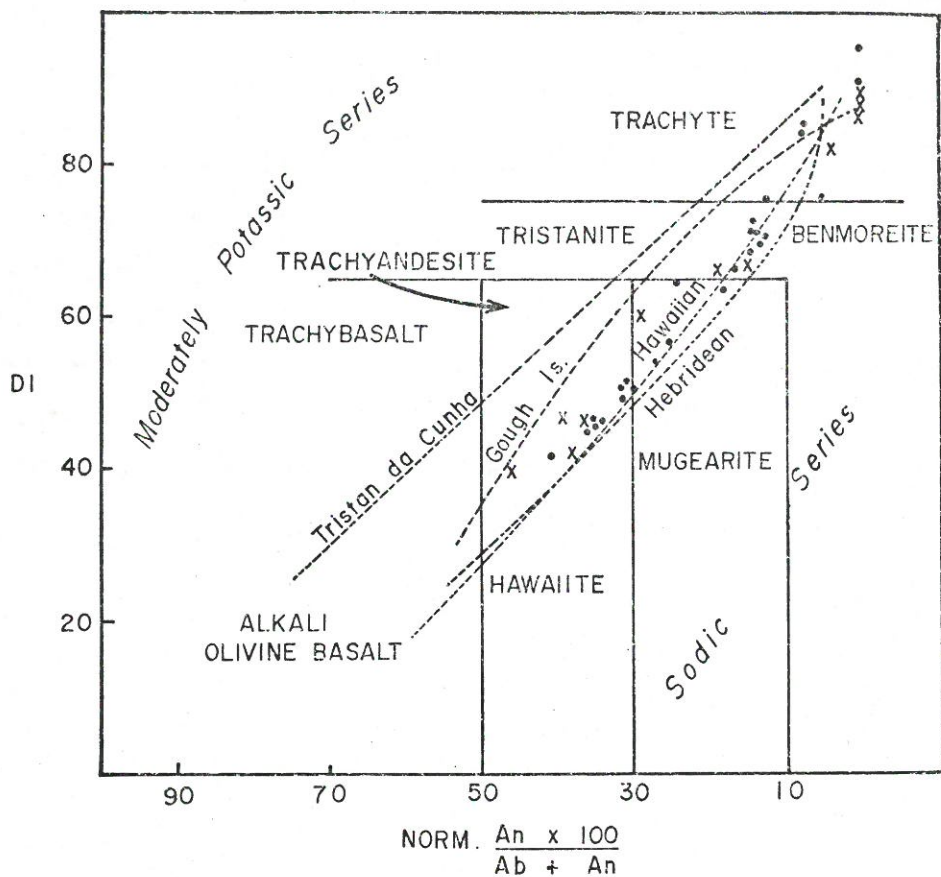
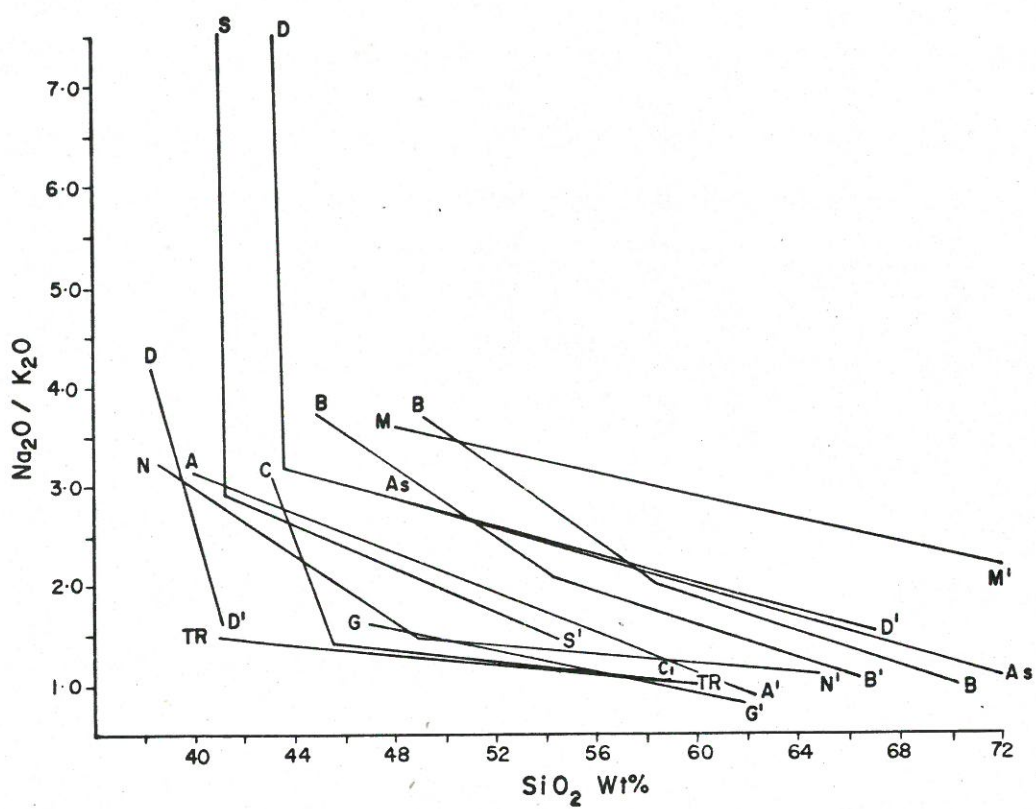


Fig. 4.2 Generalised trend lines for $\text{Na}_2\text{O}/\text{K}_2\text{O}$ plotted against SiO_2 . Sodic trends are exhibited by D - D', Danakil (Barberi et al., 1970); N - N', Nandewar (Abbott, 1969); A - A', Aden and Little Aden (Cox et al., 1970); S - S', Skye (Thompson et al., 1972); D - D', Dogo (Uchimizu, 1966); B - B', Bouvet, M - M' Madiera, As - As', Ascension (Schmincke, 1973); and contrast with the more potassic G - G', Gough Island and TR - TR', Tristan da Cunha associations (Schmincke, 1973). The Comboyne alkaline rocks C - C' follow a sodic trend.



4.2.1.2 Variation Diagrams and Comparison with other

Basaltic Associations

The plotting of total alkalis against silica (Fig. 4.3a) emphasises the alkaline affinities of the Comboyne series. This figure shows there is not a continuous variation in silica and alkali content, but rather a grouping of hawaiite and mugearite, benmoreite and mafic trachyte, and finally anorthoclase trachyte and comendite. The Nandewar series, which includes more basic rock types, also exhibits a grouping of rock types in similar fields to the Comboyne series. The trend of the Comboyne rocks on a total alkalis versus silica diagram is comparable to that of the Madeira, Sao Miguel and Terceira islands (Schmincke and Weibel, 1972), the Isle of Skye (Thompson et al., 1972), Hawaiian (Macdonald and Katsura, 1964), Nandewar (Abbott, 1969), and Silali (McCall and Hornung, 1972) alkaline provinces. Alkaline rocks from Gough Island (Le Maitre, 1962), Tristan da Cunha (Baker et al. 1964), Lake Rudolf (Brown and Carmichael, 1971) and Marquesas Islands (Bishop and Woolley, 1973), are more strongly alkaline in character. Rocks from the Danakil Depression (Barberi et al., 1970), Aden and Little Aden (Cox et al., 1970), Dogo, Iki Islands (Uchimizu, 1966), are less alkaline and exhibit a trend intermediate to that shown by the Nandewar and Comboyne series and that of tholeiitic associations including the Hebridean (Holland and Brown, 1972), Thingmuli (Carmichael, 1964), and Ferrar (Gunn, 1962) series (Fig. 4.4).

When Na_2O is plotted against K_2O (Fig. 4.1b) it is apparent that only the more basic hawaiites and mugearites in the Comboyne series have $\text{Na}_2\text{O}:\text{K}_2\text{O}$ ratios greater than two. The more evolved rock types exhibit strong enrichment in K_2O relative to Na_2O and the trend crosses the 1:2 boundary with a large angle.

Fig. 4.3a Plots of $\text{Na}_2\text{O} + \text{K}_2\text{O}$ versus SiO_2 for the Comboyne alkaline rocks. Dashed lines are the boundaries for Kuno's 'alkaline, hypersthenic and pigeonitic' series (Kunc, 1968).

Fig. 4.3b Plots of $\text{Na}_2\text{O} + \text{K}_2\text{O}$ versus SiO_2 for the Comboyne Plateau - Lorne Basin sub-alkaline rocks.

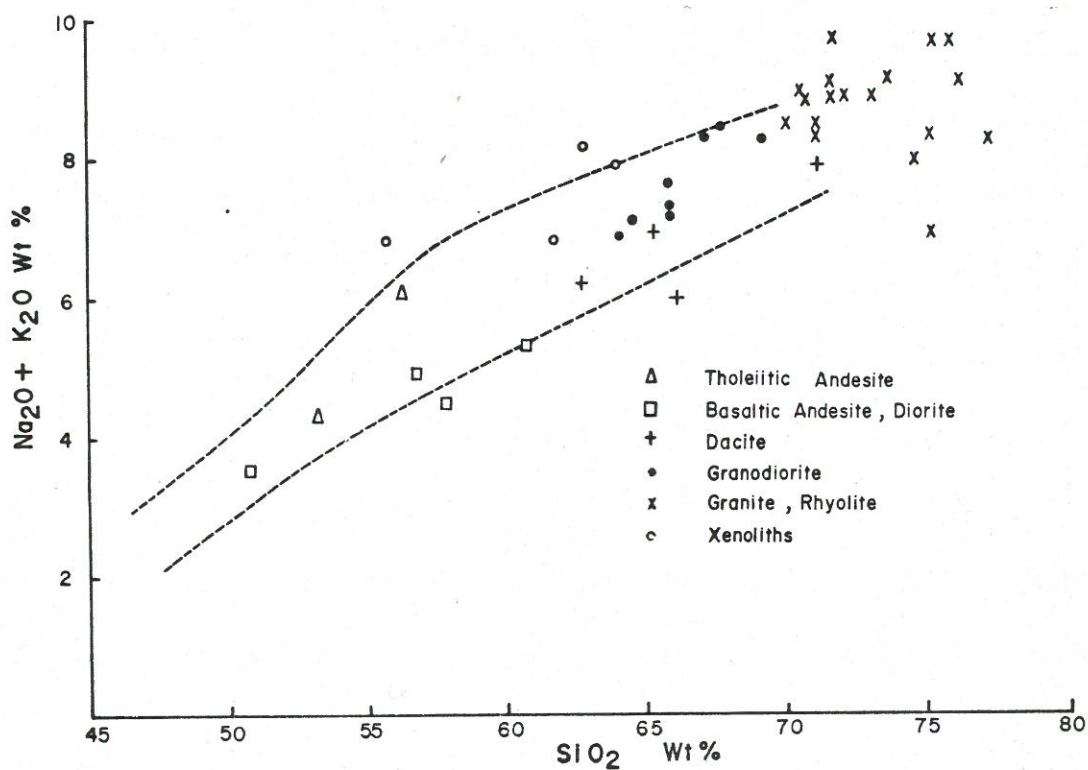
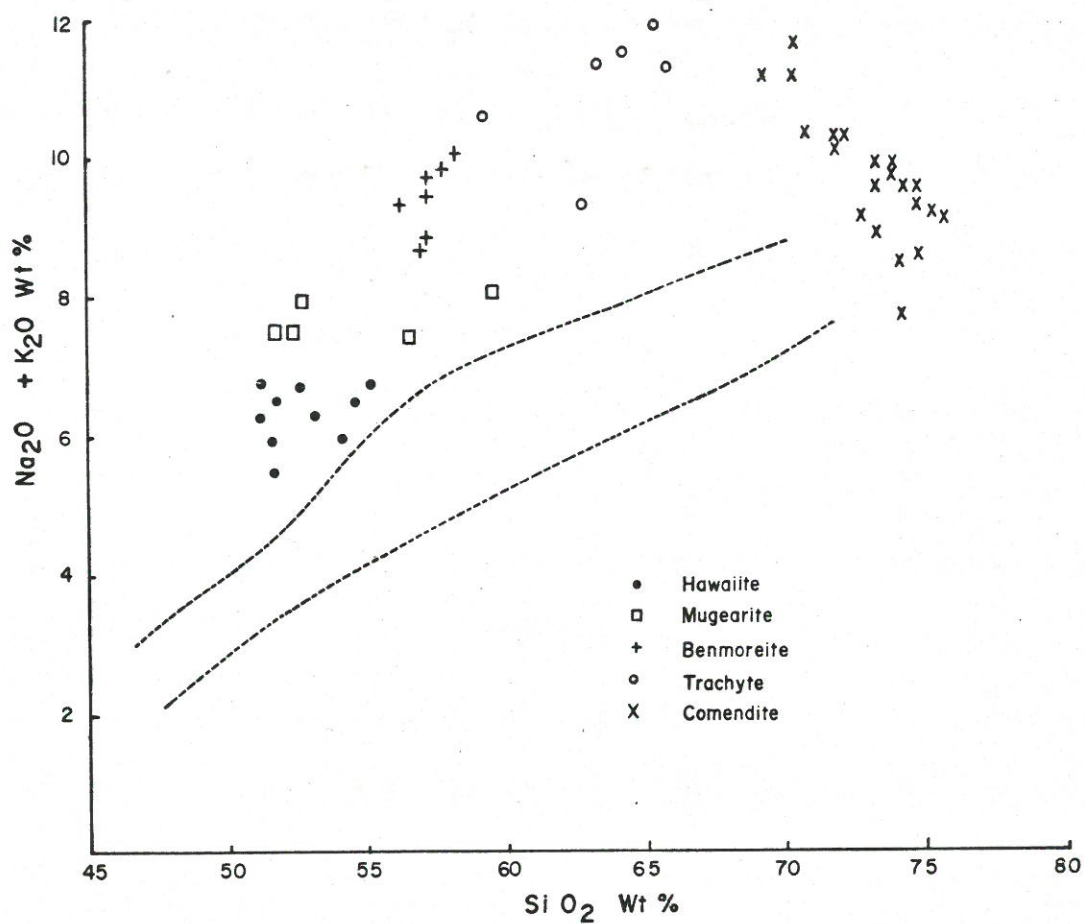
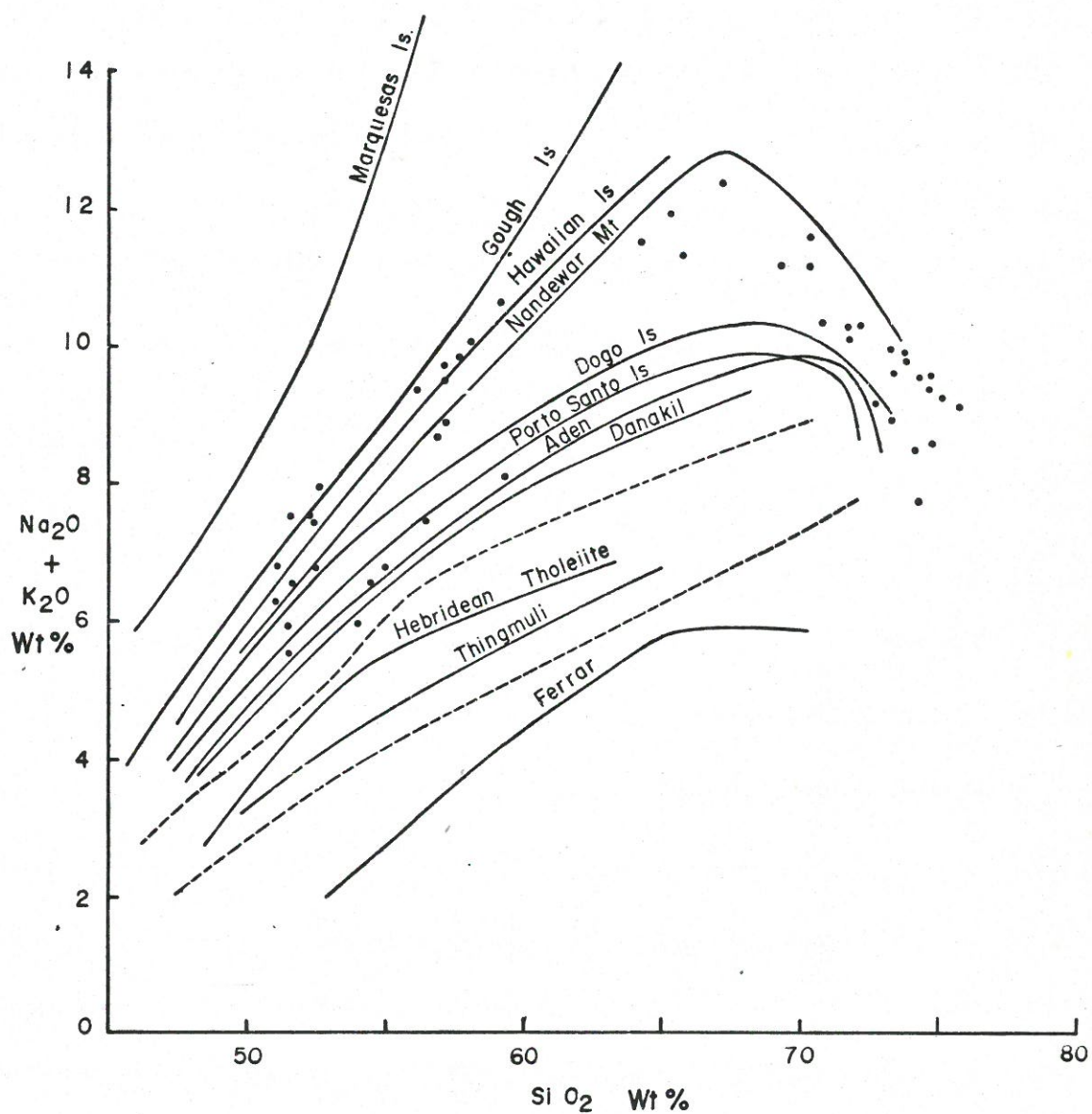


Fig. 4.4 Comparative plot of $\text{Na}_2\text{O} + \text{K}_2\text{O}$ versus SiO_2 for a selection of alkaline, 'transitional', and tholeiitic provinces with plots of Comboyne alkaline-transitional rocks (solid circles) superimposed. Generalised trend lines derived from data as follows: Marquesas Islands (Bishop and Woolley, 1973), Gough Island (Le Maitre, 1962), Hawaiian Islands (Macdonald and Katsura, 1964), Nandewar Mountain (Abbott, 1969), Dogo (Uchimizu, 1966), Porto Santo (Schmincke and Weibel, 1972), Aden (Cox et al., 1970), Danakil (Barberi et al., 1970), Hebridean tholeiite (Holland and Brown, 1972), Thingmuli (Carmichael, 1964), Ferrar (Gunn, 1962). Dashed lines are the boundaries of Kuno's (1968) 'alkaline, hypersthenic and pigeonitic series'.



A comparison of alkaline associations indicates three main fractionation paths can be identified:

- 1) Potassic Series, for example Tristan da Cunha, Gough Island - have $\text{Na}_2\text{O}:\text{K}_2\text{O}$ ratios of less than one throughout the series.
- 2) Sodic Series, for example Hawaiian and Hebridean associations - show initial enrichment in Na_2O relative to K_2O , with the trend crossing the 1:2 boundary during the latter stages of differentiation.
- 3) A less markedly sodic trend is found in mildly alkaline associations such as Comboyne, Nandewar, Danakil, Aden and Little Aden, which are chemically transitional between alkaline and tholeiitic associations. In these mildly alkaline series there is an earlier enrichment in K_2O relative to Na_2O , which probably results from the abundant early crystallisation of plagioclase, which contrasts with the early dominance of ferromagnesian minerals in the more strongly alkaline associations.

4.2.1.3 Normative Mineralogy

The mildly alkaline Skye, Nandewar, Aden and Little Aden, Danakil, Terceira, Sao Miguel, Dogo and Comboyne series characteristically contain normative hypersthene and to a lesser extent, quartz. The presence of normative hypersthene was considered by Yoder and Tilley (1962) to identify a rock as tholeiitic, and the occurrence of normative hypersthene in rocks which mineralogically were alkaline was attributed to deuteric alteration resulting in an inflated $\text{Fe}_2\text{O}_3:\text{FeO}$ ratio. It is generally accepted that in the majority of basaltic rocks deuteric alteration results in a $\text{Fe}_2\text{O}_3:\text{FeO}$ ratio greater than

that of the pre-eruptive melt (Coombs, 1963). However, despite adjustment to these ratios (page 102) in an attempt to correct for such effects, normative hypersthene occurs in all but one of the Comboyne rocks and varies from 0.00 to 15.56%. Normative quartz is rare in hawaiites and increases with differentiation as normative hypersthene decreases. The adjustment to the $\text{Fe}_2\text{O}_3:\text{FeO}$ ratio in the Comboyne hawaiites has resulted in an increase in normative hypersthene and a decrease in normative quartz, and normative hypersthene tends to be higher than in other mildly alkaline associations. The majority of analyses published from mildly alkaline associations have rather high $\text{Fe}_2\text{O}_3:\text{FeO}$ ratios and normative quartz tends to be higher than in the Comboyne rocks. A correction to lower these $\text{Fe}_2\text{O}_3:\text{FeO}$ ratios results in either a decrease in normative quartz and an increase in normative hypersthene to values more directly comparable with those in the Comboyne series or, in some instances, the appearance of minor normative nepheline.

Alkaline rocks near-saturated in SiO_2 are not uncommon and although they are normatively transitional to tholeiitic basalts, they are mineralogically typically alkaline, plot in the alkaline field in $\text{Na}_2\text{O} + \text{K}_2\text{O}$ versus SiO_2 diagrams, show only minor iron enrichment on FMA diagrams and are commonly associated with peralkaline rhyolites. The confusion caused by the complete transition between mildly alkaline and tholeiitic rocks is emphasised by the classification of the Bouvet Islands by Kempe and Schilling (1974) as tholeiitic with alkaline affinities despite the absence of a typically tholeiitic mineralogy. Similarly, Baker et al. (1964, 1974) described the Ascension, Bouvet and Easter Island basalts as tholeiitic, but in the case of Easter Island named the rocks according to terminology generally

applied to alkaline rocks. In contrast, Borley (1974) classified all the above associations as alkaline.

The mineralogical characteristics of these normatively tholeiitic rocks, including the absence of groundmass pigeonite and a reaction relationship between olivine and a Ca-poor pyroxene, together with the presence of groundmass and, more rarely, phenocryst alkali-feldspar, indicates their alkaline affinities, as does the reaction relationship between orthopyroxene and the host melt. These affinities are also emphasised on total alkalis versus SiO_2 diagrams and the ratio of total alkalis to CaO . The importance of modal characteristics was acknowledged by Yoder and Tilley (1962) who, in proposing the classification of basaltic rocks on their normative components, noted that modal composition should always have precedence over normative composition.

4.2.1.4 Indicator Ratios

Coombs (1963) established that basic alkaline rocks commonly have indicator ratios (I.R.) $(\text{Hy} + 2\text{Qz})/(\text{Hy} + 2(\text{Qz} + \text{Di}))$ of less than zero. He considered basalts with indicator ratios between zero and 0.38 as being transitional in character, with typically alkaline mineralogy despite commonly having normative hypersthene. The Comboyne hawaiites have indicator ratios ranging from zero to 0.57, with the majority being less than 0.38. Although basalts with indicator ratios in excess of 0.5 were considered by Coombs to have high-alumina and tholeiitic basalt affinities, it is noted that the Comboyne hawaiites are relatively evolved and not basalts 'sensu stricto.' Coombs also observed that basalts with indicator ratios between zero and 0.5 from Reunion, Easter Island, Galapagos, Banks Peninsular, Tutuila and Ascension are closely associated with pantellerites. Peralkaline rocks are also associated with the

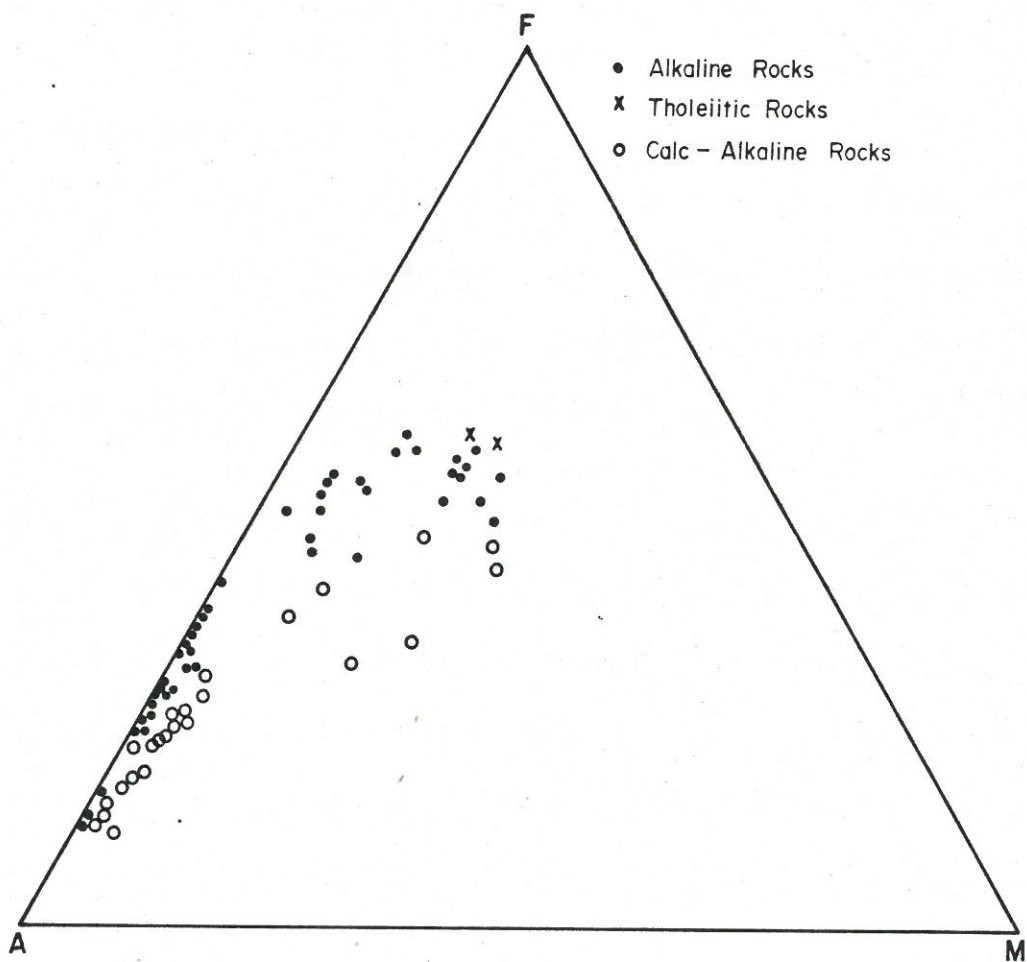
mildly alkaline-transitional Nandewar, Dogo, Aden and Little Aden, Danakil and Comboyne provinces, which further suggests a close petrogenetic relationship between these mildly alkaline and peralkaline rocks.

4.2.1.5 FMA Diagram

Plots of the Comboyne alkaline rocks on an FMA diagram (Fig. 4.5) show a trend typical of mildly alkaline associations (Skye, Hawaii, Aden and Little Aden, Danakil, Nandewar), with limited iron enrichment in intermediate rocks followed by extreme alkali enrichment. There is minor absolute as well as relative iron enrichment from hawaiite to mugearite (Fig. 4.8), with a slight decrease in total iron as differentiation proceeds to mafic trachytes. This is followed by a pronounced decrease in iron in anorthoclase trachytes and comendites. The restricted variation in iron values from hawaiite through to benmoreite results from the abundance of plagioclase and magnesium-rich megacrysts/phenocrysts relative to Fe-Ti oxides in the hawaiites and mugearites. Abundant crystallisation of Fe-Ti oxides in benmoreites and mafic trachytes is followed by a sharp decrease in total iron in the anorthoclase trachytes. The dominance of feldspar, pyroxene and olivine megacrysts/phenocrysts with bulk total silica values comparable to those of the host hawaiite can also be related to the limited variation in SiO_2 values from hawaiite to benmoreite and mafic trachyte. Prominent crystallisation of Fe-Ti oxides in benmoreites is followed by a marked increase in SiO_2 in the anorthoclase trachytes.

Figures 4.1b and 4.3a show that actual alkali enrichment proceeds only up to and including anorthoclase trachyte. At this point strong fractionation of alkali-feldspar causes a decrease in

Fig. 4.5 FMA plots for Comboyne Plateau - Lorne
Basin alkaline, tholeiitic and calc-alkaline
rocks.



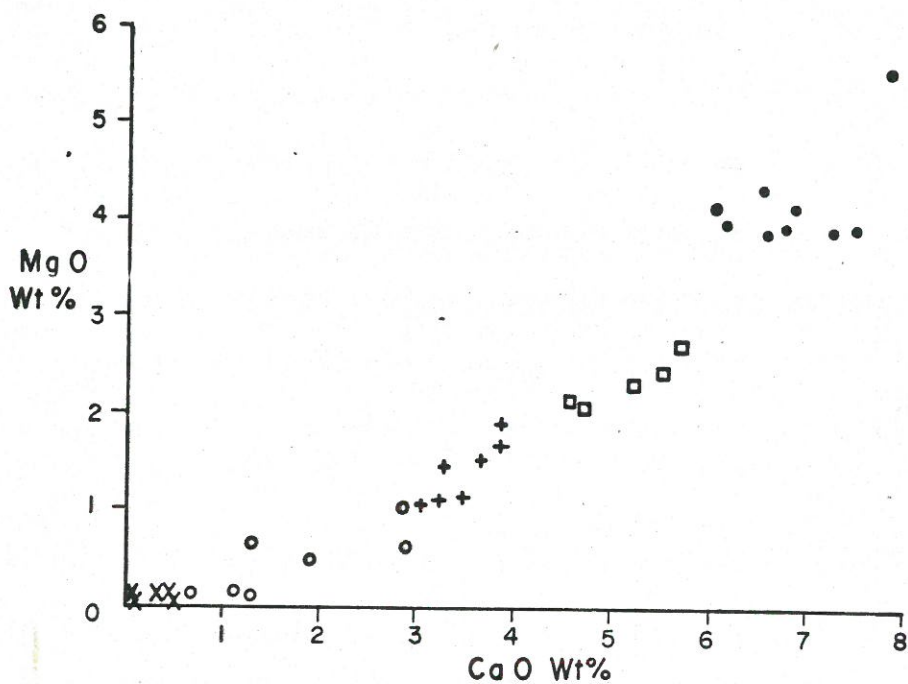
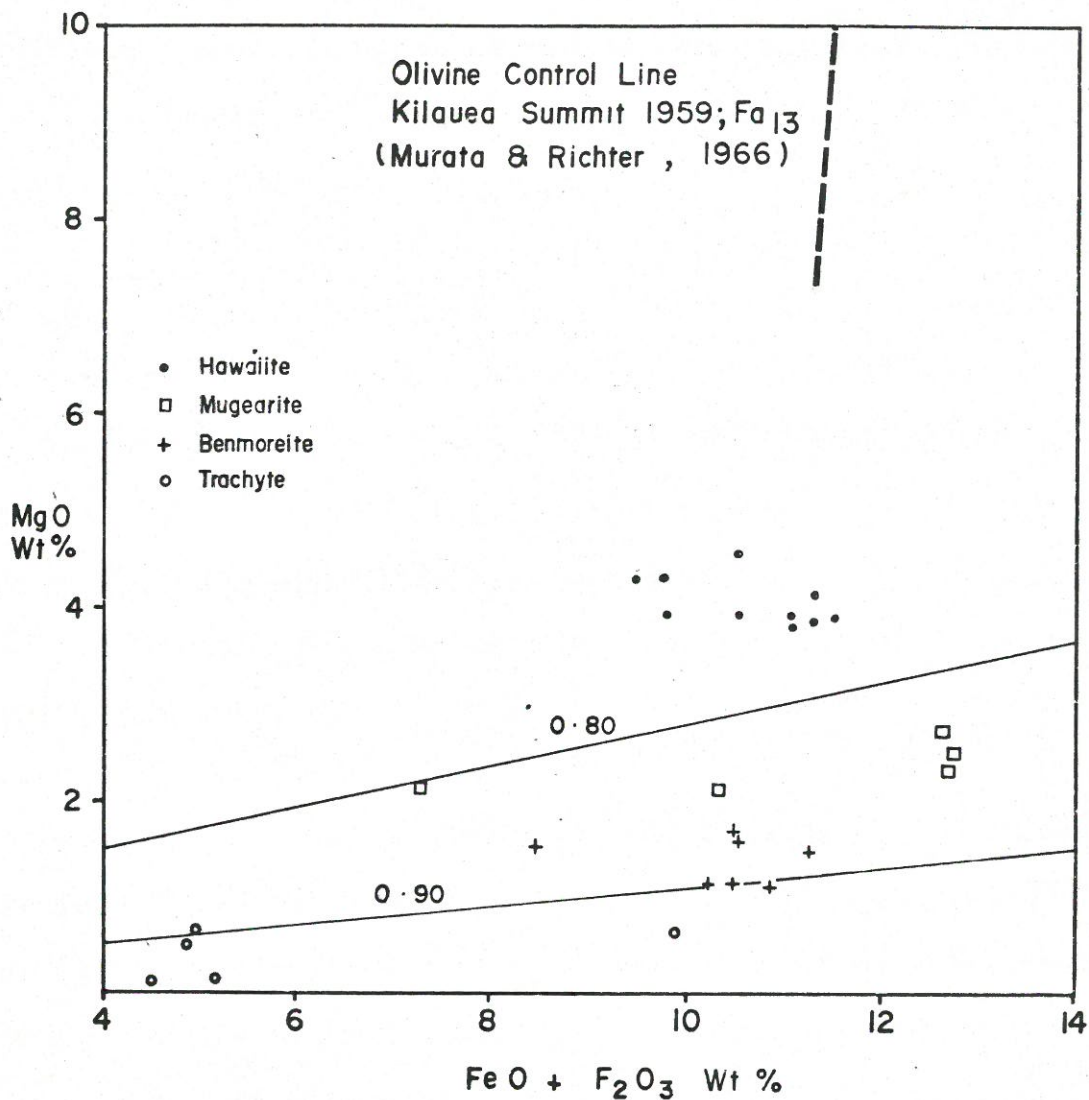
alkalis in the comendites.

The FMA differentiation trend of the mildly alkaline basalts is intermediate to that of strongly alkaline and tholeiitic basalts. In mildly alkaline associations with peralkaline differentiates (Comboyne, Nandewar, Danakil, Aden and Little Aden), marked enrichment in alkalis occurs at a more iron-rich point of the curve than is generally observed where peralkaline rocks are absent or occur as minor phases (Hawaii, Skye, Gough, Tristan da Cunha),

$100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratios in the Comboyne hawaiites vary from 49.9 to 43.8 and suggest fractionation of magnesium-rich minerals has occurred. MgO values in hawaiites vary from 4.75% to 3.72% and compare closely with values for similar rocks from Nandewar, Dogo, Danakil, Aden and Little Aden. Rapid depletion with differentiation is shown in mugearites, benmoreites and trachytes. The comendites have MgO values of less than 0.29% with the majority under 0.10%. A plot of MgO versus iron (Fig. 4.6) for the Comboyne alkaline rocks shows a broad scatter of points for the hawaiites and mugearites. Where olivine has been a major fractionation phase, olivine control lines (Murata and Richter, 1966) intersect lines representing equal iron/magnesium ratios at a large angle. Figure 4.6. shows that the trend line from hawaiite to mugearite is consistent with fractionation of magnesium-rich minerals in the hawaiite, and this is confirmed by the prevalence of olivine and orthopyroxene megacrysts, and the occurrence of olivine phenocrysts in these rocks. Plots of mugearite to trachyte on Figure 4.6 are at an angle to those from hawaiite to mugearite and fall between lines representing constant iron/magnesium ratios of 0.8 and 0.9. This is consistent with the scarcity of magnesium-rich megacrysts/phenocrysts in the mugearites, benmoreites and trachytes. The

Fig. 4.6 Plot of MgO versus iron for Comboyne hawaiiite - mugearite - benmoreite - trachyte, showing a sharp decrease in MgO between hawaiiite and mugearite reflecting the importance of Mg-rich minerals in this fractionation interval (After Thompson et al., 1972).

Fig. 4.7 Plot of MgO versus CaO showing that whereas there is a steady depletion in CaO from hawaiiite to trachyte, there is initially a marked decrease in MgO between hawaiiite and mugearite, followed by a more gradual depletion from mugearite to trachyte.



restriction of abundant magnesium-rich megacrysts/phenocrysts to hawaiites is also demonstrated when MgO is plotted against CaO (Fig. 4.7). There is a near continuum of CaO values from hawaiite to trachyte, whereas there is a marked decrease in MgO from hawaiite to mugearite, followed by a more continuous trend from mugearite to trachyte. Extremely low MgO values in the comendites are indicative of a long history of differentiation.

4.2.1.6 Fractionation Trends and Discussion

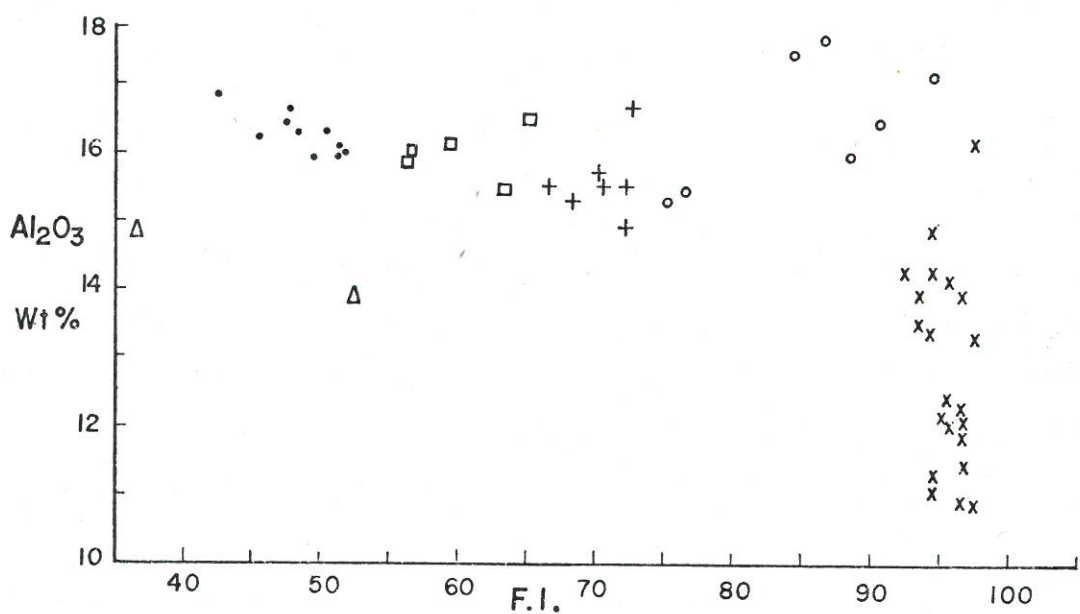
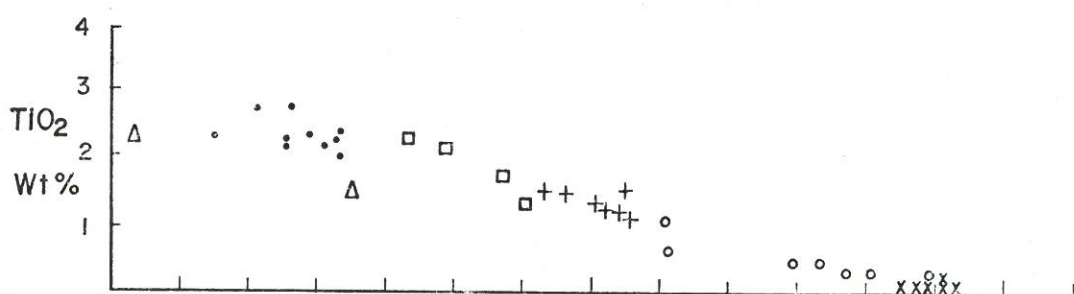
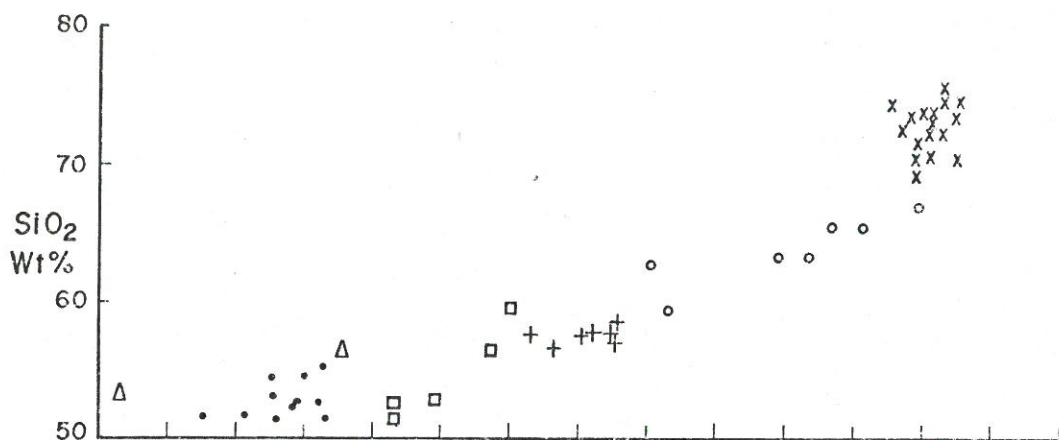
Whole rock chemical data has been plotted on variation diagrams using the Fractionation Index (F.I. = normative Qz + Or + Ab + Ac + Ns (Macdonald, 1969)) as the abscissa (Fig.4.8). This modification of Thornton and Tuttle's (1960) Differentiation Index is necessary to accommodate the late stage peralkaline liquids which Bailey and Schairer (1964) have observed fractionate towards eutectics enriched in alkali silicates rather than towards petrogeny's residua system.

The use of F.I. rather than SiO_2 also effectively separates the hawaiites from mugearites. These two rock types have a similar range of SiO_2 values, but the mugearites have lower MgO and CaO and higher F.I. consistent with fractionation in hawaiites primarily involving Mg and Ca-rich silicates.

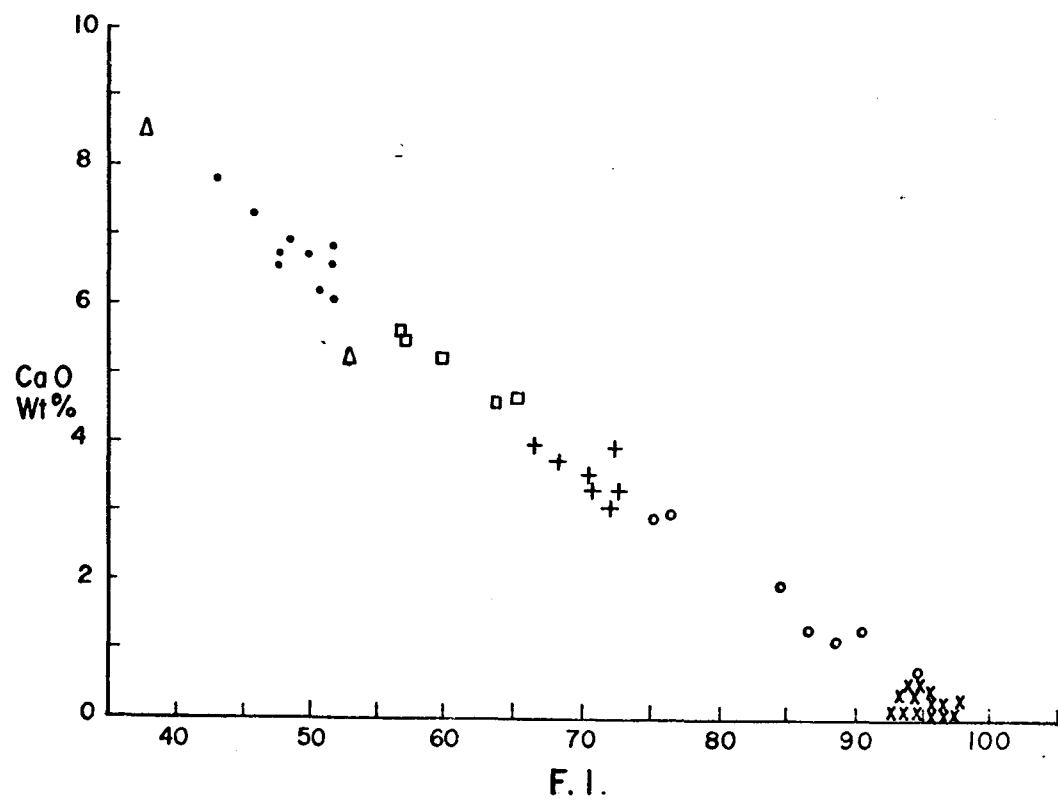
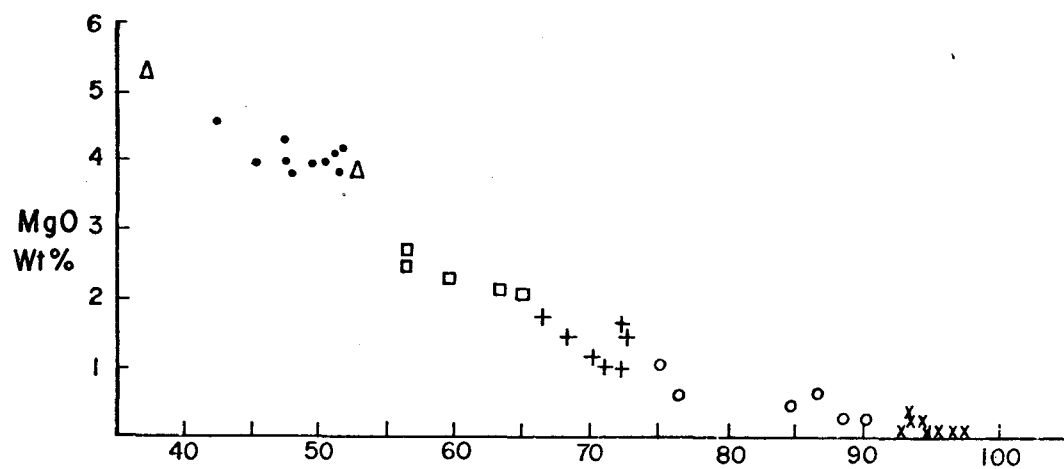
When SiO_2 is used as the abscissa, anorthoclase trachytes and comendites are more effectively separated, as the more differentiated nature of the comendites is camouflaged in the F.I. by the extremely high alkali-feldspar component in the anorthoclase trachytes.

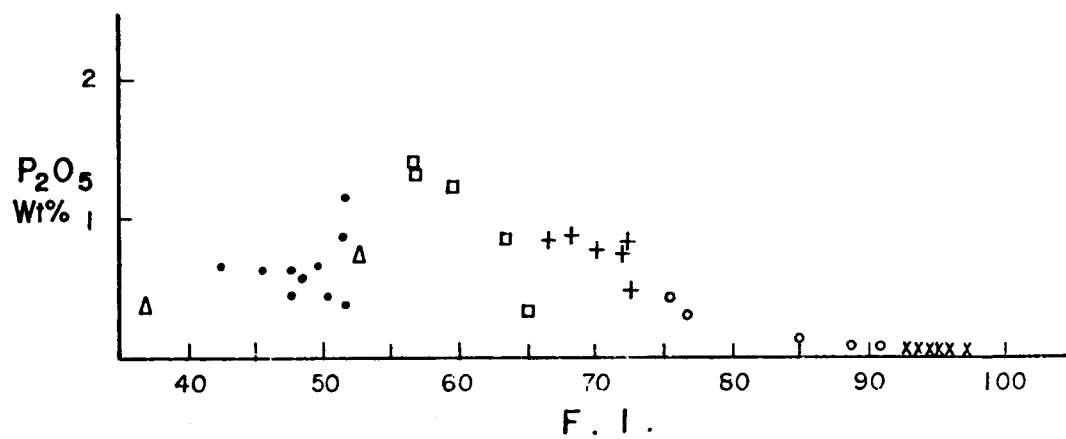
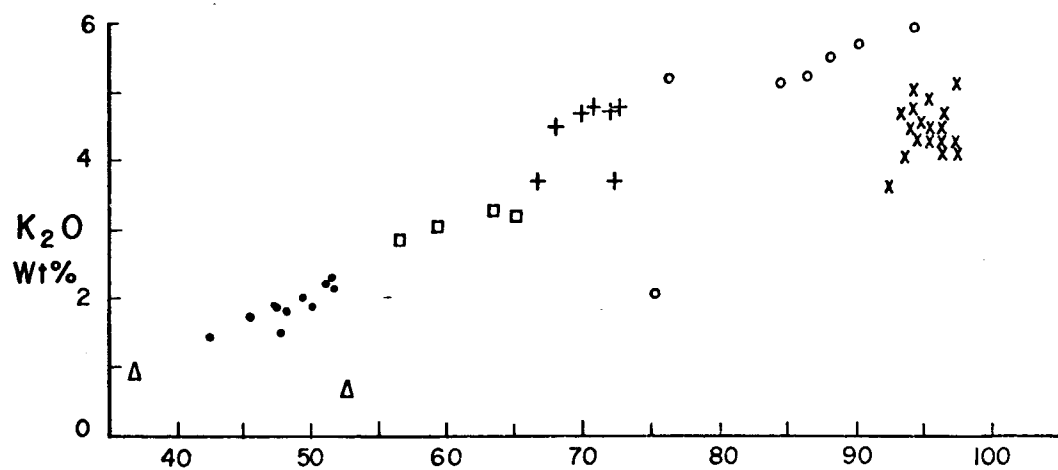
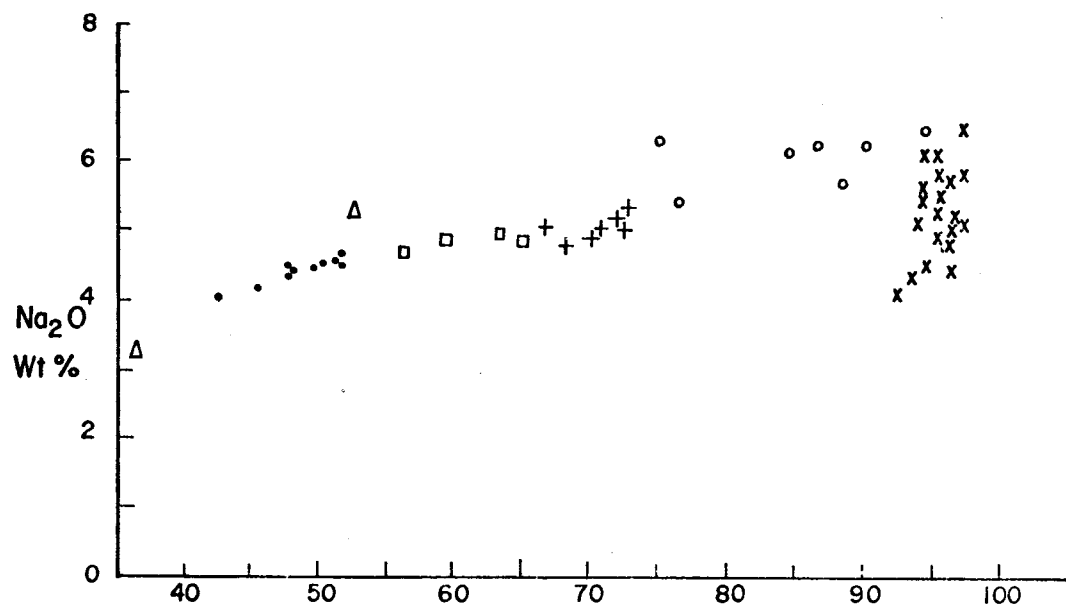
On variation diagrams plots for MgO, CaO, SiO_2 and TiO_2 lie on smooth curves suggesting that the different rock types form a single fractionation series derived from a common parental

Fig. 4.8 Variation diagrams showing Wt % oxides versus F.I. (Fractionation Index = normative Qz + Or + Ab + Ac + Ns) for the Comboyne alkaline and tholeiitic rocks.



- | | |
|------------------------------|--------------------|
| Δ THOLEIITIC ANDESITE | $+$ BENMOREITE |
| \bullet HAWAIIITE | \circ TRACHYTE |
| \square MUGEARITE | \times COMENDITE |





F. I.

liquid. There is a steady decrease with differentiation in MgO, CaO and TiO_2 , whereas SiO_2 increases from 51.43% to 75.65% in the series hawaiite to comendite. More complex curves are exhibited by FeO (total iron as FeO), Al_2O_3 , Na_2O and K_2O , but these can still be related directly to the observed mineralogy and major fractionation phases.

The similarity between trends shown by Al_2O_3 and Na_2O indicates the importance of feldspar fractionation on the host rock composition. However, whereas Na_2O increases from hawaiite to anorthoclase trachyte, Al_2O_3 shows a slight decrease from hawaiite to mugearite and then follows a similar trend to that of Na_2O , both decreasing sharply in comendites. The decrease in Al_2O_3 in mugearites appears to result from the abundant crystallisation of labradorite/andesine megacrysts/phenocrysts in the hawaiites. This suggestion is supported by a marked decrease in CaO from hawaiite to mugearite. CaO continues to decrease throughout the series. The decrease in Al_2O_3 , Na_2O and K_2O in the Comboyne rocks with F.I.'s in excess of 90 is also observed in other provinces which include highly differentiated peralkaline rocks (South Greenland - Macdonald, 1969; Upton et al., 1971; Dogo - Uchimizu, 1966; Gran Canaria - Arana et al., 1973; Aden and Little Aden - Cox et al., 1970; Nandewar - Abbott, 1969). However, this trend is not apparent in the Ethiopian Fantale (Gibson, 1972) and Danakil (Barberi et al., 1970) series which are characterised by lower Al_2O_3 and, to a lesser extent, Na_2O than most other mildly alkaline associations. This possibly results from stronger fractionation of plagioclase in the early stages of the differentiation sequences.

The trend exhibited by K_2O on variation diagrams can also be directly related to feldspar fractionation, as feldspar is the only mineral phase in the Comboyne rocks in which it is accom-

odated in significant amounts. K_2O increases rapidly from hawaiites and mugearites (where labradorite/andesine are the major feldspar megacryst/phenocryst phases), to benmoreites and trachytes where large crystal aggregates of anorthoclase are prominent. The marked drop in K_2O , Na_2O and Al_2O_3 in comendites reflects the strong fractionation of alkali-feldspar in benmoreites and trachytes. This decrease in alkalis and Al_2O_3 in comendites coincides with a sharp increase in SiO_2 and the appearance of modal quartz, indicating that in contrast to the anorthoclase trachytes, which are substantially feldspathic, the comendite liquid has intersected the feldspar-silica phase boundary resulting in the coprecipitation of quartz and feldspar. As a consequence there is a large increase in normative quartz from trachyte (0.0 - 7.54%) to comendite (12.79 - 36.81%). Phenocrysts of quartz are more common in comendite plugs than flows and the near-absence of quartz phenocrysts in flows can be correlated with relatively low SiO_2 values (Table 4.1, analyses MU6506, MU6569, MU6644). The generally higher SiO_2 values in plugs could result from higher pH_2O shifting the quartz-feldspar cotectic towards quartz relative to that in the flows.

The Comboyne comendites are characterised by rather wide ranges in Al_2O_3 , Na_2O , K_2O , SiO_2 and FeO. Some of this variation can be attributed to deuteric leaching as there is a tendency for comendites with low alkalis to have normative corundum, despite modal riebeckite/arfvedsonite and aegirine. These comendites are mainly flows and have relatively low F.I.'s, but not necessarily low SiO_2 contents.

TiO_2 ranges from 2.73% in hawaiites to 0.06% in comendites. There is steady depletion in TiO_2 with differentiation with a marked decrease from benmoreite to trachyte following the abundant

crystallisation of Fe-Ti oxides in the benmoreites. TiO_2 values in the Comboyne rocks are comparable with those for similar rocks from Skye, Nandewar, Aden and Little Aden, Danakil, Fantale, Dogo, Madeira and other 'oceanic' basalts (Chayes, 1965).

P_2O_5 values increase from hawaiite (0.46 - 1.16%) to mugearite (0.31 - 1.31%), and then decreases through benmoreites (0.47 - 0.88%) to comendites (less than 0.12%). Striated megacrysts of apatite are common in hawaiites and benmoreites, but are rare in the aphyric mugearites.

A trachytic rock (MU6665) occurring at the base of South Brother is the only specimen with alkaline affinities found in the eastern part of the Lorne Basin. It plots on the alkaline trend in the FMA diagram and in the alkaline field in a total alkalis versus SiO_2 diagram, but has lower alkalis than benmoreites and trachytes with similar SiO_2 values. It has slightly higher Na_2O and significantly lower K_2O than the Comboyne alkaline rocks with comparable fractionation indices.

Variation diagrams show that plots of MgO , CaO , Al_2O_3 , TiO_2 and P_2O_5 and, to a lesser extent, SiO_2 and FeO for the South Brother trachyte fall on the alkaline trend close to the Comboyne benmoreites and mafic trachytes. When compared with the Lorne Basin calc-alkaline rocks on variation diagrams using SiO_2 as the abscissa, the South Brother trachyte shows little relationship to the general trends for FeO , Na_2O , P_2O_5 , Al_2O_3 and CaO , which supports the conclusion that this rock is unrelated to the Lorne Basin calc-alkaline rocks (Figs. 4.8 and 4.10).

4.2.2 Tholeiitic Rocks

4.2.2.1 Classification and Discussion

Tholeiitic rocks have only been positively identified from one locality on the Comboyne Plateau. Tholeiitic andesite falls slightly to the Mg-side of the alkaline trend in close proximity to the hawaiites when plotted on an FMA diagram (Fig. 4.5). Also it falls in the sub-alkaline field on a $\text{Na}_2\text{O} + \text{K}_2\text{O}$ versus SiO_2 diagram (Fig. 4.3b). The tholeiitic andesite (MU6577) and the more strongly altered (?) tholeiitic rock (MU6568) have been plotted on both alkaline and sub-alkaline variation diagrams so that direct comparison with rocks with similar fractionation indices and SiO_2 values can be made (Figs. 4.8 and 4.10).

Major chemical differences between the tholeiitic and alkaline rocks are apparent, although in plots of MgO , CaO , Na_2O and K_2O versus F.I., the tholeiitic andesite could be considered on a continuation of the alkaline trend. However, it has significantly lower Al_2O_3 , TiO_2 and total alkalis, and higher FeO , MgO and CaO than the associated hawaiites with comparable SiO_2 values. These differences are not emphasised when oxides are plotted against F.I. as lower Al_2O_3 and alkalis in the tholeiitic andesite results in a substantially lower Fractionation Index than those in alkaline rocks with similar SiO_2 values.

The tholeiitic andesite falls at the basic end of the sub-alkaline variation curves where a scatter of points obscures any trend, but the tholeiitic rock has substantially lower Al_2O_3 than calc-alkaline rocks with comparable SiO_2 values possibly indicating its tholeiitic affinities.

Tholeiitic rocks closely associated with mildly alkaline provinces have been reported from Ethiopia (LeBas and Mohr, 1970), Aden and Little Aden (Cox et al., 1970), Uganda (Davies, 1953),

the Tweed Shield, N.S.W. (Duggan and Wilkinson, 1973), and the well documented Hawaiian (Macdonald and Katsura, 1964) and Hebridean (Holland and Brown, 1972) provinces.

Chemically the Comboyne tholeiitic andesite is comparable with tholeiitic rocks occurring at the Simien alkali basalt centre, Ethiopia, the Tweed Shield, N.S.W., and the Hebridean tholeiitic cone sheets, although it has higher MgO and CaO and lower Na₂O and K₂O than the Tweed and Hebridean tholeiites.

4.2.3 Calc-Alkaline Rocks

4.2.3.1 Classification and Comparison with other Associations

A wide variation in chemical composition is also shown by the Lorne Basin sub-alkaline rocks and SiO₂ values range from 50.64 - 77.23%.

The sub-alkaline nature of these Lorne Basin rocks is shown when total alkalis are plotted against SiO₂ when, with the exception of some xenoliths in the North Brother porphyritic microgranite, the majority of rocks fall in Kuno's (1968) hypersthenic field. The high alkali content of the rocks with SiO₂ in excess of 70% is shown in Figures 4.3 a and b, where there is a merging of alkaline and sub-alkaline rock types.

The classification of the Lorne Basin sub-alkaline rocks has been based on chemical composition rather than petrographic criteria. The observed mineralogy is consistent with the chemically derived nomenclature, although the primary mafic mineral in most of the granitic rocks is pyroxene rather than hornblende, reflecting the high-level emplacement of these rocks. Rare (?) riebeckite in the South Brother microgranite is considered to result from extreme differentiation causing depletion in Al₂O₃ and enrichment in alkalis and volatiles in the late stage fluids rather than being

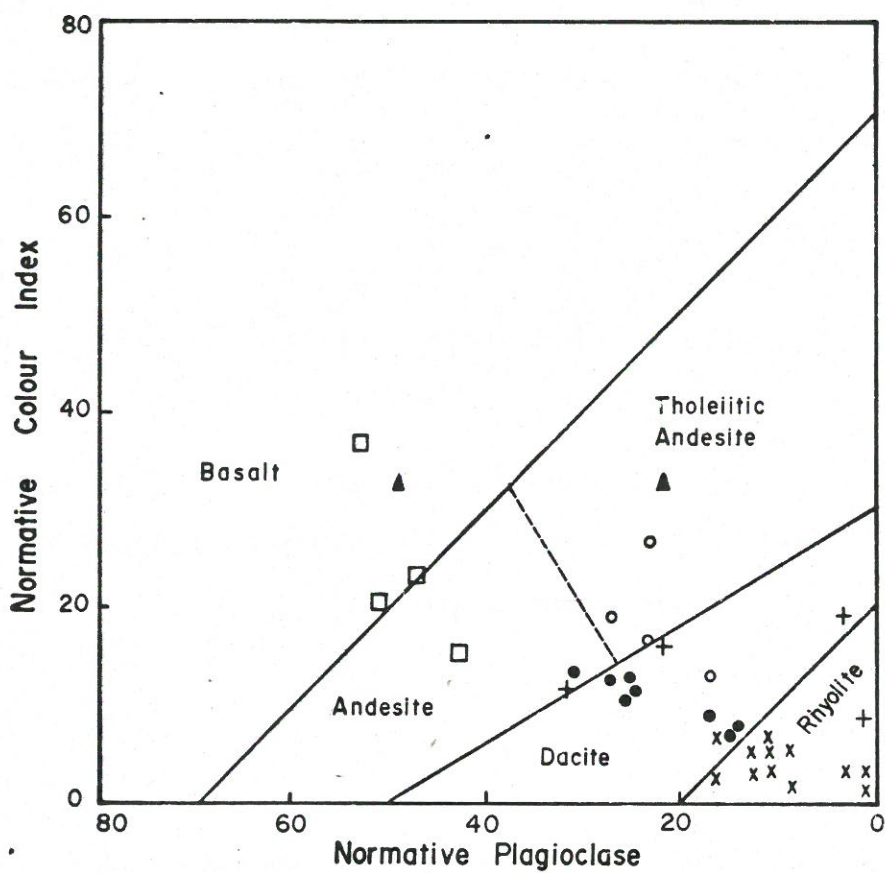
indicative of an alkaline lineage for this rock.

The classification of diorites and granodiorites follows the schemes proposed by Irvine and Baragar (1971), Gulson (1972) and Gulson et al. (1972).

In an extensive study on volcanic rocks, Irvine and Baragar have outlined a classification system based on chemical criteria, so that these rocks which have previously been classified on a mineralogical basis, can be more precisely defined. They have delineated calc-alkaline and tholeiitic fields on 1) FMA diagrams, and 2) plots of normative colour index ($Ol + Opx + Cpx + Mt + Il + Hm$) versus normative plagioclase composition (Fig. 4.9). The Lorne Basin sub-alkaline rocks plot in the calc-alkaline fields in both of these diagrams and contrast with the Comboyne tholeiitic andesite which plots in the tholeiitic field. Jakes and White (1972) note that calc-alkaline rocks typically have high Al_2O_3 (15-18%), moderate to high total alkalis, $K_2O:Na_2O$ ratios between 0.35 and 0.75, and almost constant $FeO:MgO$ ratios, and the concurrence of the Lorne Basin rocks with these limits supports the conclusion that they are calc-alkaline rather than tholeiitic (Table 4.2).

Gulson (1972) and Gulson et al. (1972) suggest a classification of diorites which follows a similar scheme to that for andesites. Under this proposal diorites are described as Low-Si, Low-K, Normal, High-K and Granodiorite. Normal diorites and granodiorites have, respectively, 56-62% SiO_2 , 0.7-2.5% K_2O , and 62-68% SiO_2 , and greater than 2.5% K_2O ; values similar to those in the Lorne Basin diorites and granodiorites. Although several of the granodiorites have slightly less than 2.5% K_2O , this could simply result from deuteric alteration. SiO_2 values greater than 70% and colour indices less than 20 (Irvine and Baragar, 1971) are used to define

Fig. 4.9 Sub-alkaline rocks plotted on Normative Colour Index versus Normative Plagioclase composition diagram, showing fields designated by Irvine and Baragar (1970) for Basalt, Tholeiitic Andesite, Andesite, Dacite and Rhyolite.



granites and rhyolites.

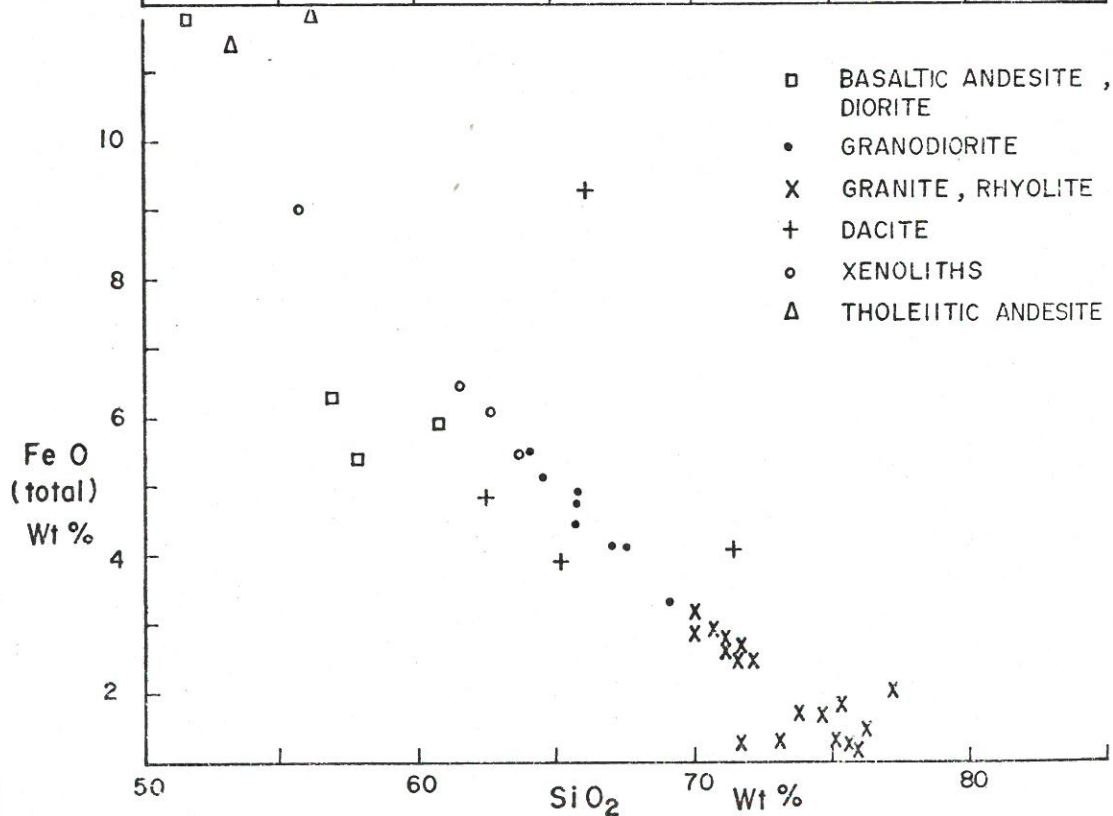
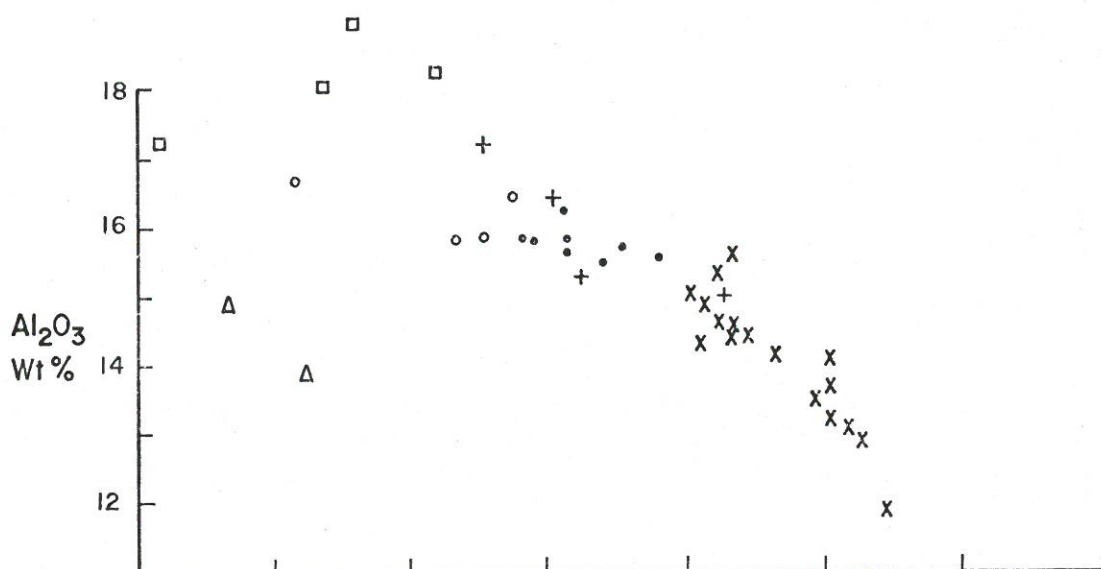
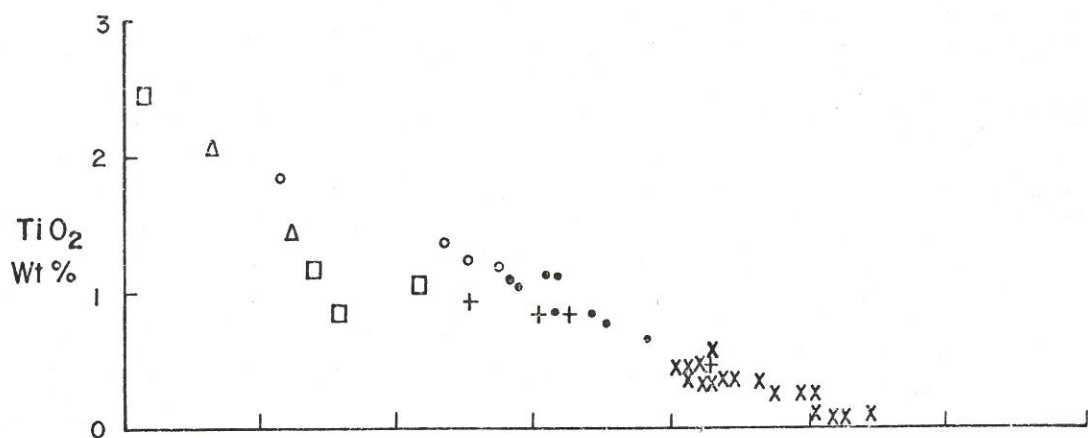
The Lorne Basin granitic rocks tend to have higher Na_2O and lower K_2O than the plutonic New England, N.S.W. (Shaw, 1964, Flood, 1971) and Cape, South Africa (Kolbe, 1966) calc-alkaline associations and average analyses for similar rock types given by Nockolds (1954). The high level Central North Island, N.Z. volcanics (Ewart and Stipp, 1968), the Westerly and Narragansett Pier granites, New England, U.S.A. (Buma et al., 1971), and the Georgetown Inlier, Queensland, volcanics, ring complexes and granites (Branch, 1966), are more directly comparable. The major element composition of the Lorne Basin pitchstones is similar to that of the Talasea, New Britain, rhyolitic obsidian (Lowder and Carmichael (1970), but slightly higher Al_2O_3 and lower FeO reflect the more strongly calc-alkaline character of the Lorne Basin rocks.

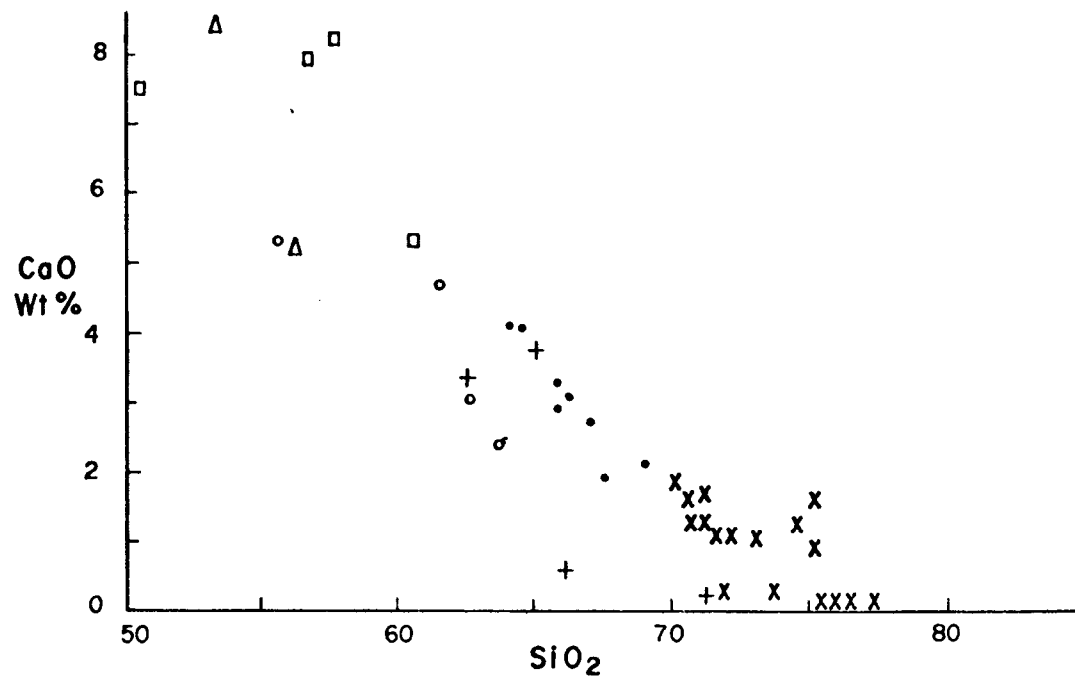
4.2.3.2 Variation Diagrams and Discussion

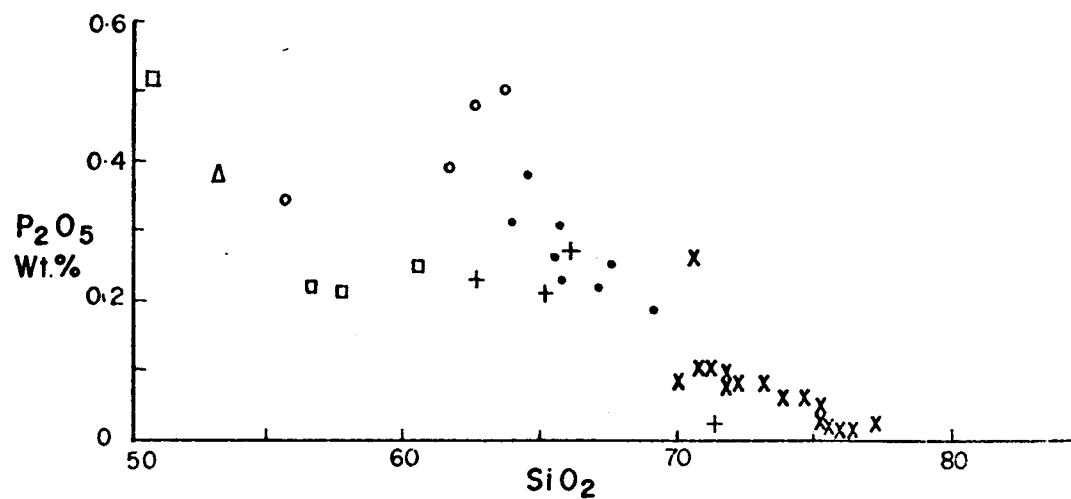
The major sub-alkaline rocks found in the vicinity of the Lorne Basin include granodiorite, granite and acid volcanics, and there are minor occurrences of diorite and basaltic andesite. Because of the wide variation in SiO_2 content in these rocks, SiO_2 has been used as the abscissa in variation diagrams (Fig. 4.10).

The basaltic andesite (MU6689), which forms dykes in the intertidal zone at Diamond Head, has extremely large phenocrysts of plagioclase, and both phenocrysts and matrix have undergone marked alteration to chlorite, white mica and clay minerals. Analysis of the groundmass (Table 4.2) shows this rock to have high FeO and low MgO and CaO. It plots at the basic end of variation diagrams on what could be considered a continuation of the trend shown by the granitic rocks, but the strong alteration throughout the basaltic andesite makes this conclusion tenuous.

Fig. 4.10 Plots of Wt % oxides versus SiO_2 for the Lorne
Basin - Comboyne Plateau sub-alkaline rocks.







The small stocks of diorite occurring to the north of the Lorne Basin are similar to the Diamond Head quartz diorite apart from lower SiO_2 and higher MgO and CaO .

The massive fine grained rocks associated with the Diamond Head rhyolite are dacitic in composition although they have anomalously low CaO . Two specimens have been analysed (MU6686, MU6687) and textural evidence indicates that the rather variable results result from differing degrees of alteration. Very low CaO values, the presence of clear groundmass albite, and the abundance of fine white mica replacing turbid multiple-twinned phenocrysts of plagioclase, indicates that leaching has resulted in the removal of CaO and the remobilisation and/or introduction of Na_2O and K_2O . High normative corundum (3.55 - 6.27%) result primarily from the removal of CaO rather than alkalis. These rocks have abundant cubes of pyrite.

Middle Brother consists primarily of granodiorite and granite, with diorite and granophyre occurring as minor phases. Similar granitic rocks form Juhle Mountain and other low hills to the north of the Lorne Basin. These rocks show limited variation in chemical composition with SiO_2 ranging from 70.04 - 75.10% in granites and 64.25 - 69.08% in granodiorites. MgO , CaO and Al_2O_3 are higher in the granodiorites and K_2O substantially lower.

The North Brother porphyritic microgranite has slightly higher CaO , FeO and Na_2O and lower SiO_2 and K_2O than the Middle Brother granites. A texturally similar microgranite at Ross Glen, just north of Middle Brother, is compositionally closer to the Middle Brother granites since it has 73.96% SiO_2 .

Porphyritic pitchstone dykes near Kew and Moorland have higher SiO_2 values than the North Brother microgranite, but lower Al_2O_3 ;

Na_2O , K_2O and FeO , reflecting the silica-rich character of the glass and the relative scarcity of feldspar and ferromagnesian minerals.

The South Brother microgranite is the most evolved rock found in the Lorne Basin area, but strong deuteric alteration makes analyses of this rock of dubious value. Significantly lower FeO and generally higher Al_2O_3 distinguish the South Brother microgranite from the Comboyne comendites. Higher K_2O than Na_2O also contrasts with the comendites which generally have Na_2O in excess of K_2O . However, a higher degree of deuteric alteration in the microgranite could have resulted in a greater removal of Na_2O relative to K_2O than in the comendites.

Plots of MgO , TiO_2 , FeO and, to a lesser extent, K_2O against SiO_2 for the majority of the Lorne Basin rocks describe a smooth curve consistent with their being representative of a single fractionation series. In contrast plots of CaO , Na_2O and P_2O_5 result in a greater scatter of points and this could indicate that many of the rocks are of independent origin, or alternatively, that differing degrees of secondary alteration have caused this scatter. The degree of secondary alteration in these rocks has already been noted (Chapter 2), and it is significant that the variation diagrams displaying the least coherence of plots represent oxides acknowledged as being the most soluble and mobile during deuteric alteration (Lipman, 1965). It is concluded that deuteric alteration has resulted in varying degrees of leaching and redistribution of the more mobile elements and caused some scattering of points on variation diagrams. However, whereas most of the rocks fall on smooth curves when Al_2O_3 , MgO , CaO and TiO_2 are plotted against SiO_2 , the Diamond Head and Lorne Forest diorites are anomalous. This is confirmed when TiO_2 is plotted

against MgO (both of which are considered to resist leaching) and strongly suggests that these diorites do not belong to the same fractionation series as the other calc-alkaline rocks in the area (Fig. 4.11).

Two samples (MU6580, MU6581) from sills low down on the southern side of the Comboyne Plateau and intruding the Lower Permian–Upper Carboniferous Giro Beds, are also apparently unrelated to the major Lorne Basin calc-alkaline series. The calc-alkaline rather than tholeiitic affinities of these rocks is suggested by relatively high Al_2O_3 values (16.49 – 17.20%) and moderately high total alkalis (6.97 – 6.24%). These rocks are strongly altered and all ferromagnesian minerals have been replaced by chlorite, (?)iddingsite, calcite and white mica. Normative corundum indicates leaching of alkalis and CaO has also occurred. These dacitic rocks plot in the sub-alkaline field of $Na_2O + K_2O$ versus SiO_2 diagrams. Low FeO relative to MgO results in plots on a FMA diagram being unrelated to the trend shown by the Lorne Basin calc-alkaline rocks (Fig. 4.5).

4.3 Trace Element Analyses

4.3.1 Alkaline Rocks

In order that trace element data for the alkaline, tholeiitic and calc-alkaline rocks can be directly compared, selected trace element abundances of all rock types are plotted against SiO_2 in Figure 4.12.

Rubidium: Rb values for the alkaline series, hawaiite through to anorthoclase trachyte show a relatively small variation from 24 – 135 ppm. In contrast the comendite Rb values range from 73 – 904 ppm, averaging 323 ppm (Fig. 4.12). These values are comparable with those determined from similar rocks at Nandewar

Fig. 4.11 Plots of TiO_2 versus MgO for Lorne Basin
calc-alkaline rocks showing lack of continuity
between the silicic granitic rocks and the more
basic diorities. Symbols as for Figure 4.10.

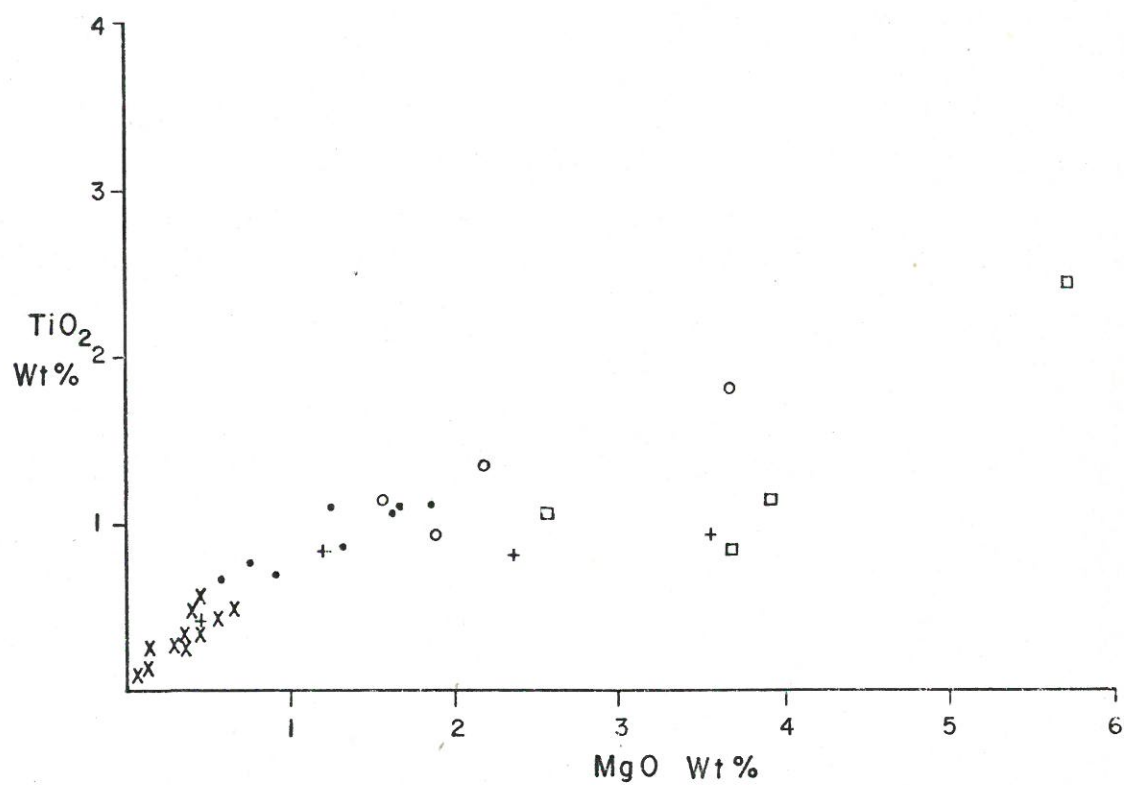
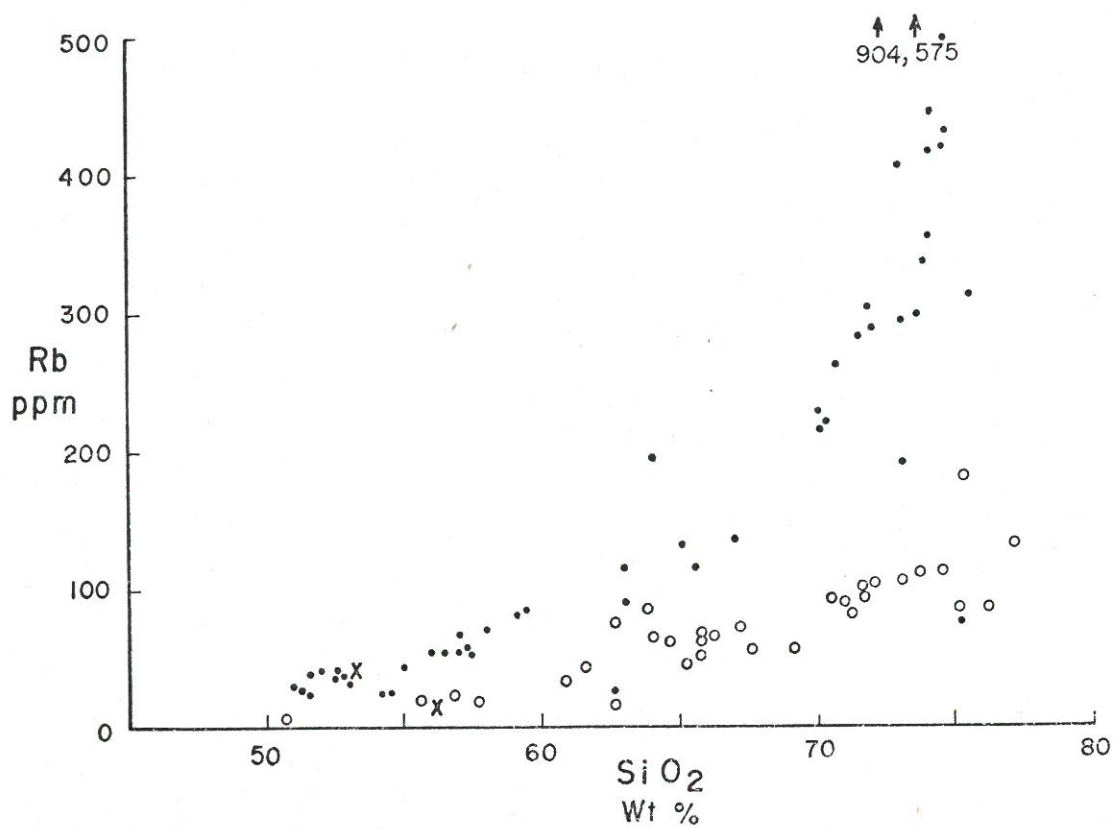
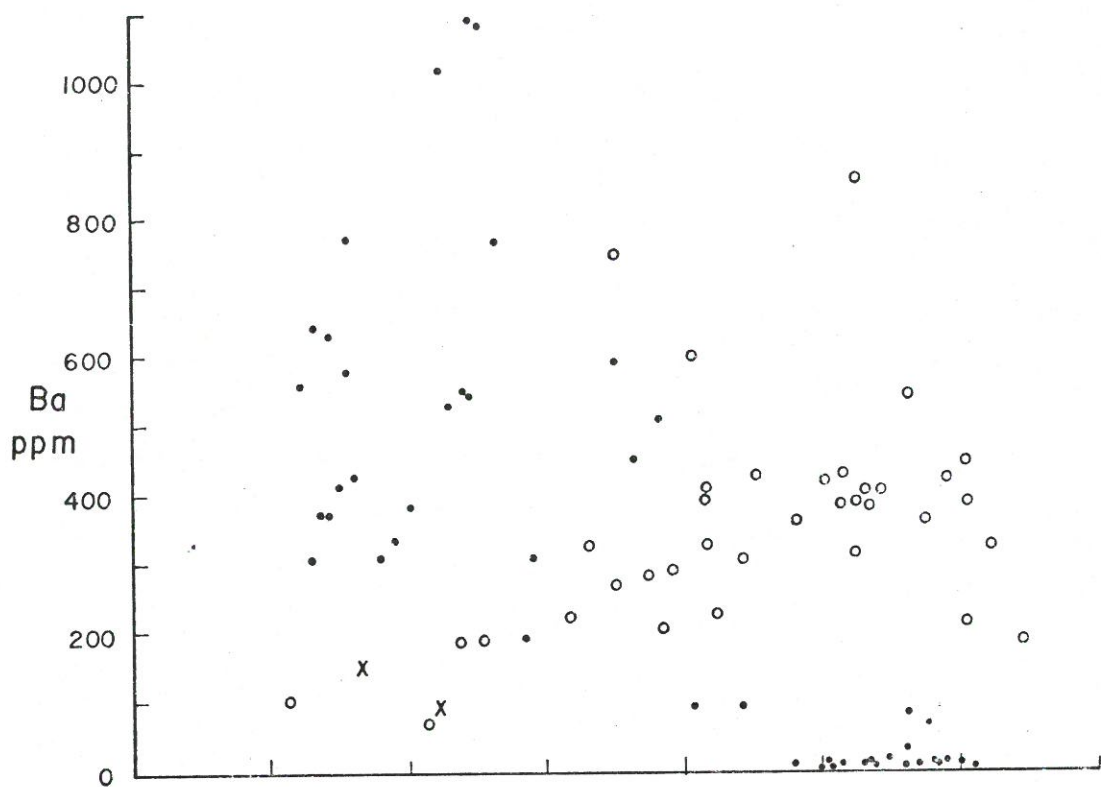
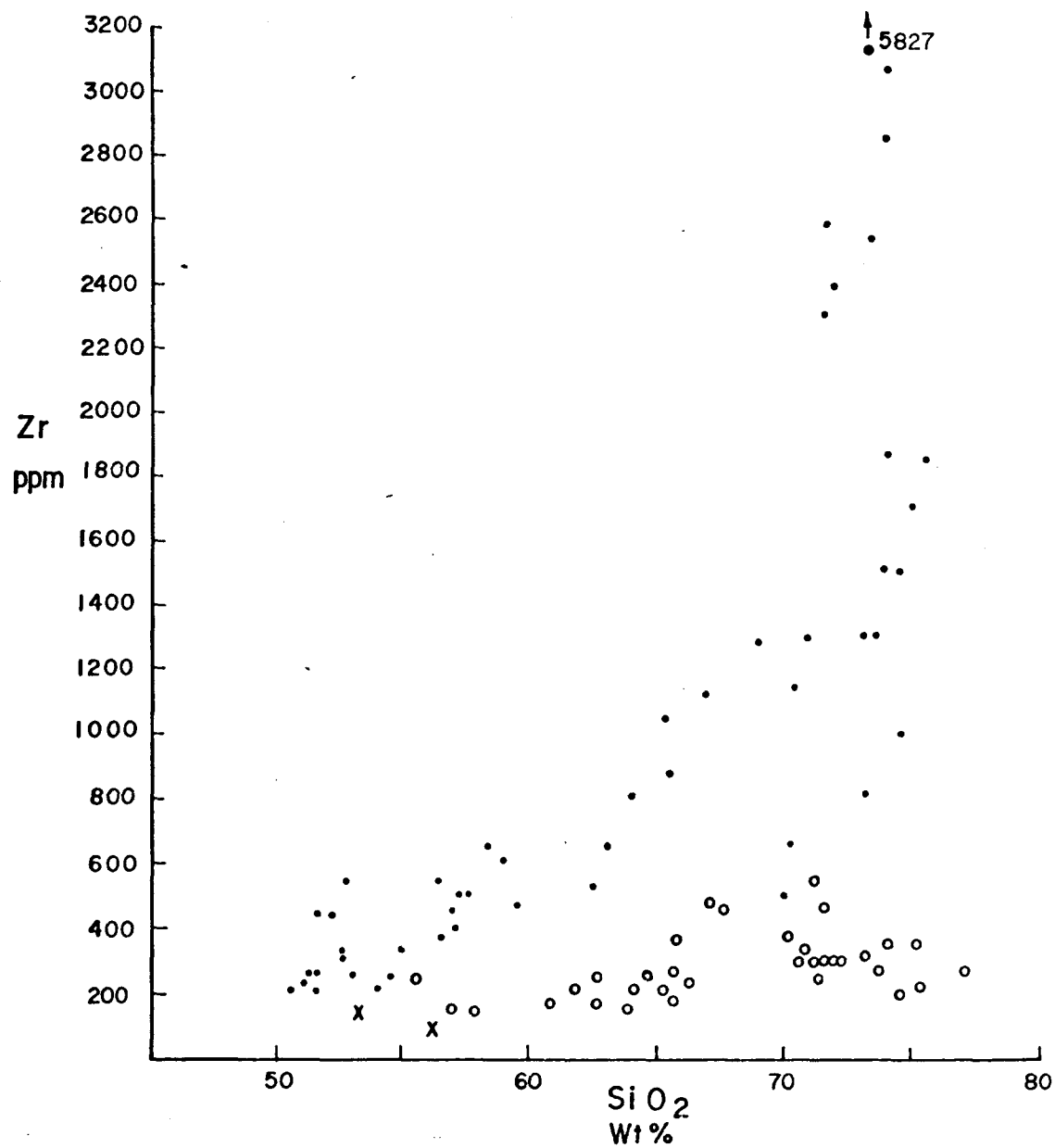
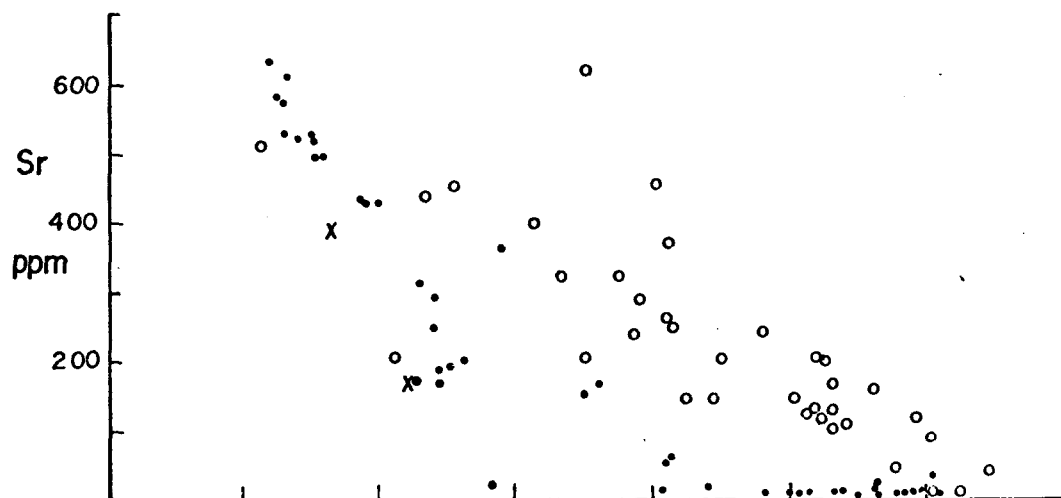
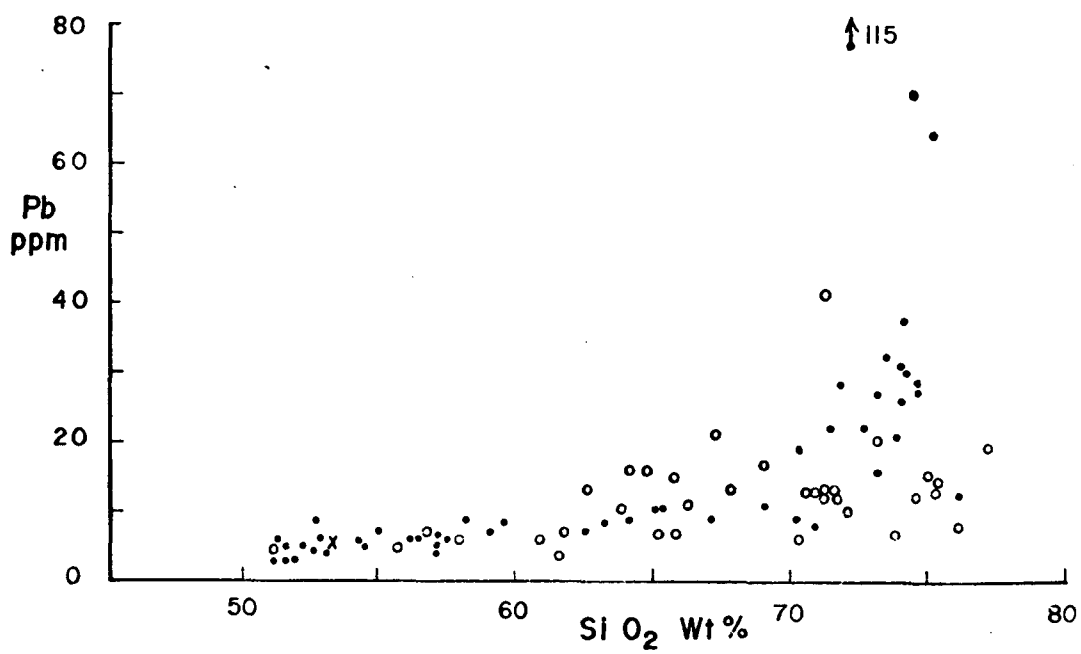
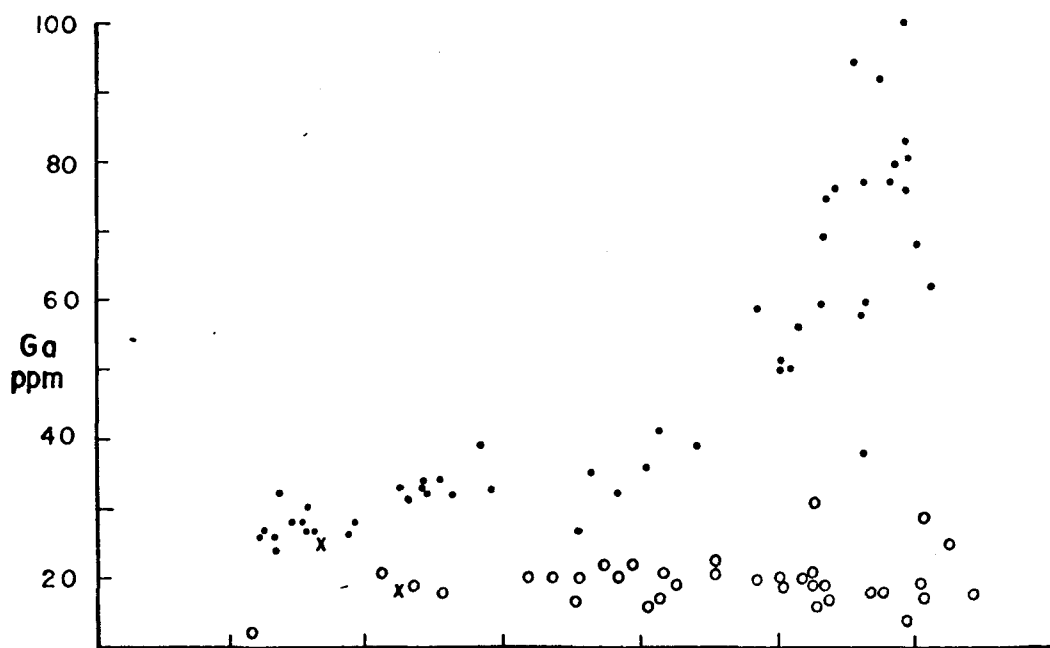
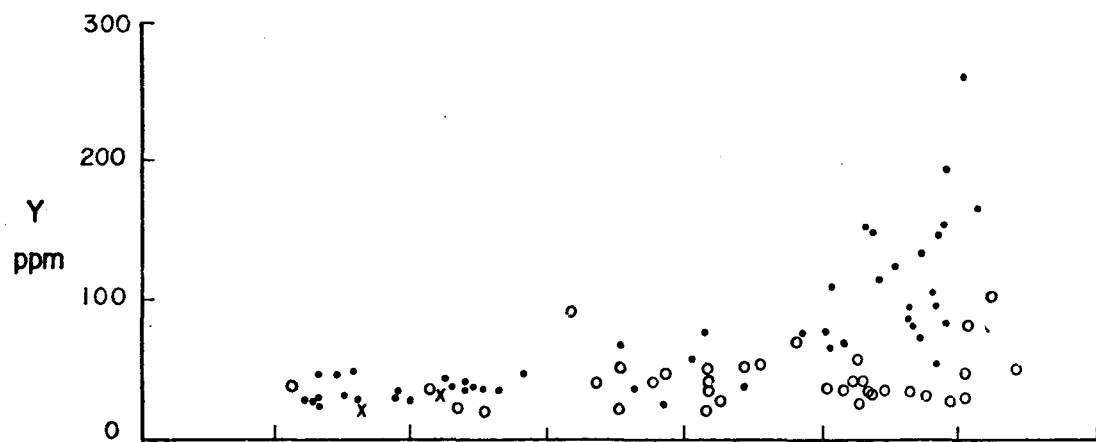


Fig. 4.12 Plots of Ba, Rb, Sr, Zr, Y, Ga and Pb (ppm) versus SiO_2 (Wt %) for the Comboyne Plateau - Lorne Basin, alkaline and sub-alkaline rocks. Solid circles = alkaline rocks. Crosses = tholeiitic rocks. Open circles = calc-alkaline rocks.







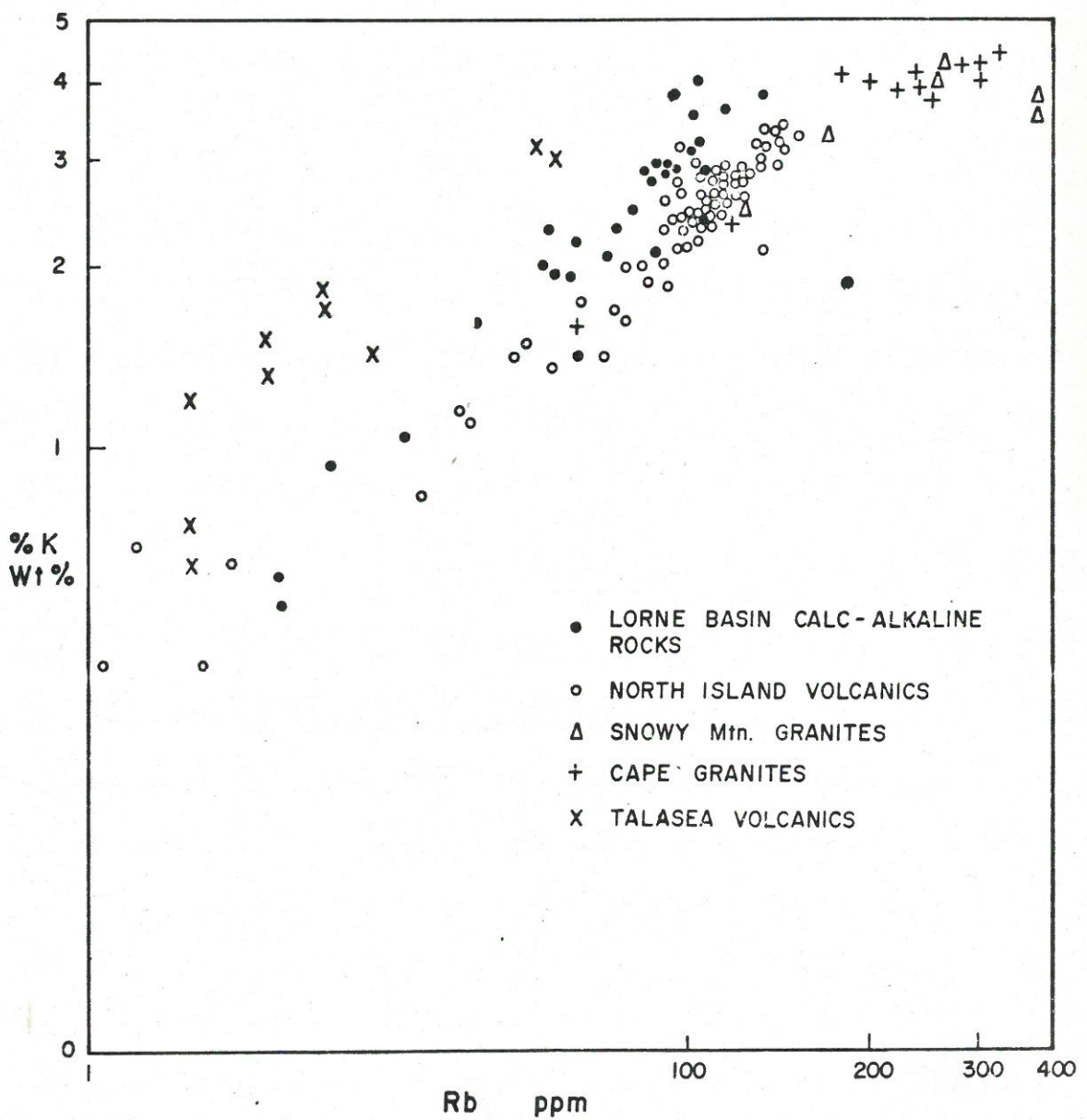
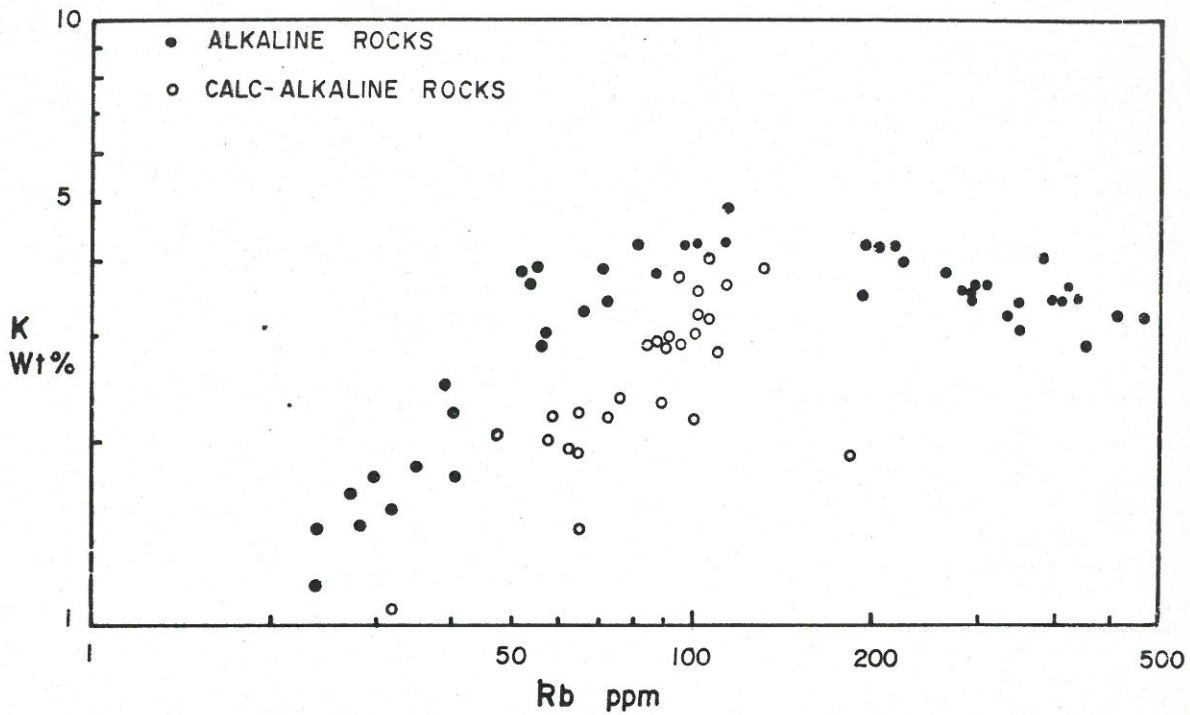
(Abbott, 1967), Easter Island (Baker et al., 1974), Madeira (Hughes and Brown, 1972), Canary Islands (Arana et al., 1973), Marquesas Islands (Bishop and Woolley, 1973), although Rb tends to be higher and more variable in the Comboyne and Marquesas peralkaline rocks. The aegirine-bearing comendites tend to have higher Rb values than the aegirine-free types, and although Prinz (1967) noted that Rb is only present in detectable amounts in feldspar and feldspathoids, it has also been identified in late-forming pyroxenes in the Skaergaard intrusion (Wager and Mitchell, 1951). The Comboyne comendites exhibit a greater decrease in K when compared with similar series and generally the plugs have higher Rb and lower K/Rb ratios than flows (Table 4.1).

K/Rb ratios for the Comboyne hawaiites range from 419-609 in rocks in which K varies from 1.17% to 1.84%. These values are comparable with basic alkaline rocks at Nandewar (Abbott, 1967b) and Hawaii (Hamilton, 1965), and contrast with lower ratios for alkaline basalts from Madeira (Hughes and Brown, 1972), Gough Island (Le Maitre, 1962), Ethiopia (Treuiel et al., 1971), and Antarctica and New Zealand (Gunn, 1965). Shaw (1968) has suggested that with the exception of pegmatites and oceanic tholeiites, igneous rock suites have a linear relationship between K and Rb with a slight decrease in K/Rb ratio with increasing K. However, as has been found in other associations which include highly evolved alkali rhyolites (Abbott, 1967b; Macdonald and Edge, 1970), Shaw's proposed relationship does not hold and there is a marked decrease in K/Rb ratio in the Comboyne comendites resulting from both a decrease in K and an increase in Rb (Fig. 4.13a).

Strontium: Sr values in the Comboyne alkaline series range from 633 ppm in hawaiites to 5 ppm in comendites. Values for the hawaiites are higher than the Easter Island hawaiites, similar to

Fig. 4.13a Plots of Wt % K versus Rb (ppm) for the
Comboyne Plateau - Lorne Basin alkaline and
sub-alkaline rocks.

Fig. 4.13b. Plots of Wt % K versus Rb (ppm) for the
Lorne Basin calc-alkaline rocks with similar
plots for other sub-alkaline provinces.
North Island Volcanics (Ewart and Stipp,
1968), Snowy Mountain and Cape Granites
(Kolbe and Taylor, 1966b), Talasea
Volcanics (Lowder and Carmichael, 1970).



those for the Hawaiian and Madieran hawaiites, and lower than those reported for the Marquesas Island and, to a lesser degree, the Nandewar hawaiites.

The extreme depletion in Sr in an aphyric trachyte (MU6501) (21 ppm) relative to porphyritic benmoreites and anorthoclase trachytes (178-297 ppm and 54-198 ppm respectively) suggests that Sr is concentrated in feldspar phenocrysts rather than the ground-mass. This trend is also reported by Seal and Weaver (1971), Noble et al. (1969), and Dickinson and Gibson (1972). It is also noted that the feldspar-phyric comendites have substantially higher Sr values (23-37 ppm) than the aphyric comendites (5-9 ppm). Noble et al. (1969) pointed out that peralkaline silicic glasses exhibit strong depletion in Mg and Sr, with Ca being relatively less depleted, and the Comboyne comendites show similar trends. This concurs with the observation by Berlin and Henderson (1967) that Sr is enriched in plagioclase relative to Ca.

Depletion in Sr is particularly evident after anorthoclase becomes the major phenocryst phase, which supports the findings of Heier (1962) and Butler and Skiba (1962) that K-feldspar and intermediate plagioclase are comparatively enriched in Sr relative to highly calcic and sodic plagioclase.

In a survey of basaltic rocks, Prinz (1967) found Rb/Sr ratios of 0.06 for tholeiites, 0.05 for alkali basalts and 0.06 for all rocks of basaltic composition. Rb/Sr ratios for the Comboyne hawaiites vary from 0.04 to 0.10 and show a negative correlation with increase in Sr. In contrast, Rb/Sr ratios for comendites range up to 95.83, resulting from marked enrichment in Rb and depletion in Sr.

Barium: There is an increase in Ba values from 304-776 ppm in hawaiites and mugearites to 544-1096 ppm in benmorites. This is

followed by a rapid decrease from trachyte (189 ppm in mafic trachyte, 509-97 ppm in anorthoclase trachytes) to comendite (83-2 ppm). A strong positive correlation is noted between higher Ba content and the more porphyritic rock types, as is reflected in the anorthoclase trachytes where aphyric specimens (MU6520, MU6635) have 97 ppm Ba, whereas porphyritic types have Ba values in the vicinity of 500 ppm. These trends result from the incorporation of Ba in early formed K-rich minerals (Taylor, 1965), and depletion in Ba follows strong fractionation of K-feldspar. Similar values and trends are shown by the alkali basalt-trachyte Paka series, Kenya (Sceal and Weaver, 1971), the Mayor Island, N.Z. trachybasalt and pantellerites (Ewart et al., 1968), Jebel al Abyad (Baker et al., 1973), Nandewar (Abbott, 1969), Oki-Dogo Island (Nagasawa, 1973), Gough Island (Le Maitre, 1962) and Tristan da Cunha (Baker et al., 1964). In contrast, the Marquesas and Madiera alkaline rocks are enriched in Ba, whereas in the Easter Island rocks there is enrichment in Ba in the more evolved rocks relative to the basic types.

Berlin and Henderson (1969) found that when plagioclase is the first feldspar to crystallise the Ba/Sr ratio increases as the result of the preferential inclusion of Sr relative to Ba. The early and abundant crystallisation of plagioclase in the more basic rocks is reflected by an increase in the Ba/Sr ratio up to and including trachytes. Large-scale crystallisation of anorthoclase in benmoreites and trachytes is followed by a decrease in Ba/Sr ratios in comendites.

Uranium and Thorium: U in the alkaline basic and intermediate rocks is low and has a narrow range from below the limit of detection to 5 ppm. Peralkaline rhyolite U values vary from 5 ppm to 29 ppm, and show a tendency to be higher in plugs than in flows. These comendites are considerably enriched in U and Th when

compared with the Mayor Island pantellerites.

Thorium distribution in the alkaline series shows similar trends to U, varying from 2 to 18 ppm in hawaiites through to anorthoclase trachytes, with comendites exhibiting a significantly greater variation from 22-108 ppm. Th/U ratios are high, ranging up to 8.33 in samples where U is above the limit of detection, and the ratio varies considerably in any one rock type. This suggests U or Th may be concentrated in a minor, irregularly distributed mineral phase, or that there were varying degrees of removal of U and/or Th during deuteric alteration. A tendency towards increased Th/U ratios with differentiation indicates either concentration of Th has occurred, or there has been a greater loss of U relative to Th during late stage secondary processes.

Lead: Pb is considered to be concentrated in residual melts and in the Comboyne alkaline series ranges from 3 ppm in hawaiite to 115 ppm in comendites. There is a tendency for comendite plugs to have higher Pb values than flows. This could result from the porphyritic nature of these plugs with Pb having been preferentially incorporated into feldspar phenocrysts. Kolbe (1966) has shown K-feldspar to be the main Pb-bearing mineral, so it may be anticipated that Pb would be incorporated into early forming K-feldspar. Pb values in the Marquesas Island series are higher than in the Comboyne series and both show little variation in Pb content in the basic and intermediate rock types. It is only in the Comboyne comendites that there is a marked increase in Pb content.

Gallium: The Comboyne alkaline series has consistently higher Ga (24-100 ppm) than the associated calc-alkaline rocks (13-25 ppm) and the two associations can be effectively separated using Ga as

the criterion (Fig. 4.12). This trend has also been reported by Goldschmidt (1954) and Noble and Häfhty (1969) and is particularly evident when calc-alkaline and alkaline rocks with similar SiO_2 values are compared.

In the alkaline series Ga shows a steady increase with degree of differentiation, and there is little variation within rock types, with the exception of comendites. In these Ga varies from 50-100 ppm. Macdonald and Edge (1970) attribute considerable scatter of Ga in the Tugtutoq microgranites to loss of Ga in late stage Na-rich fluids. There is no evidence that this has been a significant process in the Comboyne comendites.

There is also a tendency for Ga to increase with Zr. This is anomalous as Zr has been shown to concentrate to some degree in pyroxene (Chao and Fleischer, 1960; Taylor, 1965; Wilkinson, 1969), whereas Goodman (1972) has noted that Ga is strongly depleted in pyroxene phenocrysts relative to plagioclase phenocrysts and groundmass. Goodman's work, however, is only related to basic rocks, and he did note that Ga showed a strong coherence with Fe^{3+} . Therefore it is possible that NaFe^{3+} -rich pyroxenes will be enriched in Ga relative to NaFe^{3+} -poor pyroxenes.

Yttrium: Y in the Comboyne alkaline series shows a steady increase with degree of differentiation, varying from 24 ppm in hawaiite to 262 ppm in comendite. Y values given for Easter Island, Madiera and the Marquesas Island rocks are similar to those determined for the Comboyne rocks, although there is a tendency for the Comboyne values to be slightly lower. A comparison of the alkaline and calc-alkaline rocks (Fig. 4.12) shows that there is greater enrichment of Y in the more evolved alkaline rocks than in the calc-alkaline rocks. The concentration of Y in residual liquids is clearly demonstrated in the Comboyne comendites with values varying

from 53-262 ppm. This compares closely with values for the over-saturated syenites and alkali granites of the Central Complex, Tugtutoq (60-290 ppm) (Upton et al., 1971). The Comboyne comendite plugs tend to have higher Y than the aphyric flows

Zirconium: Zr values range from 204 ppm in hawaiite to 5827 ppm in comendite. It tends to be high relative to similar rock types from other areas (Abbott, 1969); Weaver et al., 1972; Ewart et al., 1968; Gibson, 1972). Hughes and Brown (1972), Baker et al. (1974) and Cox et al. (1970) quote Zr values for hawaiites, mugearites, benmoreites and trachytes from Madeira, Easter Island and Aden and Little Aden comparable with those for similar rocks at Comboyne. However, peralkaline rhyolites from Easter Island and Aden are considerably less enriched in Zr (up to 1500 ppm compared with 5627 ppm for the Comboyne comendites). The Comboyne values are similar to those reported by Macdonald and Parker (1970) for the South Greenland intrusive alkaline rocks which are considered to be among the most Zr-rich in the world.

Chao and Fleischer (1960) have noted wide regional variation for Zr and record basaltic averages of 10-60 ppm for island arc rocks from Japan, Palau and the Aleutian Islands, and 120-520 ppm for oceanic basaltic rocks from Polynesia. It is apparent that Zr values in alkaline rocks are generally high throughout any one series and that high Zr is a function of initial composition of the parental melt as well as degree of evolution.

Zr is considered to be highly soluble in alkali melts (Dietrich, 1968; Nicholls and Carmichael, 1969), but the relatively high Zr in the Comboyne hawaiites, mugearites and benmoreites suggests that Zr is incorporated in a crystal phase unless there is glass at grain boundaries. Interstitial glass

has only been observed in the porphyritic mugearite which has 464 ppm Zr which is similar to values in glass-free mugearites and benmoreites. Feldspar, olivine, ilmenite and magnetite are essentially Zr-free, however, apatite, zircon and alkali-amphibole and pyroxene can contain appreciable Zr, and it is probable that apatite, a common accessory mineral in these rocks, is in fact the Zr-bearing phase.

Vanadium, Chromium, Nickel: V values vary from 100 ppm in hawaiites down to below the limit of detection in the comendites. Taylor (1965) quotes 250 ppm as typical for V in basalts. The Comboyne V values compare closely with those determined for the Mayor Island volcanics (Ewart et al., 1968) where a trachybasalt has 130 ppm V, and it is below the limit of detection in the pantellerites.

Chromium decreases from 86 ppm in the hawaiites to below the limit of detection in the comendites. The relatively low Cr values in the hawaiites compare with values typical of basalt (in excess of 200 ppm, Taylor, 1965), indicating that Cr is rapidly depleted during differentiation.

Depletion is also evident with Ni. Values range from a maximum of 39 ppm in hawaiite to below the limit of detection in comendite. Trends for both Cr and Ni are similar to those reported for the basalt-trachyte-phonolite series from Ua Pu, Marquesas Islands (Bishop and Woolley, 1973) and Gough Island (Le Maitre, 1962). However, the Marquesas series is considerably richer in V than the Comboyne series.

Copper and Zinc: In the Comboyne series hawaiite through to anorthoclase trachyte Cu exhibits restricted variation from 39-21 ppm. Two aphyric comendites (MU6506, MU6569) have 21-25 ppm Cu, and the remainder have Cu values of less than 9 ppm. These

values are comparable with values quoted for alkaline rocks from Marquesas Islands (Bishop and Woolley, 1973), Erta'Ale, Danakil, (Trueil et al., 1971) and Mayor Island (Ewart et al., 1968).

The Comboyne comendites show extreme enrichment in Zn with values ranging from 237-811 ppm. These are considerably higher than values reported for peralkaline rhyolites from Erta'Ale (Danakil) (65-115 ppm) and syenitic rocks from Shonkin Sag (75-105 ppm, Nash and Wilkinson, 1971). There is only a small variation in Zn content in the remainder of the Comboyne series hawaiite through to anorthoclase trachyte (71-107 ppm). These values tend to be slightly less than in similar rocks from Erta'Ale. However, in the more evolved phonolite dykes from Gronnedal-Ika, South Greenland (Gill, 1972) and pantellerite ash flows from Ethiopia (Gibson, 1970), Zn varies from 406-615 ppm and 215-290 ppm respectively.

4.3.1.1 Discussion

Variations in Y, Ga, Zr, Rb, Ba and Sr content effectively separate the alkaline and calc-alkaline rocks (Fig. 4.12). A greater concentration of Rb, Ga, Y and Zr occurs in the silica-rich alkaline rocks compared with calc-alkaline rocks with similar silica contents. Sr is strongly depleted in the comendites relative to the calc-alkaline rhyolites and granites, and there are substantial differences in K/Rb and Rb/Sr ratios (Tables 4.1 and 4.2).

The steady enrichment in Rb, Zr, U, Th, Pb, Y and Ga in the alkaline series, with extreme enrichment in these elements in the comendites, is consistent with trends typically exhibited by basaltic fractionation series (Nash et al., 1969). Similarly, initial enrichment in Ba followed by depletion in the comendites, is consistent with anticipated fractionation trends. Low Cr, Ni and

V in hawaiites and throughout the series, indicates that fractionation of olivine and possibly spinel occurred prior to the formation of the hawaiites. Strong depletion in Sr and Ca with marked decrease in K/Rb and increase in Rb/Sr ratios in the comendites, suggests these were derived from highly fractionated liquids, where there has been substantial removal of plagioclase and sodic alkali-feldspar. The marked increase in Rb/Sr ratios in the benmorites corresponds with the abundant crystallisation of large anorthoclase crystals in these rocks. This is followed by a decrease in the Ba/Sr ratio which reflects the depletion of Sr and Ba in the melt as the result of fractionation of both plagioclase and K-feldspar.

K/Rb ratios in the Comboyne alkaline rocks tend to be higher than those typical of undersaturated alkaline provinces and a comparison of alkali basalts from a number of associations (Table 4.3) suggests this ratio is generally higher in near-saturated compared to undersaturated alkaline rocks. Ridley (1971b) notes that transitional basalts with tholeiitic affinities are depleted in Rb and Sr relative to those with alkaline affinities. Undersaturated associations also generally have higher total alkalis than do near-saturated alkaline associations. Schmincke (1973) noted that the more alkalic Atlantic islands tend to be more potassic, and that it is an increase in K rather than Na that determines the degree of alkalinity. The above data suggest that near-saturated alkaline and transitional basalts, such as the Comboyne series, are derived from parental material where amphibole rather than phlogopite has been the major hydrous phase.

There is substantial experimental and natural rock evidence indicating that amphibole and phlogopite are present in the upper mantle and that the K/Rb ratio reflects the proportion of amphibole

TABLE 4.3

COMPARISON OF K/RB VALUES FOR SELECTED BASALTIC ROCKS

<u>Location</u>	<u>Rock Type</u>	<u>K/Rb</u>	<u>Reference</u>
Comboyne	Hawaiite	609-419	This study
Aden	Transitional Basalt	638	Civetta et al.(1974)
Ert'a Ale (Danakil)	Transitional Basalt	500	" "
Zubair, Red Sea	Transitional Basalt	525	" "
Limosa, Italy	Alkali Basalt	606	" "
Easter Island	Basalt	2075	Baker et al.(1974)
Nandewar	Olivine Basalt	385-474	Abbott (1969)
Southwest Utah	Alkali Olivine Basalt - Transitional	149-540	Lowder (1973)
Assab Range	Ankaramite	471	Civetta et al.(1974)
Assab Range	Alkali Olivine Basalt	345-358	" "
Azores	Alkali Basalt	329-385	Goodman (1972)
Azores	Ankaramite	269-390	" "
Hawaii	Ankaramite	273	" "
Juan Fernandez	Alkali Basalt	533-1120	" "
San Juan Mts.	Alkali Olivine Basalt	400	Lipman (1969)
Madiera	Basalts	299-454	Hughes & Brown (1972)
Madiera	Hawaiite	247-420	"
Cape Verde Is.	Melilite Nephelinite	159	Klerkx et al.(1974)
Cape Verde Is	Olivine Nephelinite	145	" "
Dunedin Volcano	Alkali Olivine Basalt	394	Price & Taylor (1973)
Dunedin Volcano	Hawaiite	285	"
Marquesas Is.	Alkali Olivine Basalt	41-636	Bishop & Woolley (1973)
Marquesas Is.	Hawaiite	90-221	"
Tristan da Cunha	Ankaramite	133	Baker et al.(1964)
Tristan da Cunha	Olivine Basalt	154	" "
Gough Island	Basalt	140-400	Le Maitre (1962)

relative to phlogopite in the melt. K/Rb ratios in amphibole are high (> 1000) compared with phlogopite (200-450), and the addition of phlogopite to the melt lowers the K/Rb ratio and significantly increases the amount of K and Rb (Griffin and Murthy, 1969; Jakes and White, 1970).

Subrounded striated apatite megacrysts are prominent in the more basic porphyritic Comboyne rocks and a high pressure origin is suggested for these and similar apatites described from other localities (Table 3.14). Green and Ringwood (1967) note that alkali olivine basalts are enriched in such 'incompatible elements' as Zr, Rb and U and suggest that these elements are strongly partitioned into any liquid phase in the mantle, and in the absence of a liquid phase are probably incorporated in minor phases such as amphibole, phlogopite, Mg-ilmenite and apatite. It is significant that Zr tends to be higher in undersaturated alkaline rocks relative to tholeiitic rocks (Chao and Fleischer, 1960) as experimental work by Green and Ringwood has shown that at 35-70 km depth (13.5-18 kb) 20% partial melting of a 'pyrolite' mantle will result in a liquid with the composition of alkali olivine basalt, whereas 30% melting will result in an olivine-rich tholeiite. This would result in greater dilution of the component comprising minor phases such as apatite with increased degrees of partial melting. The enrichment of 'incompatible elements' in alkali olivine basalts relative to tholeiitic rocks, and the common occurrence of large apatite crystals (and amphibole) in alkaline rocks and their apparent absence in tholeiitic rocks supports the above suggestion.

It is concluded that trace element abundances in the Comboyne alkaline rocks are consistent with these rocks being members of a single fractionation series.

4.3.2 Tholeiitic Rocks

Trace element abundances (Table 4.2) in the Comboyne tholeiitic andesite (MU6577) are generally similar to those in rocks with comparable SiO_2 values from other tholeiitic associations (Taylor, 1968; Jakes and White, 1970; Gunn, 1971; Jakes and White, 1972). However, the K/Rb ratio in the Comboyne tholeiitic andesite (200) is lower than the value given by Taylor (1968) for average andesite (430), and substantially lower than that quoted by Jakes and White (1970) for island arc tholeiites. These K/Rb ratios contrast with those reported for continental tholeiites which typically exhibit a wide spread from 150 to 1000 (Jakes and White, 1970).

Variations in K/Rb ratios are considered to result from 1) different proportions of amphibole and mica involved in fractional melting, and 2) high level fractionation and contamination. As noted previously, K/Rb ratios in amphibole are substantially higher than in phlogopite. However, low K_2O relative to Na_2O in the tholeiitic andesite indicates that the breakdown of phlogopite was not significant in the formation of the parental melt. The presence of megacrysts and cumulates in the Comboyne tholeiitic andesite indicates that high-level fractionation after the formation of the observed megacryst assemblage has been minor, although major element composition indicates that fractionation did occur prior to this. There is no evidence to indicate that crustal contamination has been a factor in causing low K/Rb ratios in the tholeiitic andesite. The gabbroic inclusions in this rock might be considered foreign xenoliths incorporated in the melt during its ascent through the crust, but the major phase in these inclusions is plagioclase which typically has high K/Rb ratios (Ewart and Taylor, 1969). It is therefore concluded that

the relatively low K/Rb ratios in the Comboyne tholeiitic andesite probably result primarily from fractionation processes involving the removal of a phase with a high K/Rb ratio such as plagioclase. This in turn suggests that fractionation of some of the large plagioclase-rich gabbroic inclusions possibly occurred at upper crustal levels by gravity settling.

4.3.3. Calc-Alkaline Rocks

Rubidium: The average concentration of Rb in the Lorne Basin calc-alkaline rocks with silica in excess of 70% is 104 ppm (Table 4.2). The total range for Rb in the Lorne Basin calc-alkaline rocks is 21 ppm (Lorne Forest diorite and microdioritic xenolith in North Brother microgranite) to 184 ppm (rhyolitic pitchstone). The average Rb value for the Lorne Basin granitic rocks is significantly lower than average values for rhyolite (217 ppm, Heier and Adams, 1964) and granite (150 ppm, Taylor, 1965). In view of the coherence between K and Rb (Taylor, 1965) it is noteworthy that the average K_2O content in the Lorne Basin granites and rhyolites is 3.42% (range 2.50-5.09%) which is also less than average values quoted by Nockolds (1954) for rhyolite (5.46%) and granite (5.35%).

In detail, the Lorne Basin calc-alkaline rocks have low Rb values compared with the Southern Californian Batholith, Lassen Peak, Medicine Lake Highlands, East Central Sierra Nevada granitic rocks (Nockolds and Allen, 1953), the Cape and Snowy Mountain granites (Kolbe and Taylor, 1966b), and the New England Batholith granites (Shaw, 1964; Flood, 1971). Rb in the Lorne Basin calc-alkaline granitic rocks is closer in value to the Central North Island rhyolites (Ewart and Stipp, 1968), and the Talasea, New Britain (Lowder and Carmichael, 1970), Lesser Antilles and Crater Lake (Nockolds and Allen, 1953) volcanics, and in this respect

the Lorne Basin rocks have more affinity with continental margin and some island arc volcanics than to continental plutonic calc-alkaline rocks (Fig. 4.13b).

The Lorne Basin calc-alkaline rocks exhibit only a small variation in K/Rb ratio from basic to silicic types, with no consistent trend towards decrease in K/Rb ratio with increase in silica content (Table 4.2). The K/Rb ratio ranges from 106 to 580 as SiO_2 varies from 51% to 76%. With one exception (pitchstone MU6690 with K/Rb of 106), K/Rb ratios in the granitic rocks range from 255-431. The moderately high K/Rb ratios in the Lorne Basin granitic rocks are consistent with the comparatively low K_2O values in these rocks, and contrast with lower ratios in the Cape granites (98-245), the North Island rhyolitic lavas (172-303), and the Snowy Mountain granitic rocks (100-217), which generally have higher K_2O values.

Strontium: The majority of the calc-alkaline rocks in the Lorne Basin area show only limited variation in Sr content (11-457 ppm). In rhyolites and granites Sr tends to be lower (11-208 ppm) than values quoted by Turekian and Kulp (1956) for average rhyolite (235 ppm) and Taylor (1965) for average granite (285 ppm). The Lorne Basin granitic rocks generally have low calcium values which are reflected in the low Sr values.

Maximum Sr content occurs in the more calcic basaltic andesite and diorites. Low Sr values are also reported for the continental calc-alkaline provinces of Cape granite (Kolbe, 1966), Chilean ignimbrites (El Hinnawi et al., 1969), Southern Californian Batholith, Medicine Lake Highlands (Nockolds and Allen, 1953), the Snowy Mountain granites (Kolbe and Taylor, 1966a), and the Central North Island rhyolites (Ewart and Stipp, 1968).

Rb/Sr ratios vary from 0.01 in the basaltic andesite at Diamond Head to 3.30 in a rhyolite. Typically Rb/Sr ratios for the Lorne Basin silicic rocks are less than one, which contrast markedly with the high ratios exhibited by the peralkaline rhyolites - 8.09-95.83 (Tables 4.1 and 4.2). The Lorne Basin granitic rocks tend to have lower Rb/Sr ratios than the Cape and Snowy Mountain granites and the North Island rhyolitic lavas, resulting from the generally higher Sr and lower Rb values in the Lorne Basin rocks.

Barium: Ba in the Lorne Basin calc-alkaline rocks varies from 101 ppm in the Diamond Head basaltic andesite to 860 ppm in the strongly altered Diamond Head dacite. Values for the granodiorites and granites range from 206-430 ppm and 319-549 ppm respectively. These values are lower than those for rocks with similar SiO_2 values in the Chilean ignimbrite province, Cape and Snowy Mountain granites, the Southern Californian Batholith, Lassen Peak, Crater Lake, Medicine Lake Highlands, East Central Sierra Nevada, the Central North Island rhyolites and the New England Batholith and, as noted previously, it is significant that in many instances K_2O values are also lower and in this respect the Lorne Basin rocks show greater correlation with continental margin calc-alkaline rocks.

Ba/Sr ratios vary from 0.19 - 0.42 in the more basic Lorne Basin rocks, and from 0.85 - 2.13 in the granodiorites. With the exception of the South Brother microgranite which has a value of 30.9, the Lorne Basin granites and rhyolites have Ba/Sr ratios ranging from 1.53 - 6.54. These are generally lower than the North Island rhyolites (resulting from lower Ba values) which in turn are higher than "average granite" (Ewart and Stipp, 1968). Both Ba and Sr are depleted in the South Brother microgranite. Similar depletion occurs in some of the strongly altered rhyolites and it

is suggested that this probably partly results from the leaching of alkalis, Ba and Sr during deuteric alteration.

Yttrium: The Lorne Basin calc-alkaline rocks tend to have high Y values when compared with other calc-alkaline associations. Values range from 20 to 92 ppm, and these extremes occur in the Lorne Forest diorite and the Diamond Head quartz diorite respectively. Taylor (1965) gives average values for granite and granodiorite of 40 and 30 ppm respectively. In comparison, the range for similar rock types in the Lorne Basin is 31 - 104 ppm.

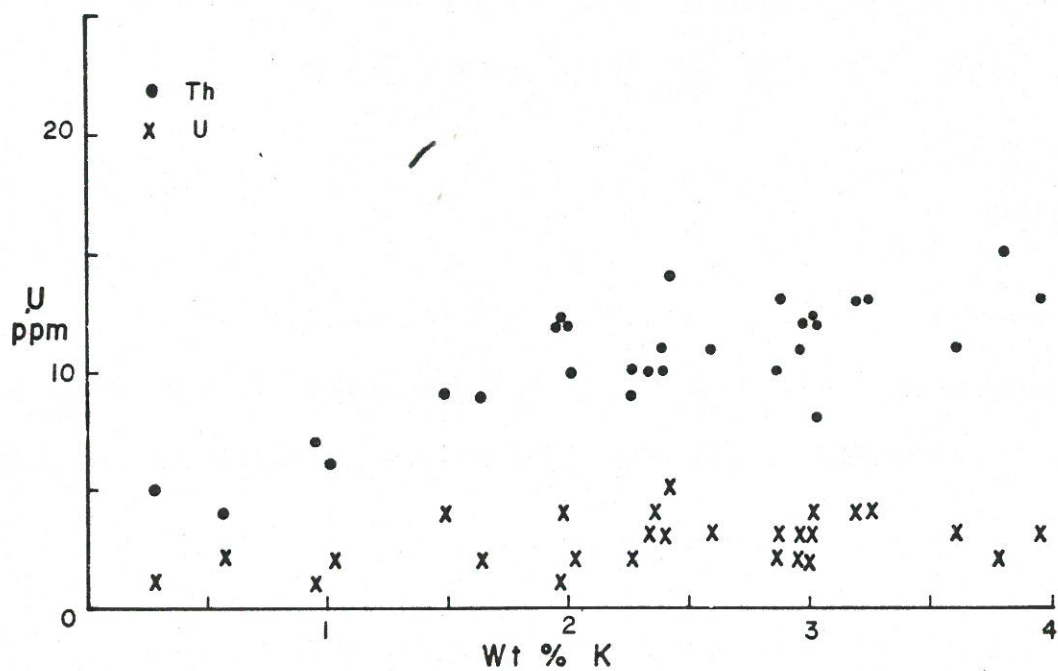
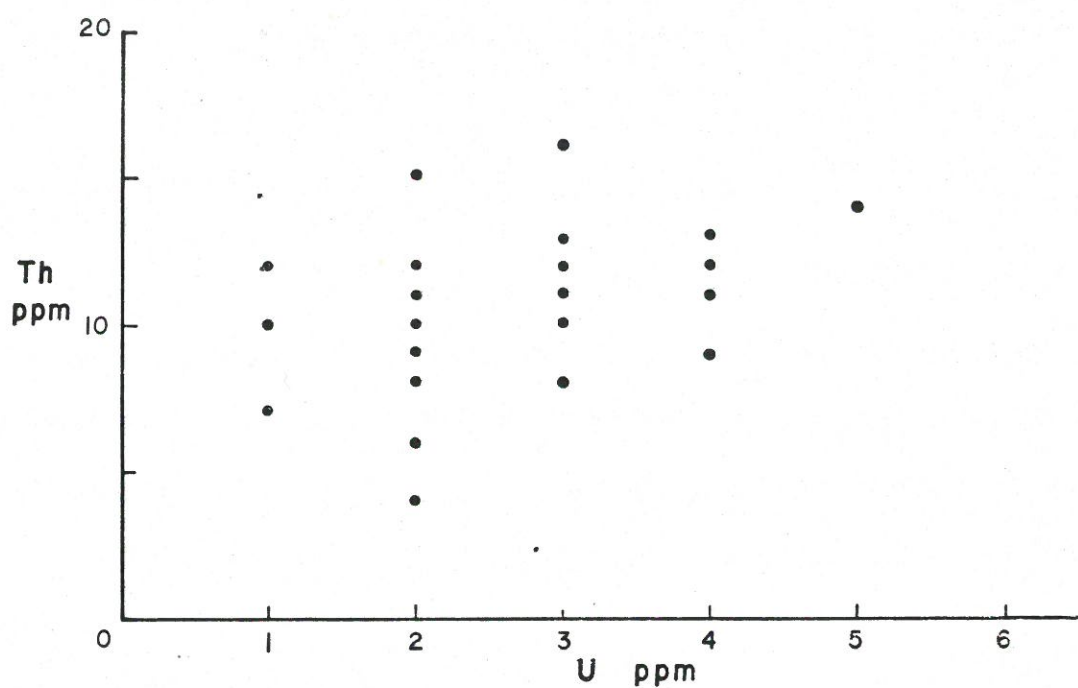
Nockolds and Allen (1953) have shown in a comparison of calc-alkaline associations (Southern Californian Batholith, Lassen Peak, Crater Lake, Lesser Antilles, Medicine Lake Highlands, East Sierra Nevada) that Y tends to be constant through individual series. More recently Macdonald and Edge (1970) have noted that Y is generally concentrated in residual liquids. The calc-alkaline rocks in the Lorne Basin exhibit rather wide scatter with no consistent tendency towards enrichment with increase in SiO_2 content.

Thorium and Uranium: It has been established that Th, U and K generally exhibit marked coherency, particularly in single differentiation suites (Heier and Roger, 1963; Ewart and Stipp, 1968; Gill, 1970). However, Figures 4.14 a and b show that although Th in the Lorne Basin calc-alkaline rocks increases steadily with increase in K, plots of U against K, and Th against U, show rather wide scatter with no consistent trends.

Th content varies from 6 - 16 ppm in the Lorne Basin granitic rocks (with the exception of MUG672 which has anomalously high Th - 104 ppm). The range 6 - 16 ppm is lower than the values quoted by Taylor (1965) for average granite (17 ppm), and Kolbe and Taylor (1966b) and Flood (1971) for plutonic granites (5 - 62 and 11 - 40 ppm respectively). Lowder and Carmichael (1970) give an

Fig. 4.14a Plots of Th versus U (ppm) in Lorne Basin
calc-alkaline rocks.

Fig. 4.14b Plots of Th (ppm) versus Wt % K, and U (ppm)
versus Wt % K in the Lorne Basin calc-alkaline
rocks.



average Th value of 5 ppm for five rhyolites and rhyodacites from the transitional tholeiitic - calc-alkaline island arc rocks at Talasea, New Britain. Commonly, U and Th are lower in island arc calc-alkaline provinces relative to continental provinces.

Variations in U content in the Lorne Basin calc-alkaline rocks is restricted, ranging from 1 ppm in the more basic rocks, to 5 ppm in rhyolites and granites. As with Th values, these tend to be slightly lower than the Snowy Mountain and Cape granites and the New England granites. Th/U ratios vary from 2.0 to 11.0, the more basic rocks generally having the higher values. Rhyolites and granites mostly have Th/U ratios between 3.0 and 4.0 which compare with the crustal average of 3.5 (Taylor, 1965).

Zirconium: Zr in the Lorne Basin calc-alkaline rocks varies from 154 - 756 ppm. Average values for granodiorites and granites/rhyolites are 379 ppm and 377 ppm respectively, and these contrast with values given for average granodiorite and granite of 140 ppm and 180 ppm respectively (Taylor, 1965). The Lorne Basin Zr values tend to be higher than those given for other continental calc-alkaline associations by Nockolds and Allen (1953), Taylor (1965), Ewart and Stipp (1968), Kolbe (1966), and Kolbe and Taylor (1966b). In contrast, the New England, U.S.A., Narragansett Pier and Westerly calc-alkaline granites (Buma et al., 1971) have comparable Zr values to the Lorne Basin granites, and the Chilean ignimbritic rhyolites show extreme enrichment in Zr with 215 - 1200 ppm (El Hinnawi et al., 1969).

Gallium: The Lorne Basin rocks generally exhibit only limited variation in Ga from 18 - 25 ppm. The range in Ga values is similar to those reported by Nockolds and Allen (1953), Ewart and Stipp (1968), Kolbe and Taylor (1966b) for a number of calc-alkaline associations, and comparable with average values for

granitic rocks (Taylor, 1965).

Lead: The maximum range observed in Pb values in the Lorne Basin calc-alkaline rocks varies from 5 ppm in a xenolith in the North Brother microgranite to 41 ppm in a Diamond Head dacite. Pb is generally lower in the basic and intermediate rocks, and ranges from 5 - 7 ppm in the basaltic andesite and diorites. Pb varies from 6 - 20 ppm in the granites and rhyolites and from 7 - 21 ppm in the granodiorites. These values are comparable with average values given by Taylor (1965) for granites and granodiorites (20 ppm and 15 ppm respectively). Pb tends to be slightly higher in the Snowy Mountain, Cape and New England, N.S.W. granitic rocks. In contrast, the Chilean ignimbritic rhyolites are enriched in Pb relative to calc-alkaline rocks with similar SiO_2 values.

Vanadium, Chromium, Nickel: V values vary from 223 ppm in the Diamond Head basaltic andesite to below the limit of detection in the North Brother microgranite. Values for the Lorne Basin granitic rocks range from < 1 - 81 ppm, the higher values being for granodiorites. The Diamond Head and Lorne Forest diorites have V values between 103 - 138 ppm. Generally, V values in the Lorne Basin calc-alkaline rocks tend to be slightly lower than average values quoted by Taylor (1965) and Taylor and White (1965) for granites, granodiorites and andesites, and are also lower than rocks with similar SiO_2 values from the Central North Island rhyolites and the Chilean ignimbrites. There is a closer correlation with the Snowy Mountain and Cape granites.

Low Cr contents are found in all the Lorne Basin calc-alkaline rocks consistent with their generally silicic character. The granitic rocks have Cr values varying from below the limit of detection to 6 ppm, which is comparable to Cr contents in the Snowy Mountain leucogranites, the Cape granites and the North

Island rhyolites. A maximum value of 47 ppm Cr occurs in a Lorne Forest diorite, and this value is lower than values quoted for andesites by Taylor and White (1965), but equivalent to those determined in the Yeoval diorites (Gulson et al., 1972).

The diorite Ni values (11 - 20 ppm) are similar to those reported for the Yeoval diorites, but lower than those quoted by Taylor and White (1965) for andesites. In the granitic rocks Ni is generally below the limit of detection and compares with the value of 0.5 ppm given by Taylor (1965) for average granite and values quoted by Koble and Taylor (1966b) for the Snowy Mountain leucogranites and the Cape granites, and Ewart and Stipp (1968) for the North Island rhyolites.

Copper and Zinc: The Lorne Basin granitic rocks have Cu values varying from 12 - 30 ppm which compare closely with values quoted by Taylor (1965) for average granites and granodiorites. Cu values in the North Island rhyolites, the Snowy Mountain and Cape granites are within a similar range, whereas in the Chilean ignimbrites Cu shows wide variation from 'normal' values to 1800 ppm. In the Diamond Head and Lorne Forest diorites Cu ranges from 23 - 40 ppm. These values tend to be slightly lower than those given by Taylor and White (1965) and Gulson et al. (1972) for andesites and diorites.

Zn in the granitic rocks varies from 46 - 86 ppm, and in the diorites from 46 - 52 ppm. The highest value (148 ppm) occurs in a sulphide-rich dacitic rock at Diamond Head. There is little published data on Zn values in igneous rocks, but Taylor (1965) gives average values for basalt, granodiorite and granite of 100, 60 and 40 ppm respectively.

4.3.3.1 Discussion

Trace element variation is considered to be particularly sensitive to fractionation processes, and the general lack of systematic variation in both major and trace element chemistry between the basic basaltic andesite and diorites and the more silicic intermediate and acid Lorne Basin calc-alkaline rocks indicates that these rocks are not related by normal fractionation processes unless subsequent alteration has destroyed the evidence for this relationship.

K/Rb ratios for the Lorne Basin granitic rocks are higher than those typical of rocks which have been derived by crystal fractionation, and low Rb, K, U, Th and Pb also suggest that extreme differentiation by crystal fractionation has not occurred. The granites and rhyolites are depleted in both Sr and Ca, and in this respect are similar to the Central North Island rhyolites, the Snowy Mountain gneissic and leuco-granites, and the Cape granites. Low Sr and Ba values can be attributed to strong fractionation, but this is inconsistent with low Rb, K, Th and Pb values and comparatively high K/Rb ratios, and could be a direct result of low Sr, Ca and Ba content in the parental material.

V, Cr and Ni values in the Lorne Basin granitic rocks are comparable or slightly lower than values quoted for similar rock types in other areas, and are consistent with an origin for the host rock by crystal fractionation during which time these elements would have been selectively incorporated into early forming minerals. However, low V, Cr and Ni values would also be expected in a melt formed by the partial melting of crustal material, as these ferromagnesian elements would mainly be concentrated in the more refractory ferromagnesian silicates and oxides.

Griffin and Murthy (1969) give Ba contents of hornblende and

mica as 220 ppm and 1400 ppm respectively. The low Ba values in the Lorne Basin rocks suggest that the host melt was derived from the fusion of an amphibole-bearing parental material. This is also supported by the relatively high K/Rb ratios shown by these rocks. Low Ba could also result from the fusion of parental material in which plagioclase rather than alkali-feldspar was the major feldspar phase.

Marked enrichment in Zr occurs in the granitic rocks and although this increase could be attributable to extreme fractionation, it can also be related to a melt obtained by anatexis of sedimentary material (Taylor, 1965). A similar anatectic origin has been proposed for the Central North Island rhyolites (Ewart and Stipp, 1968), the Westerly and Narragansett Pier calc-alkaline granites (Buma et al., 1971), the Cape and Snowy Mountain granites (Kolbe and Taylor, 1966b) and the Chilean ignimbritic rhyolites (El Hinnawi et al., 1969). Zr values in these associations are variable, but are consistently higher than those reported by Turekian and Wedepohl (1961) and Taylor (1965) for granitic rocks. Variable Zr could be anticipated in rocks formed from crustal material by anatexis as it would be directly dependent on the composition of the parental material.

As previously noted, both major and trace element data indicate that the Lorne Basin basaltic andesite and diorites are not related to the associated granitic rocks by crystal fractionation. Trace element abundances in the more silicious rocks are consistent with an origin by anatectic processes. However, both the major and trace element compositions of the more basic rocks indicate that they could not have formed from a partial melt of crustal material. Except for significantly lower Ba and Rb, trace element contents in the diorites are closely comparable with those given by Gulson

(1972) for the Yeoval dioritic complex. To a lesser extent they are similar to trace element values quoted by Taylor and White (1965), Gill (1970), Lowder and Carmichael (1970) and Jakes and White (1972) for island arc andesites. The high Y value in the Diamond Head quartz diorite is puzzling, but it does have approximately 10% more modal amphibole than the Lorne Forest diorites, and Gulson (1972) has determined that more than 90% of Y in the Yeoval diorites is in hornblende.

In summary, it is concluded that trace element data for the silicic Lorne Basin rocks are consistent with their derivation by partial melting of crustal material. In contrast, trace element values in the more basic rock types suggests they formed from a melt derived from fractionation of a basalt which in turn was obtained by melting of the mantle.

CHAPTER 5 - EXPERIMENTAL STUDY ON A

NEAR-SATURATED HAWAIIITE

5.1 Introduction

The common occurrence in the Comboyne hawaiites of megacryst and cumulate phases which show evidence of being in disequilibrium with the host melt during the final stages of crystallisation suggests that these are either foreign xenoliths or cognate phases which crystallised at depth.

High-pressure lherzolite inclusions are widely reported in silica-undersaturated alkaline basalts and these have been interpreted as substantiating experimental evidence (Yoder and Tilley, 1962; Green and Ringwood, 1967; O'Hara and Yoder, 1967; Ito and Kennedy, 1967; O'Hara, 1968; Kushiro, 1968; Green, 1970; Bultitude and Green, 1971) that the host basalts are primary melts formed by partial melting of mantle material. In contrast, it has generally been accepted (Green and Ringwood, 1967; Ito and Kennedy, 1967) that more evolved alkaline rocks, including hawaiites and mugearites, have formed by crystal fractionation at low pressures from a mantle derived parent basaltic magma. Recently, however, postulated high-pressure phases have been recognised in a limited number of hawaiites, mugearites, benmoreites, trachytes and phonolites (Kuno, 1964; Uchimizu, 1966; Wright, 1969; Binns et al., 1970; Wass, 1971; Irving, 1971; Borley et al., 1971; Green et al., 1974) indicating that at least some of these more evolved alkaline rocks may have formed at elevated pressures.

Green et al. (1974) have attributed the origin of lherzolite nodule-bearing, nepheline-normative hawaiite, mugearite and benmoreite to fractionation of parental basanite or alkali olivine basalt at depths of > 30 km. The $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ value for these rocks

is < 60, contrasting with values of about 70 for nodule-bearing alkali olivine basalt and basanite, and indicating fractionation of Mg-rich phases. This proposed fractionation proceeded under hydrous conditions and was controlled by kaersutitic hornblende, with subordinate olivine and possibly clinopyroxene. These phases are found as megacrysts in the nepheline-normative hawaiites.

The proposed high-pressure megacryst and cumulate assemblages in the Comboyne hawaiites contrast in several ways with the compositions discussed above. The Comboyne hawaiites are hypersthene-normative, lack lherzolite nodules and contain megacryst/cumulate assemblages of olivine, orthopyroxene, clinopyroxene, Fe-Ti oxide, plagioclase and apatite (i.e. kaersutitic hornblende is notably absent). The evolved nature of the Comboyne hawaiites is indicated by their $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ values of 44 - 50. Proposed high-pressure phases have been reported in compositionally similar hawaiites (Kuno, 1964; Binns et al., 1970) but so far no high-pressure experimental work has been carried out on such compositions. Accordingly, an experimental study of one of the Comboyne hawaiites (from Mt. Baldy - MU6498) has been undertaken to evaluate the postulated high-pressure origin of the megacryst/cumulate phases and delineate their P-T conditions of crystallisation.

5.2 Host Rock and Megacryst/Cumulate Assemblage

The composition of the Mt Baldy hawaiite (MU6498) is given in Table 5.1. It falls in the alkaline field of the $\text{Na}_2\text{O} + \text{K}_2\text{O}$ versus SiO_2 plot of Macdonald and Katsura (1964) but in terms of normative components it is transitional between the alkaline and tholeiitic igneous suites, since it has minor normative hypersthene (0.7%) after the $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio has been adjusted to 0.3 (page 102). Other megacryst-bearing hawaiites in the area have up to 15.6%

TABLE 5.1ANALYSES OF NATURAL HAWAIIITE AND HAWAIIITE GLASS USED
IN THE EXPERIMENTAL RUNS.

	<u>Experimental Hawaiiite</u>	<u>Natural Hawaiiite</u>
SiO ₂	52.20	51.13
TiO ₂	2.54	2.36
Al ₂ O ₃	16.21	16.01
Fe ₂ O ₃	.92	2.63
FeO	10.45	8.76
MnO	ND	.16
MgO	4.45	4.17
CaO	6.94	6.83
Na ₂ O	4.18	4.63
K ₂ O	2.11	2.16
P ₂ O ₅	ND	1.16
$\frac{100\text{Mg}}{\text{Mg} + \text{Fe}^{2+}}$	43.1	45.9
Or	12.46	12.76
Ab	35.35	39.16
An	19.23	16.53
Di	12.65	8.08
Hy	.25	.74
Ol	13.81	11.74
Mt	1.33	3.81
Il	4.92	4.48
Ap	-	2.69
$\frac{100\text{An}}{\text{An} + \text{Ab}}$	35.3	29.7

Analyses recalculated to 100%.

normative hypersthene. Mineralogically the hawaiites have closer affinities to alkali olivine basalts than tholeiites, with both olivine and alkali-feldsparⁱ occurring in the groundmass. A normative plagioclase composition of andesine (An_{30}) and a D.I. value of 52 classifies the Mt. Baldy rock as a hawaiite falling close to the hawaiite-mugearite boundary (Coombs and Wilkinson, 1969).

Megacrysts of olivine, clinopyroxene, orthopyroxene, ilmenite, titanomagnetite, (?)plagioclase and apatite have been observed in the Comboyne hawaiites. Cumulate assemblages include: olivine-orthopyroxene-plagioclase, olivine-orthopyroxene-ilmenite, olivine-clinopyroxene-ilmenite, orthopyroxene-clinopyroxene-olivine, plagioclase-clinopyroxene-orthopyroxene, clinopyroxene-apatite, titanomagnetite-apatite. Marked corrosion of the megacryst and cumulate phases indicates they were not in equilibrium with the host melt during the final stages of crystallisation. Megacryst/cumulate olivine is enriched in Mg relative to the coexisting phenocryst and groundmass phases and corrosion has caused rounding of crystals. Large, clear crystals of clinopyroxene generally exhibit fracturing and not cleavage, and have significantly higher Ca-Tschermak component and Na_2O , and lower CaO (Table 3.3) values than the associated phenocrysts. The megacrysts/cumulates have strongly sieved rims where in contact with the melt. Orthopyroxene megacrysts are rimmed and in some instances completely replaced by a granular aggregate of olivine and clinopyroxene. They have higher Al_2O_3 values than is typical for low-pressure orthopyroxenes in

ⁱ Fine grain-size in the groundmass in the Mt. Baldy hawaiite has prevented positive identification of alkali-feldspar. Alkali-feldspar has been analysed in a similar but coarser-grained hawaiite.

tholeiitic rocks. The $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratios for the clinopyroxene, orthopyroxene and olivine megacrysts/cumulates in the hawaiiite used in the experimental study vary from 67.0 - 77.9, 61.5 - 73.7 and 64.8 respectively (Fig. 5.1a).

Megacrysts of corroded ilmenite range up to 4 mm and commonly have rims and intergrowths of titanomagnetite. These ilmenites have higher Al_2O_3 and MgO (Table 3.10) than groundmass ilmenites. Euhedral but slightly rounded apatite megacrysts up to 1.5 mm long are common in the Comboyne hawaiiites. They occur as discrete crystals or associated with titanomagnetite, olivine, clinopyroxene and plagioclase as inclusions or cumulates. The cores of these large apatite crystals characteristically have tiny acicular inclusions of an (?)Fe-Mg oxide intersecting at 60° . Plagioclase megacrysts and phenocrysts are chemically indistinguishable. Crystallisation of plagioclase contemporaneously with the megacryst/cumulate phases is indicated by the presence of plagioclase in some of the cumulates. There is overlap in chemical composition between megacryst and cumulate phases of olivine, orthopyroxene, clinopyroxene and ilmenite.

5.3 Experimental Results

5.3.1 Qualitative Results

Experimentally determined phase relationships for the hawaiiite at pressures ranging from 5 to 15 kb under hydrous and anhydrous conditions are listed in Table 5.2 and illustrated in Figure 5.1b.

The major features are summarised below:

1. In 'dry' runs olivine and plagioclase are the near-liquidus phases at 5 kb, olivine, clinopyroxene and plagioclase are the near-liquidus phases at 7.5 kb, and clinopyroxene and plagioclase are the near-liquidus phases at 10 kb.

Fig. 5.1a Left: Composition of pyroxenes and olivines in natural hawaiiite. Filled circles = clinopyroxene megacrysts/cumulates. Open circles = clinopyroxene phenocrysts. Crosses = orthopyroxene megacrysts/cumulates. Filled triangle = olivine cumulate. Open triangles = olivine phenocrysts.

Right: Composition of pyroxenes and olivine from hawaiiite experimental runs. Filled circles = clinopyroxene. Filled triangle = olivine.

Fig. 5.1b Experimentally determined phase relationships for the Mt. Baldy hawaiiite.

Al₂O₃
wt%

PRESSURE Kb

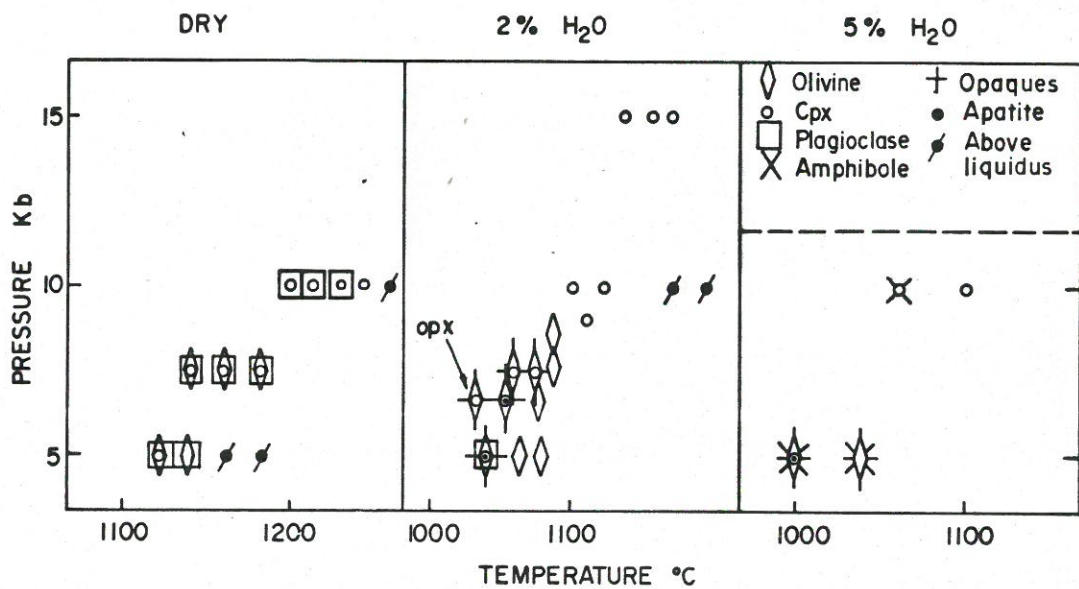
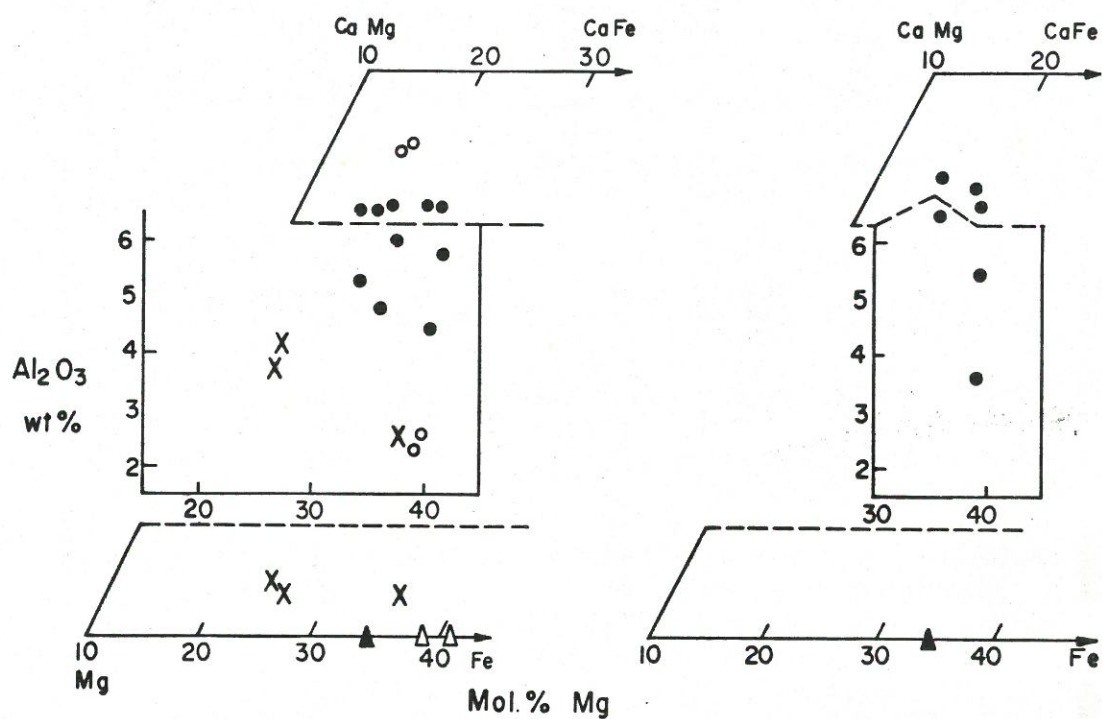


TABLE 5.2

RESULTS OF EXPERIMENTAL RUNS ON MOUNT BALDY HAWAIIITE

Run	P.kb	T. °C.	Time mins.	Capsule	Results
<u>DRY CONDITIONS</u>					
186	5	1180	60	Pt	Above liquidus
189	5	1160	60	Pt	Above liquidus
266	5	1145	55	Pt	Olivine + subordinate plagioclase. Near liquidus
205	5	1140	60	Pt	Olivine + plagioclase
265	5	1135	60	Pt	Olivine + plagioclase + (?)clinopyroxene. Well crystallised
194	5	1120	60	AgPd	Olivine + plagioclase + clinopyroxene. Well crystallised
206	7.5	1180	30	Pt	Olivine + clinopyroxene + plagioclase. Near liquidus
200	7.5	1160	60	Pt	Olivine + clinopyroxene + plagioclase
199	7.5	1140	30	Pt	Olivine + clinopyroxene + plagioclase. Well crystallised.
152	10	1260	30	Pt	Above liquidus
264	10	1240	40	Pt	Clinopyroxene. Near liquidus
153	10	1230	40	Pt	Clinopyroxene + plagioclase. Near liquidus
155	10	1215	45	Pt	Clinopyroxene + plagioclase. Well crystallised
145	10	1200	45	Pt	Clinopyroxene + plagioclase. Well crystallised
143	10	1160	45	Pt	Clinopyroxene + plagioclase. Well crystallised
<u>2% H₂O</u>					
178	5	1080	90	AgPd	Olivine. Very near liquidus
177	5	1070	90	AgPd	Olivine. Near Liquidus
176	5	1050	120	AgPd	Olivine + clinopyroxene + plagioclase + ilmenite + apatite. Well crystallised
202	6.5	1080	90	AgPd	Olivine. Very near liquidus
208	6.5	1060	60	AgPd	Olivine + clinopyroxene + (?)apatite + opaque
215	6.5	1040	60	AgPd	Olivine + clinopyroxene + orthopyroxene + opaque.
183	7.5	1090	90	AgPd	Olivine. Very near liquidus
193	7.5	1080	60	AgPd	Olivine + clinopyroxene + ilmenite. Well crystallised
179	7.5	1070	90	AgPd	Olivine + clinopyroxene + opaque. Well crystallised
214	8.25	1090	45	AgPd	Olivine + clinopyroxene. Near liquidus
201	9	1110	90	AgPd	Clinopyroxene. Very near liquidus
162	10	1180	45	Pt	Above liquidus
166	10	1160	60	Pt	Above liquidus
167	10	1120	60	AgPd	Clinopyroxene. Near liquidus
172	10	1100	90	AgPd	Clinopyroxene. Well crystallised

TABLE 5.2 Cont'd

Run	P.kb	T. °C.	Time mins.	Capsule	Results
<u>2% H₂O</u>					
174	15	1160	60	Pt	Clinopyroxene. Very near liquidus
173	15	1150	60	Pt	Clinopyroxene. Near liquidus
175	15	1130	60	Pt	Clinopyroxene. Well crystallised
<u>5% H₂O</u>					
159	5	1040	120	AgPd	Olivine. Very near liquidus
160	5	1000	60	AgPd	Olivine + amphibole + ilmenite + apatite + plagioclase
157	10	1100	60	AgPd	Clinopyroxene. Near liquidus
158	10	1060	120	AgPd	Clinopyroxene + amphibole + apatite. Well crystallised.
<u>5% Oxalic Acid</u>					
191	5	1100	120	AgPd	Above liquidus - saturated
216	5	1080	45	AgPd	Olivine + clinopyroxene
195	10	1140	15	Pt	Clinopyroxene. Near liquidus
188	10	1120	60	AgPd	Clinopyroxene.
187	10	1100	60	AgPd	Clinopyroxene + amphibole. Well crystallised.

Note: Glass phase present in all runs.

2. Olivine is the liquidus phase at 5 kb with 2% H₂O.
3. In all hydrous runs clinopyroxene is the liquidus phase at 10 kb.
4. With 2% H₂O present, olivine is replaced by clinopyroxene as the liquidus phase in the pressure range 7.5 to 9 kb.
5. Orthopyroxene, clinopyroxene, Fe-Ti oxide and (?)apatite crystallise together at approximately 40^o C below the liquidus at 6.5 kb with 2% H₂O.
6. Amphibole is not present in runs with 2% H₂O, but appears as an important near-liquidus phase with 5% H₂O.
7. The appearance of bubbles in runs with 5% oxalic acid supports the observation of Hill and Boettcher (1970) that there is little solubility of CO₂ in basic silicate liquids at low to moderate pressures. The few exploratory runs conducted in the presence of CO₂ indicate it has no significant effect on the nature or sequence of phases in the near-liquidus interval.
8. The narrow temperature interval over which several phases appear close to the liquidus points to the near-cotectic nature of the hawaiite composition, especially at 6 - 7 kb for runs with 2% H₂O.

5.3.2 Analytical Results

Electron microprobe analyses of olivine, clinopyroxene, amphibole and ilmenite are shown in Table 5.3. In addition to these, plagioclase and apatite have been positively identified, although fine grain-size has prevented their accurate analysis. Olivine has been analysed from two experimental runs at 5 kb with 2% and 5% H₂O. They are similar in composition and have 100Mg/Mg + Fe²⁺ ratios of 63.8 and 64.9, which compare with the value

TABLE 5.3

MICROPROBE ANALYSES OF EXPERIMENTAL AND NATURAL MEGACRYST/CUMULATE PHASES

	Olivine	Olivine	Olivine	Cpx	Cpx	Cpx	Cpx	Amphibole	Amphibole	Amphibole	Amphibole	Ilm	Ilm	Ilm	
H ₂ O	2	5	Natural	2	2	5	Natural	5	5	5	Natural	2	5	Natural	
kb	5	5	Cumulate	5	10	10	Cumulate	5	5	10	Cumulate	5	5	Megacryst	
T°C	1070	1040		1050	1120	1060		1000	1040	1060	Benmoreite	1050	1000		
SiO ₂	36.36	36.06	36.36	49.67	49.40	51.08	49.00	41.95	40.81	41.56	45.78	-	-	-	
TiO ₂	.12	.11	-	2.00	1.93	1.07	1.41	4.55	4.61	4.25	1.21	51.86	51.64	51.99	
Al ₂ O ₃	-	.13	-	5.45	6.55	3.59	5.84	11.91	12.92	12.68	7.18	1.13	.76	.39	
Cr ₂ O ₃	-	-	.11	.18	.18	.15	-	-	-	.14	-	.22	.14	-	
FeO	30.84	31.50	30.98	11.77	9.13	11.21	12.70	13.14	14.64	13.36	15.67	37.63	40.61	40.84	
MnO	.28	.63	.34	.34	.48	.45	.16	.26	.33	.34	.22	.86	.79	.49	
MgO	32.03	31.16	32.05	13.29	13.62	13.56	13.11	12.18	11.95	12.06	11.98	5.40	4.35	4.28	
CaO	.33	.40	.16	16.68	17.89	18.48	17.02	10.79	9.63	10.02	11.25	.58	.51	.16	
Na ₂ O	-	-	-	.47	.76	.40	.64	2.54	2.71	2.64	2.82	-	-	-	
K ₂ O	-	-	-	.15	.06	-	-	.79	1.11	1.08	1.59	-	-	-	
Cation Content on Basis of:															
	4 Oxygen Atoms			6 Oxygen Atoms				23 Oxygen Atoms				Ilmenites recalculated after the method of Anderson (1968 b)			
Si	.9897	.9861	.9895	1.8561	1.8320	1.9101	1.8383	6.1972	6.0415	6.1460	7.1720	Al ₂ O ₃	1.13	.76	.39
Al	-	.0042	-	.2401	.2864	.1582	.2581	2.0730	2.2536	2.2109	1.3241	Cr ₂ O ₃	.22	.14	-
Ti	.0024	.0023	-	.0562	.0537	.0300	.0399	.5057	.5129	.4731	.1427	CaTiO ₃	1.40	1.24	.41
Cr	-	-	.0024	.0054	.0052	.0046	-	-	-	.0163	-	MgTiO ₃	16.10	13.00	12.74
Mg	1.2983	1.2701	1.2999	.7400	.7530	.7559	.7330	2.6819	2.6359	2.6587	2.7963	MnTiO ₃	1.83	1.67	1.06
Fe	.7014	.7205	.7051	.3679	.2833	.3507	.3983	1.6233	1.8122	1.6529	2.0530	FeTiO ₃	74.78	78.61	81.19
Mn	.0066	.0145	.0077	.0107	.0152	.0142	.0048	.0325	.0408	.0426	.0283	Fe ₂ O ₃	3.07	3.99	2.72
Ca	.0095	.0118	.0046	.6680	.7110	.7405	.6840	1.7082	1.5273	1.5872	1.8878				
Na	-	-	-	.0339	.0545	.0291	.0466	.7276	.7786	.7563	.8345				
K	-	-	-	.0073	.0028	-	-	.1494	.2105	.2029	.3173				
Total	3.0079	3.0095	3.0092	3.9885	3.9971	3.9931	4.0033	15.6989	15.8132	15.7468	16.5560				
100Mg Mg + Fe ²⁺	64.9	63.8	64.8	66.8	72.7	68.3	64.8	62.3	59.3	61.7	57.7	Hm	3.89	4.83	3.24
Pyroxene end members calculated by Kushiro's (1962) method. Total iron as FeO															
NaAlSi ₃ O ₆				4.02	5.23	2.81	4.22								
CaTiAl ₂ O ₆				5.95	5.71	3.09	4.28								
CaAl ₂ SiO ₆				4.36	6.11	3.49	6.11								
CaSiO ₃				29.27	31.02	34.85	29.97								
MgSiO ₃				33.14	33.94	33.74	32.63								
FeSiO ₃				22.30	17.68	21.37	23.62								

of 64.8 for cumulate olivine in the natural rock.

Clinopyroxene has been analysed from experimental runs with 2% H₂O at 5 kb and 10 kb and 5% H₂O at 10 kb. 100Mg/Mg + Fe²⁺ ratios for the experimentally produced clinopyroxenes vary from 66.8 to 72.7, which compare closely with the range 64.8 to 78.9 for the natural megacryst and cumulate clinopyroxenes. The close similarity in Al₂O₃ values, and in particular the Ca-Tschermak component, in the natural and experimental clinopyroxenes is significant considering the pressure dependence of the Al^{VI} content in pyroxenes (Fig. 5.1a).

As naturally occurring amphiboles have not been observed in the Comboyne hawaiites, no direct comparison can be made with amphiboles produced experimentally in runs with 5% H₂O at 5 and 10 kb. However, megacryst amphiboles do occur in more evolved rocks (benmoreites) in the area. The experimentally produced amphiboles are kaersutites and contrast with the edenitic amphiboles analysed in the Comboyne benmoreites in having lower SiO₂, and much higher TiO₂ and Al₂O₃.

The experimentally obtained ilmenites are chemically similar to ilmenite megacrysts in the natural hawaiite. There is an increase in FeO from 37.6 to 40.8% in runs with 2% and 5% H₂O respectively. As the temperature of the experimental runs is known, analyses of the experimentally produced ilmenites enables an estimate of the oxygen fugacity of the host melt to be made (Buddington and Lindsley, 1964). Oxygen fugacity values thus estimated range from 10^{-11.2} atm at 5 kb, 1050° C and 2% H₂O, to 10^{-11.8} atm at 5 kb, 1000° C and 5% H₂O. These values compare closely with oxygen fugacity and temperature estimated from ilmenite-magnetite megacryst intergrowths in the natural hawaiites which range from 10^{-11.4} atm at 1000° C to 10^{-15.4} atm at 825° C.

The experimental results follow the normal trend of decreasing oxygen fugacity with decreasing temperature, and are consistent with oxygen fugacity values estimated for natural Comboyne hawaiites and other basaltic rocks (Wass, 1973).

5.4 Discussion

Orthopyroxene has only been identified in one experimental run with 2% H₂O at 6.5 kb and 1040° C, where it coexists with olivine, clinopyroxene, Fe-Ti oxide and (?)apatite. Experimental investigations on basaltic compositions at elevated pressures by Green and Ringwood (1967), Green and Hibberson (1970), Bultitude and Green (1971), Irving (1971) and Thompson (1974) have shown that the stability field of orthopyroxene in alkaline compositions is restricted, and that these Ca-poor pyroxenes form on or near the liquidus at pressures intermediate to those where olivine is replaced by clinopyroxene as the liquidus phase. The replacement of olivine by clinopyroxene as the liquidus phase in experimental runs with 2% H₂O at 9 kb indicates this is the upper pressure limit for the formation of orthopyroxene and the observed megacryst/cumulate mineral assemblage. The crystallisation of orthopyroxene at 6.5 kb in a closely similar paragenesis to that observed in the natural rock suggests that the hawaiite megacryst/cumulate assemblage precipitated at a depth of approximately 20-24 km prior to rapid ascent to higher crustal levels.

Plagioclase is prominent in some hawaiite cumulates and textural evidence indicates that crystallisation of plagioclase continued after that of the ferromagnesian minerals. The absence of plagioclase coexisting with olivine, clinopyroxene and orthopyroxene in the experimental run at 6.5 kb suggests that plagioclase crystallisation was delayed relative to these phases. This is

supported by experimental results of D.H. Green and Ringwood (1967), T.H. Green and Ringwood (1968), Nesbitt and Hamilton (1970), Green and Hibberson (1970) and Irving (1971) which indicate that the appearance of plagioclase in hydrous runs is delayed relative to ferromagnesian phases at elevated pressures. However, plagioclase is expected to occur at 6.5 kb with 2% H₂O at slightly lower temperatures.

Recent experimental investigations at atmospheric pressures on differentiated rocks from Skye (Thompson, 1972b), Aden (Humphries and Cox, 1972) and Reunion (Humphries, 1972), show that plagioclase is the liquidus phase in most instances, followed by olivine, and then by clinopyroxene at lower temperatures. The dominance of plagioclase phenocrysts over olivine, and the scarcity of clinopyroxene phenocrysts in the Comboyne hawaiites support the conclusion that crystallisation of the megacryst/cumulate assemblage at elevated pressures was followed by a period of phenocryst crystallisation at upper crustal levels when partial re-equilibration (e.g. reaction at rims and resorption) of megacrysts occurred.

The experimentally obtained ilmenite with a composition similar to that occurring naturally supports the suggestion that the ilmenite megacrysts precipitated at elevated pressures. The positive identification of apatite in the hawaiite experimental runs at 5 kb is also consistent with the suggestion that the large striated apatites in the Comboyne hawaiites appeared early in the crystallisation sequence at elevated pressures.

5.5 Conclusions

The results of this study indicate that the experimentally obtained near-liquidus assemblage of the hawaiite with 2% H_2O at 6.5 kb is similar to the partly resorbed megacryst/cumulate assemblage in the natural hawaiite. The close similarity in composition and paragenesis between the natural and experimental mineral assemblages and, in particular, the very restricted field of orthopyroxene stability shown in this and other experimental investigations, indicates that the megacryst/cumulate assemblage in the Comboyne hawaiites crystallised in the pressure range of approximately 6.5 - 8 kb, at approximately 1040 - 1080° C.

The appearance of amphibole in near-liquidus experimental runs with 5% H_2O at both 5 and 10 kb limits the H_2O content in the natural melt to a value of less than 5%. Similarly, the early appearance of plagioclase in the 'dry' experimental runs contrasts with the apparent dominance of ferromagnesian phases near the liquidus in the natural megacryst assemblage. The above data indicate that the H_2O content in the natural hawaiite melt was in the vicinity of 2%. The occurrence of olivine and pyroxene (chemically equivalent to the naturally occurring phases) as near liquidus phases in experimental runs with 2% H_2O substantiates this conclusion.

CHAPTER 6 - PETROGENESIS AND CONCLUSIONS

6.1 Alkaline Rocks

The mineralogical and chemical data for the Comboyne alkaline rocks are consistent with these rocks being representative of a single fractionation series derived from a common parental melt. Variations in mineral and major and trace element compositions (as discussed in Chapters 3 and 4), between hawaiites, mugearites, benmoreites, trachytes and comendites, follow smooth curves on compositional diagrams and can be attributed to the fractionation of the observed megacryst/cumulate and phenocryst phases.

Mineral extract calculations provide a means of testing the hypothesis that the Comboyne alkaline rocks represent a single fractionation series. Table 6.1 shows that the removal of observed megacryst/cumulate and phenocryst phases from the different rock types results in a near approach in composition to the associated more evolved types. As a first step in the fractionation sequence, the extraction of megacryst/cumulate phases from the hawaiites results in a composition closely similar to the mugearites and consistent with the observed trends in major and trace element composition.

Although the abundance of megacryst/cumulate phases indicates that extrusion to the surface has been rapid, it is impossible to determine accurately the initial proportion of the megacryst/cumulate phases. However, distribution coefficients for Mg:Fe for both naturally and experimentally produced olivines (Tables 3.1 and 5.3) which are consistent with equilibrium values determined by Nicholls (1974) indicate the bulk composition of the hawaiites may be representative of the composition of the melt at the time

TABLE 6.1

CALCULATED MINERAL EXTRACT COMPOSITIONS - ALKALINE ROCKS

Ref.No:	1	2	3	4	5	6	7	8	9	10	11	12	13
Sample No:	6498		6607		6566		6630		6569		6645		6658
SiO ₂	51.13	52.15	52.20	56.63	56.31	64.73	65.68	68.88	69.23	71.41	73.24	70.84	70.22
TiO ₂	2.36	2.30	2.29	1.01	1.42	.18	.37	.09	.26	.33	.16	.09	.14
Al ₂ O ₃	16.01	15.77	15.83	15.52	15.35	17.50	15.98	14.97	14.26	13.39	12.45	15.39	16.18
Fe ₂ O ₃	2.63												
FeO	8.76	12.38 ⁺	12.39 ⁺	12.00 ⁺	10.97 ⁺	4.19 ⁺	4.96 ⁺	3.24 ⁺	4.16 ⁺	2.87 ⁺	3.40 ⁺	.85 ⁺	.98 ⁺
MnO	.16	.15	.19	.15	.14	.12	.16	.21	.08	.10	.06	.22	.04
MgO	4.17	2.56	2.43	1.13	1.44	.24	.16	.09	.07	.09	.15	.09	.05
CaO	6.83	5.44	5.57	3.95	3.77	1.03	1.17	.25	.47	.60	.39	.26	.25
Na ₂ O	4.63	5.08	4.73	5.15	4.79	5.74	5.69	5.70	6.13	5.90	5.52	5.86	6.42
K ₂ O	2.16	2.81	2.80	3.42	4.54	6.18	5.56	6.21	5.00	4.97	4.41	6.36	5.17
P ₂ O ₅	1.16	1.01	1.30	1.01	.88	.06	.10	.13	.12	.15	.05	.13	.02

+ = Total iron as FeO.

1. Mt. Baldy hawaiiite
2. Mt. Baldy hawaiiite less 2½% olivine, 5% clinopyroxene, 2½% orthopyroxene, 15% plagioclase, 1% ilmenite, 1% apatite.
3. Hannon Vale mugearite
4. Hannon Vale mugearite less 15% plagioclase, 2% olivine 2½% clinopyroxene, 2½% orthopyroxene, 1% ilmenite, 4% magnetite, 2½% alkali-feldspar, 1½% apatite.
5. Comboyne benmoreite
6. Comboyne benmoreite less 5% plagioclase, 10% anorthoclase, 2½% clinopyroxene, 2½% amphibole, 2½% olivine, 2% apatite, 7% magnetite
7. Mt. Gibraltar anorthoclase trachyte
8. Mt. Gibraltar anorthoclase trachyte less 20% alkali feldspar, 1% clinopyroxene, 2% magnetite, 0.5% ilmenite
9. Comboyne comendite
10. Comboyne comendite less 20% alkali-feldspar, 2% magnetite
11. Mt. Coxcomb comendite
12. Mt. Gibraltar anorthoclase trachyte less 20% alkali-feldspar, 1% clinopyroxene, 4% magnetite, 0.5% ilmenite
13. Comboyne mafic-poor comendite.

the observed megacryst/cumulate assemblage precipitated.

The proportion of the different phases removed to obtain the mugearite composition is consistent with proportions observed in the Mt. Baldy hawaiite. Marked decreases in MgO and Ni in the mugearites reflect the importance of olivine and pyroxene fractionation in the crystallisation interval from hawaiite to mugearite. Slight enrichment in iron in the mugearites parallels trends observed in other mildly alkaline-transitional associations and results from the early crystallisation of magnesium-rich olivine and pyroxene. The tendency towards absolute iron-enrichment and depletion in the anorthite component observed in the intermediate rocks would be compounded by conditions of low p_{H_2O} causing the early appearance of plagioclase in these rocks relative to more strongly alkaline, silica-undersaturated basalts.

This interpretation is similar to the conclusions of Barberi et al. (1971) for the mildly alkaline-transitional association of the Erta'Ale range, Ethiopia. However, the Erta'Ale rocks have markedly lower Al_2O_3 and higher FeO than the corresponding Comboyne rocks suggesting that either the parental melt had higher Al_2O_3 and FeO, and possibly closer tholeiitic affinities, or alternatively, low p_{H_2O} could have caused a greater proportion of plagioclase fractionation in the more basic rock types.

The removal of megacryst/cumulate phases observed in the hawaiites and mugearites from the mugearites results in a near approach to the benmoreite composition (Table 6.1). In turn, the extraction of the benmoreite (?) megacryst/cumulate phases from the benmoreite gives a composition similar to that of the Comboyne anorthoclase trachytes. However, in order to derive anorthoclase trachyte from benmoreite it has been necessary to extract 7% magnetite from the benmoreite composition. This exceeds the amount

observed in the benmoreites, although magnetite-rich cumulates do occur in this rock indicating that pronounced fractionation of this phase may have taken place. It is also possible that unsampled compositions intermediate to benmoreite/mafic trachyte and anorthoclase trachyte occur.

Although the Comboyne series, up to and including the anorthoclase trachytes, can be explained by the removal of the observed megacryst/cumulate and phenocryst phases of the preceding less evolved rock type, derivation of the comendites is more complex. In a number of associations, and in particular in the East African Rift Valley, the abundance of rhyolites and phonolites relative to intermediate types has led to the suggestion that these highly evolved rocks have formed by the melting of crustal material. Although some continental peralkaline rhyolites may have an anatectic origin (Bailey and Macdonald, 1970), their frequent association with mildly alkaline-transitional rocks and the strong similarity of these associations in widely scattered locations, suggests that many peralkaline rhyolites are highly evolved fractionates of alkaline basalts. This conclusion is supported by 1) the unlikelihood of lower crustal rocks being of suitable composition to provide anatectic peralkaline melts, 2) Sr isotope values consistent with a mantle origin and 3) variations in major and trace element trends generally consistent with a crystal fractionation origin.

The marked enrichment in SiO_2 observed between trachytes and the associated comendites follows strong fractionation of Fe-Ti oxides in the benmoreites and mafic trachytes, resulting in rapid depletion in FeO and enrichment in SiO_2 (page 108). The relatively SiO_2 -poor comendites (commonly occurring as flows) can be derived from the anorthoclase trachytes by removal of alkali-feldspar,

clinopyroxene, magnetite and ilmenite. In turn, extraction of alkali-feldspar and magnetite from the SiO_2 -poor comendite composition results in a near-approach to the more evolved comendites. An increase in pH_2O at this stage would further shift the quartz-feldspar cotectic towards quartz, and it is significant that it is only in those rocks which commonly form plugs that quartz phenocrysts occur.

As noted in Chapter 4 there is some evidence to indicate loss of alkalis during the crystallisation or post-crystallisation history of the comendites (rocks with normative corundum and modal riebeckite/arfvedsonite and aegirine, scattering of plots on $\text{SiO}_2 - \text{Al}_2\text{O}_3 - 3\text{SiO}_2 - (\text{Na}_2\text{O} + \text{K}_2\text{O})3\text{SiO}_2$ ternary diagrams with some rocks with peralkaline mineralogy plotting in the peraluminous field). However, feldspar and major element data (page 65) indicate the Comboyne rocks are not strongly peralkaline and this alkali-loss was not of major significance.

Alkali rhyolites which have lower SiO_2 and FeO and higher Al_2O_3 than the majority of comendites are closely associated with the comendites in the field. Low FeO in these rhyolites reflects the almost complete absence of ferromagnesian minerals. Crystal extract calculations indicate these rocks are direct differentiates of the anorthoclase trachytes, and result from more pronounced fractionation of magnetite than occurred in the formation of the majority of comendites.

As has already been noted, hawaiite is the most basic rock found in the Comboyne area. The possibility that more basic rock types may be buried beneath the volcanic pile is supported by the occurrence of alkali olivine basalts in the closely comparable Nandewar Mountain series (which Abbott (1969) argued resulted from high-level fractionation of alkali olivine basalt \rightarrow rhyolite) and

similar mildly alkaline associations. However, experimental work on the Mt. Baldy hawaiite and other natural basaltic compositions (Green and Hibberson, 1970; Irving, 1971; Thompson, 1974), and reports of high-pressure phases in rocks as evolved as trachytes (Wright, 1969), suggest the hawaiites rose rapidly from 20 - 24 km. This removes the necessity for more basic parental melts at upper crustal levels and indicates a deep-level hawaiite may be the parental magma of the Comboyne sequence. Thus high-level fractionation need not have been extensive, at least for the composition range hawaiite - benmoreite/mafic trachyte, and thus there is less of a problem in explaining large volumes of intermediate and silicic rocks without the accompanying basic fraction. Also, the derivation of evolved rocks such as hawaiites, mugearites, benmoreites and trachytes from depth, when linked with the tectonic controls proposed by Bailey (1964) and Gill (1973), provides an explanation for the abundance of intermediate and/or silicic rocks relative to the associated basic types commonly found in continental provinces (South Greenland, Upton et al., 1971; East Africa, Bailey, 1964; Aden and Little Aden, Cox et al., 1970; Nandewar, Abbott, 1969) in contrast to oceanic island environments (St. Helena, Baker, 1969). However, a deep level origin for the more evolved rock types can only be inferred if high-pressure megacrysts or lherzolite xenoliths are observed and generally, to date, this is not the case.

Thompson (1972a) notes a chemical gap between basalts and intermediate lavas in anorogenic volcanic suites and attributes this to a mantle origin for both magma batches. He describes a mechanism whereby the parental basaltic magma is held at depth (cf Gill, 1973) and undergoes high-pressure partial crystallisation. When these bodies are approximately half solid, rupturing (caused by "stresses

associated with heating, cooling and magma production in the mantle beneath a volcanic area") results in the forceful extrusion upwards of more evolved interstitial fluid. There is also evidence of such compositional gaps between basic and more evolved rock types in many mildly alkaline associations (e.g. Nandewar, Abbott, 1969; Dogo, Uchimizu, 1966; Danakil, Barberi et al., 1970; Skye, Thompson et al., 1972; Easter Island, Baker et al., 1974). The Comboyne series shows similar groupings of the various rock types (particularly when a few anomalous rock types are excluded) and it is only in the anorthoclase trachytes and comendites that there is some continuum in composition (Fig. 4.3a). These gaps could result from insufficient sampling, but the apparent widespread occurrence of this trait suggests that it reflects some fundamental mechanism. Thus it is envisaged that the different rock types were derived from magma batches formed by crystal fractionation at depth resulting from the parental melt being held at upper mantle or lower crustal levels prior to extrusion to higher crustal levels. The identification of possible megacryst/cumulate phases in the Comboyne hawaiites, mugearites and benmoreites supports this suggestion. Experimental work has not been undertaken on the mugearites and benmoreites, but textural evidence and data from other alkaline associations suggests that at least some mugearites, benmoreites and trachytes have been derived at elevated pressures (Green et al., 1974). Although cumulates in the Comboyne benmoreites consisting of anorthoclase, olivine, edenitic amphibole, apatite and Fe-Ti oxides cannot be unequivocally interpreted as crystallising at depth, the crystallisation of amphibole, apatite, olivine and ilmenite from the hawaiite experimental runs with 5% H₂O and 5 kb is consistent with this suggestion.

The derivation of basalts and more evolved rocks at high pressures also explains the occurrence of slightly silica under-saturated and saturated rocks spanning the low pressure thermal divide of the basalt tetrahedron (Ridley, 1971b); Thompson et al., 1972). Such basalts typically occur in mildly alkaline-transitional associations which include more evolved saturated rock types. The transitional nature of these rocks is considered to be a direct function of the composition of the parental melt, and the low oxygen fugacity of these mildly alkaline melts does not support suggestions (Kuno, 1965) that near-saturation results from the removal of magnetite.

Two approaches can be used to speculate on the composition of the parental melt from which the Comboyne alkaline series was derived. Firstly, some indication can be obtained by comparison with the more basic rocks from similar mildly alkaline-transitional series (Nandewar, Dogo, Aden, Danakil, Skye). These basalts are not strongly alkaline and, like the associated intermediate types, are only mildly silica-undersaturated or near-saturated in character. This suggests that the melt parental to the Comboyne hawaiite showed similar characteristics and thus is probably a near-saturated alkaline olivine basalt.

Secondly, consideration of experimental work on basaltic compositions also provides an insight on the melting behaviour and subsequent crystallisation trends of liquids derived from postulated mantle compositions. Green and Ringwood (1967) and Green (1970) have shown that melts formed from 'pyrolite' are dependent on the depth and degree of melting, the depth of separation of magma from residual crystals, and the water content. A small degree of partial melting and subsequent magma separation at >40 km depth result in a melt with alkali olivine basalt affinities, whereas more marked

melting results in tholeiitic liquids. It may be anticipated, therefore, that intermediate degrees of melting will produce a spectrum of 'transitional' liquids. It is concluded that these transitional melts are derived from 20 - 30% melting of 'pyrolite' at a depth of 40 - 70 km (Green, 1970). Variations within this range of melting would result in a continuum of mildly alkaline-transitional compositions from those more nearly alkaline in character (Nandewar, Dogo, Comboyne), to those with greater tholeiitic affinities (Easter, Ascension and Bouvet Islands).

It is emphasised that the megacryst/cumulate phases in the Comboyne hawaiites are compositionally markedly different from similar minerals in lherzolite inclusions described from strongly undersaturated alkaline basalts, and that lherzolite inclusions have not been observed in the Comboyne rocks. Frey and Green (1974) have concluded from major and trace element abundances, that lherzolite inclusions in the Western Victorian basaltic lavas represent residual material after the melting of mantle domes in the lithosphere. This residual material has subsequently been modified by a melt which has migrated to the top of the low velocity zone. Mineral phases in these lherzolite inclusions have $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ values generally in excess of 90, and these contrast with values between 65 - 81 in the Comboyne hawaiite megacryst/cumulate phases. This supports the conclusions already made that these hawaiites do not represent a primitive mantle melt, but have evolved from a mantle derived melt (i.e. one with $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ values of 65 - 75) by the separation of olivine, pyroxene and possibly plagioclase. The fractionation of these phases from mildly alkaline near-saturated basalts at a maximum of 40 - 70 km and a minimum of 20 - 24 km depth, contrasts with the amphibole-dominated fractionation of wetter, more undersaturated,

lherzolite-bearing, nepheline-normative suites representing much smaller degrees of melting of 'pyrolite' than the drier transitional melts (Green et al., 1974).

Alkaline rocks are generally recognised as forming in tectonically stable continental areas, and both on and away from mid-ocean ridges in oceanic environments (Bailey, 1974; Borley, 1974). Large-scale alkaline activity has also occurred in areas adjacent to major fault zones, including the African Rift system, the Rhine-Oslo graben, the Gardar Rift (Bailey, 1974), and the Monteregian province, Quebec (Philpotts, 1974). Bailey (1972) noted that both rifting and alkaline magmatism are related to cratogenic uplift and suggests this uplift may be the mechanical response to forces transmitted through the rigid plate. This in turn may result in partial melting and the tapping of this mantle melt initiated by a decrease in pressure.

It is envisaged that during a period of crustal uplift the primitive Comboyne melt rose to the base of the crust and fractionated there to hawaiite (possibly as the result of the crust acting as a density filter (Gill, 1973)). The hawaiite in turn was rapidly extruded to the surface during periods of crustal extension. The more evolved mugearites and benmoreites could also have formed at a similar level as the hawaiites by fractionation. Alternatively, they could have formed at lower, but still elevated, pressures as batches of magma were held in the lower crust prior to rapid rises to shallower levels in an episodic manner. The absence of megacrysts in the anorthoclase trachytes indicates that these were derived by crystal fractionation from benmoreites and/or mafic trachytes in the upper crust. Similarly, fractionation of the anorthoclase trachytes towards a comendite composition also occurred at shallow levels.

Wellman and McDougall (1974) have noted in the Cainozoic igneous province of Eastern Australia that slightly undersaturated alkaline volcanism (which includes the Comboyne province) appears to have migrated southward with time over the period 33 my (Cape Hillsborough, Queensland) to 6 my (Mt. Macedon, Victoria). As sea-floor spreading associated with the formation of the Tasman Basin is reported to have ceased 60 my ago (Hayes and Ringis, 1972), it would appear that the mildly alkaline volcanism in Eastern Australia is not directly related to this event. However, as noted by Wellman and McDougall, this volcanism does show a remarkable correlation with the separation of Australia and Antarctica, considered to have started 55 my ago. This suggests that the mildly alkaline volcanism in Eastern Australia was initiated by uplift and tensional stresses (cf Bailey, 1964, 1972, 1974) associated with the northward migration of the Australian (Indian) plate (Wellman and McDougall, 1974).

6.2 Tholeiitic Rocks

The occurrence of volcanic provinces with both alkaline and tholeiitic rocks (including the well documented Hawaiian Island and British Tertiary provinces) has been ascribed to variable depth of initiation of diapiric upwelling of solid mantle resulting in differing degrees of partial melting and depths of magma separation from the residual crystals (Green and Ringwood, 1967). Rare tholeiitic rocks are associated with the mildly alkaline Comboyne series, and similar occurrences are reported from Aden (Cox et al., 1970), Ethiopia (LeBas and Mohr, 1970), and Nandewar (Abbott, 1965), and large volumes of tholeiitic rocks are closely associated with alkaline rocks in the northeast of New South Wales (Duggan and Wilkinson, 1973).

Textural and experimental data (see below) indicate that olivine and orthopyroxene megacrysts in the Comboyne tholeiitic andesite crystallised at elevated pressures, and the presence of these megacrysts suggests this rock has not undergone significant low-pressure fractionation.

Experimentally determined phase relationships on selected samples from the olivine-rich lavas of Nuanetsi (Cox et al., 1974) indicate that equilibration of olivine and orthopyroxene in these lavas occurred in a pressure range of 6 - 12 kb. Similarly, Nicholls and Lorenz (1973) experimentally duplicated pyroxenes in K-rich tholeiites from S.W. Germany under conditions of $p_{\text{Load}} = 6 - 10 \text{ kb}$, $T = 1280 - 1080^{\circ} \text{ C}$ and 2 - 4 % H_2O , and found that the maximum pressure of olivine crystallisation with 2% H_2O was 6 - 7 kb, and that with 4% H_2O it increased to 8 - 10 kb. Green and Ringwood (1967) report that olivine and orthopyroxene are near-liquidus phases in olivine tholeiite between 9 - 13 kb, whereas at lower pressures liquidus olivine is followed by plagioclase and clinopyroxene. The Comboyne tholeiitic andesite contains at least two generations of olivine, the more Mg-rich type having $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ values slightly lower than those in the associated orthopyroxene megacrysts (76 and 78 respectively) suggesting that these phases precipitated together (Green and Ringwood, 1967). The tholeiitic andesite also contains gabbroic aggregates which consist primarily of plagioclase associated with clinopyroxene, olivine and ilmenite. $100\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ ratios for the ferromagnesian minerals in these aggregates indicate they crystallised from a melt less Mg-rich than that from which the orthopyroxene and olivine megacrysts precipitated. However, the plagioclase is less calcic than the associated groundmass plagioclase, suggesting that the gabbroic aggregates crystallised at elevated p_{Load} or $p_{\text{H}_2\text{O}}$ relative to the groundmass phases (but subsequent to

the associated orthopyroxene and olivine megacrysts).

Experimental investigations carried out by T.H. Green (1968, 1972) and D.H. Green and Ringwood (1967) on tholeiitic and calc-alkaline compositions show that up to 9 kb under dry conditions plagioclase is the dominant liquidus phase and is followed by subordinate clinopyroxene.

Although the experimental work of Cox et al. (1974), D.H. Green and Ringwood (1967) and T.H. Green (1968, 1972) on Mg-rich picritic lavas, olivine tholeiites and Al-rich tholeiitic and calc-alkaline rocks cannot be directly applied to the relatively low Mg and Al Comboyne tholeiitic andesite, the observed textural and mineralogical relationships do concur with their results. The tholeiite used by Nicholls and Lorenz (1973) in their experimental study is more closely comparable. This rock is broadly similar in composition to the Comboyne tholeiitic andesite, but has higher MgO and K_2O and lower FeO, CaO and Na_2O . Their results also support the conclusion that the orthopyroxene and olivine megacrysts initially crystallised at pressures of approximately 6 kb. This was followed by abundant crystallisation of plagioclase (with minor clinopyroxene and olivine) during the melt's rapid ascent to higher crustal levels.

Experimental data (T.H. Green et al., 1967; D.H. Green and Ringwood, 1967) indicate that the parental melt of the Comboyne tholeiitic andesite formed from the melting of 'pyrolite' at up to 60 km depth followed by fractionation of Mg-rich phases (as indicated by the relatively low $100Mg/Mg + Fe^{2+}$ value (50) of the Comboyne tholeiitic andesite) to a depth of approximately 20 km, before a rapid rise through the crust prior to extrusion.

6.3 Calc-Alkaline Rocks

The Lorne Basin cal-alkaline rocks are characterised by an abundance of highly evolved granodiorites, granites and rhyolites and the complete absence of more basic rocks that can be positively identified as parental to these. There is strong chemical evidence to support the conclusion that the associated diorites do not belong to the same volcanic episode as the more silicic rocks (Chapter 4), although marked alteration of a large proportion of these rocks makes interpretation based on chemical criteria tenuous.

Field relationships, mineralogy and the generally fine grain-size of the granitic rocks indicate that they were emplaced at high crustal levels. Zoning and sieving of feldspars suggests that equilibrium was not maintained between crystals and melt.

Although the major element composition of the evolved granitic rocks is consistent with a fractionation origin, trace element data indicate that these rocks were not formed by extreme differentiation by crystal fractionation (page 139). The major and trace element abundances in these evolved rocks support the conclusion that the parental melt was derived by partial fusion of crustal material and that differences in the compositions of the rocks results from variations in fractionation and consolidation history, and subsequent post-crystallisation alteration processes.

An anatectic origin has been proposed for the Central North Island rhyolites (Ewart and Stipp, 1968), the Cape and Snowy Mountain granites (Kolbe and Taylor, 1966b), and the Chilean ignimbritic rhyolites (El Hinnawi et al., 1969), and these all show chemical similarities to the Lorne Basin granitic rocks, particularly in trace element abundances. These highly siliceous rocks are not strongly enriched in the residual elements Rb, K, Th and Pb and in

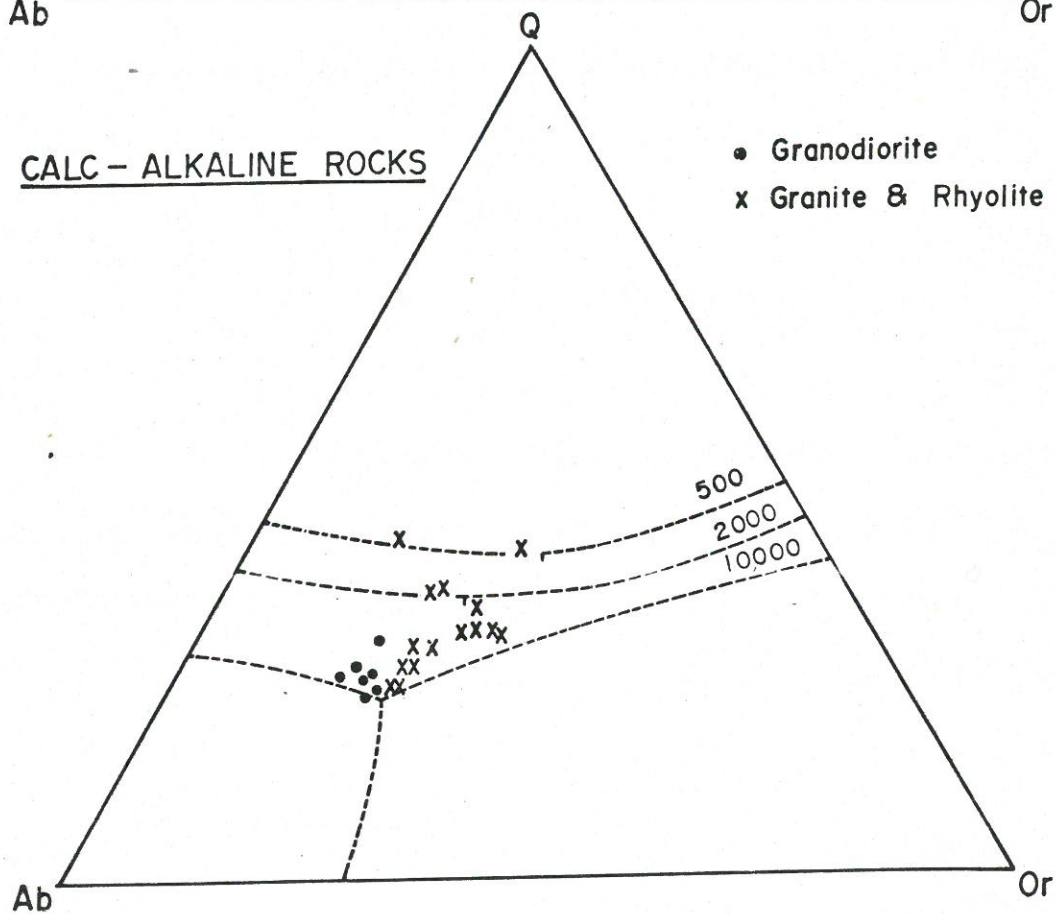
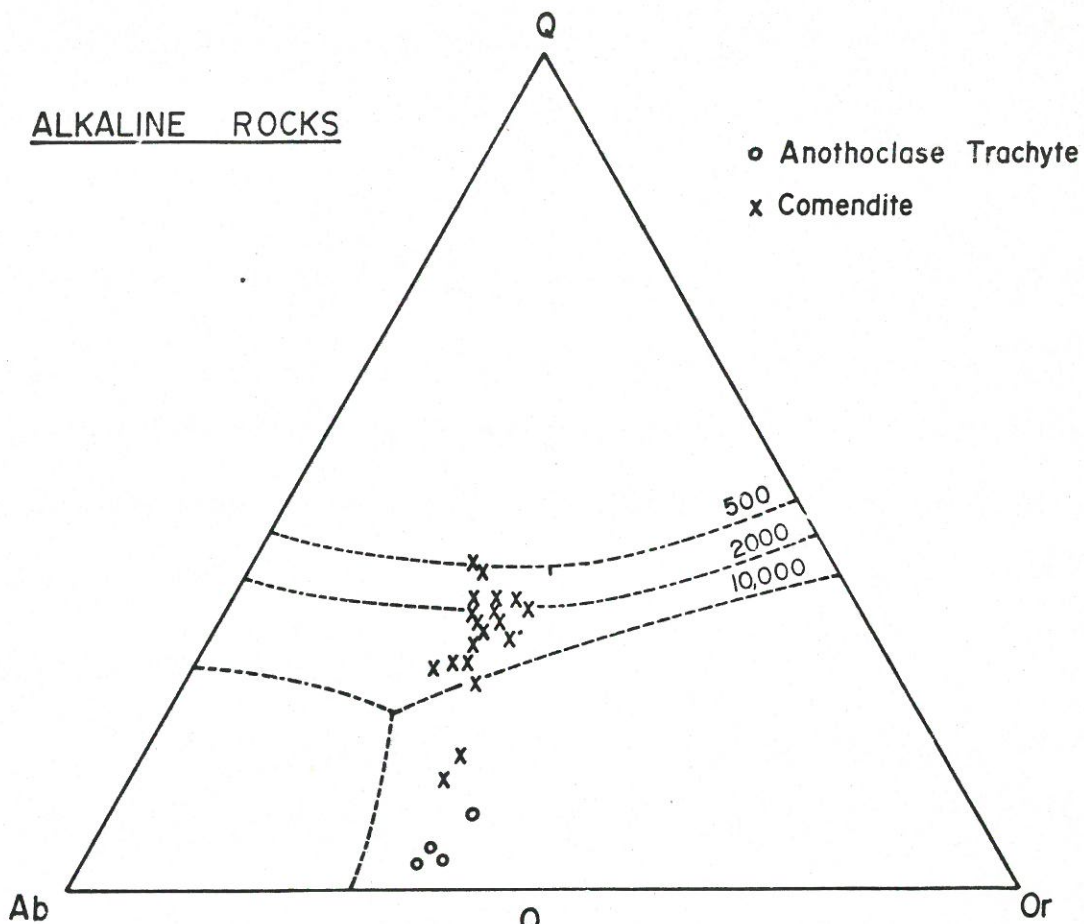
this regard contrast markedly with the Comboyne peralkaline rhyolites for which a long history of crystal fractionation has been proposed.

Experimental studies by Winkler and Von Platen (1960), Piwinski and Wyllie (1968), Wyllie and Tuttle, 1957; and Brown and Fyfe (1970) have indicated that granitic melts can be derived from the partial melting of crustal material and support suggestions that a large proportion of continental granitic rocks are of anatectic origin. The negative slope of the water-saturated solidus curves places a limitation on the upward movement of water-saturated melts. Thus, high-level granites and rhyolites such as those occurring in the Lorne Basin must have crystallised from hot, dry melts (Brown and Fyfe, 1970). The general occurrence of pyroxene rather than amphibole and the scarcity of mica in the silicic Lorne Basin rocks support this conclusion. Ewart and Stipp (1968) suggest that low Rb, U and Th values and "high temperature" mineralogy in the Central North Island rhyolites result from the parental crustal material being of intermediate andesitic composition. A similar conclusion can be made for the Lorne Basin granodiorites, granites and rhyolites, and this is supported by K/Rb and Ba values.

Brown and Fyfe (1970) have shown experimentally that a range of water-undersaturated granitic melts can be generated at moderate pressures and temperatures by the partial fusion of crustal material containing water in the hydrated solid phases only. Their results indicate that melts involving the amphibole breakdown reaction are enriched in CaO and Na_2O and depleted in SiO_2 relative to those in which the breakdown of mica is involved. With the exception of the South Brother microgranite, the Lorne Basin granitic rocks cluster around the minima in the system

Quartz - Albite - Orthoclase between 2 - 10 kb (pH_2O), with a concentration of plots between 4 - 10 kb (Fig. 6.1). This is consistent with both a crystal fractionation or a partial melt origin. The Lorne Basin granodiorites plot on the Quartz - Albite side of the minimum in the Quartz - Albite - Orthoclase ternary diagram and Brown and Fyfe (1970) suggest that rocks plotting in this area form from high-temperature melts involving the breakdown of amphibole. In contrast, the granites and rhyolites plot towards the Quartz - Orthoclase side of the minimum which Brown and Fyfe consider consistent with rocks formed from a low-temperature partial melt of crustal rocks where muscovite is the hydrous phase. However, as noted on page 164, the high emplacement level and "high temperature" mineralogy of these rocks indicate that they formed from hot, dry melts and it is proposed that the granites and rhyolites were derived from the less evolved granodioritic melt by minor crystal fractionation rather than by direct partial melting of crustal material. This is supported by the relatively high K/Rb ratios of these rocks which indicate they formed from a melt derived from the fusion of amphibole rather than mica-bearing parental material. The removal of early forming plagioclase from the granodiorite melt would drive the residual liquid towards a more quartz-orthoclase enriched composition. This sequence is supported by field relationships indicating that the granodiorites were intruded ahead of the granites. If the granites and granodiorites were both direct partial melts it would be anticipated that the initial melt extruded from depth would be granitic in composition, followed by a more basic composition as the degree of partial melting increased. The range in composition of pyroxenes associated with calcic-oligoclase and sodic andesine in the Lorne Basin granitic rocks indicates that some of these are xenoliths (page 67)

Fig. 6.1 a and b. Plots of normative Q - Ab - Or for evolved alkaline (D.I. > 85) and calc-alkaline (D.I. > 65) rocks. Dashed lines show position of cotectic and minimum in the simple system at 500, 2,000, and 10,000 bars (pH_2O). After Tuttle and Bowen (1958) and Luth et al. (1964).



picked up at depth (and possibly of refractory nature), which exhibit varying degrees of disequilibrium with the melt.

The relatively high MgO, FeO and CaO values of the Lorne Basin diorites precludes a crustal anatectic origin for these rocks. The diorites are quartz-normative and have very high Al_2O_3 and relatively low K_2O values. In both major and minor element composition these rocks are similar to 'Normal' andesites and diorites found in island arcs and active continental margins.

Although there are a number of hypotheses relating to the origin of calc-alkaline rocks, only two are considered to be relevant in discussing the more basic gabbroic-dioritic types similar to the Lorne Basin diorites. These are:

- 1) Crystal fractionation from a mantle derived basaltic melt (Bowen, 1928; Osborn, 1959, 1962), and
- 2) A two-stage process in which mantle derived basalts are extruded to the surface and subsequently sink back into the mantle where they undergo partial melting to give rise to calc-alkaline magmas (Green and Ringwood, 1968).

The absence of more basic rocks considered to be parental to the Lorne Basin diorites makes conclusions relating to (1) above difficult. However, residual element abundances are low and suggest that crystal fractionation has not been an important process in the formation of these diorites. Additionally, it is generally recognised that crystal fractionation of primary tholeiitic basalts results in more evolved tholeiitic rather than calc-alkaline rocks (Green and Ringwood, 1968).

Experimental work by Green and Ringwood (1968) shows that andesitic (dioritic) melts can be obtained by the postulated two-stage process by wet melting of amphibolite ($pH_2O < pLoad$), at pressures of 9 - 10 kb, or dry melting of eclogite at pressures of

27 - 36 kb. The probable site for this process is along active Benioff zones resulting in island arc and continental margin volcanism. The age of the Lorne Basin diorites cannot be closely determined, but field relationships indicate they are post-Carboniferous and pre-Triassic, that is, approximately 270 - 230 my old. At this time the New England High was rising and the Newcastle geosyncline to the west and basins to the east were also tectonically active. Additionally, there was extensive block faulting in an area which includes the Lorne Basin. Andesitic and rhyolitic volcanism occurred throughout the Permian in the basins to the east of the New England High, and the New England Batholith formed during the Permian and Early Triassic as the result of intense plutonic activity in northeastern New South Wales and eastern Queensland (Brown et al., 1968). It has been proposed that the Permian tectonism and related igneous activity in eastern Australia resulted from the area being representative of an active continental margin or island arc (Leitch, 1974). This would provide a mechanism whereby the Lorne Basin diorites could have formed from partial melting of amphibolite or eclogite along a dipping Benioff zone.

APPENDIX I - LOCALITIES OF ANALYSED ROCKS FROM THE
COMBOYNE PLATEAU - LORNE BASIN AREA.

Four figure numbers are catalogue numbers for the rock collection at the School of Earth Sciences, Macquarie University. Grid references are given to one hundred feet and refer to the Hastings 1:250,000 Geological Sheet and the New South Wales Military Maps for Wingham, Comboyne, Cundle and Camden Haven. As far as possible the appropriate geographic name follows the grid reference. Finally the rock type of the specimen is given. This is not a complete list of rocks collected, but it includes all rocks analysed plus any additional rocks referred to in the text.

<u>Catalogue</u> <u>Number</u>	<u>Grid</u> <u>Reference</u>	<u>Location</u>	<u>Rock Type</u>
MU6498	549.9 079.9	Mt. Baldy	Hawaiite
6499	545.8 082.0	Mt. Killabank	Hawaiite
6500	566.6 088.3	Lorne Forest	Hawaiite
6501	555.6 092.4	Comboyne Plateau	Mafic Trachyte
6505	542.1 084.6	Comboyne Plateau	Benmoreite
6506	547.1 086.7	Comboyne Plateau	Comendite
6507	546.2 086.3	Comboyne Plateau	Benmoreite
6520	552.7 088.6	Mt. Bulli	Anorthoclase Trachyte
6562	555.5 084.4	Comboyne Plateau	Hawaiite
6563	552.2 084.5	Comboyne Plateau	Benmoreite
6564	552.2 084.6	Comboyne Plateau	Benmoreite
6565	548.8 094.1	Comboyne Plateau	Hawaiite
6566	548.2 088.9	Comboyne Plateau	Benmoreite
6567	545.7 091.7	Comboyne Plateau	Mugearite
6568	548.2 086.2	Comboyne Plateau	(?) Tholeiitic Andesite
6569	544.5 087.4	Comboyne Plateau	Comendite
6570	537.8 087.4	Comboyne Plateau	Hawaiite
6571	536.6 087.1	Comboyne Plateau	Hawaiite
6572	536.1 086.8	Comboyne Plateau	Hawaiite
6573	544.9 087.2	Comboyne Plateau	Comendite
6575	545.3 086.8	Comboyne Plateau	Benmoreite
6576	545.3 086.3	Comboyne Plateau	Benmoreite
6577	546.8 083.3	Comboyne Plateau	Tholeiitic Andesite
6578	547.2 083.4	Comboyne Plateau	Mugearite
6579	546.1 081.8	Comboyne Plateau	Comendite
6580	546.6 078.5	Wingham-Comboyne	Dacite
6581	546.6 078.5	Wingham-Comboyne	Dacite
6596	565.9 090.4	Lorne Forest	Hawaiite
6597	565.7 091.4	Lorne Forest	Hawaiite
6607	562.5 079.8	Hannon Vale	Mugearite
6608	562.5 079.8	Hannon Vale	Mugearite

<u>Catalogue Number</u>	<u>Grid Reference</u>	<u>Location</u>	<u>Rock Type</u>
MU6625	564.2 079.4	Hannon Vale	Mugearite
6627	550.0 085.5	Mt. Kophi	Anorthoclase Trachyte
6630	547.3 083.2	Mt. Gibraltar	Anorthoclase Trachyte
6631	548.2 084.2	Mt. Gibraltar	Anorthoclase Trachyte
6635	552.2 088.0	Mt. Bulli	Anorthoclase Trachyte
6636	553.1 078.9	Lansdowne Valley	Alkaline Pitchstone
6637	546.3 083.4	Mt. Killabahk	Comendite
6638	545.3 084.3	Mt. Killabahk	Comendite
6639	546.8 084.3	Mt. Killabahk	Comendite
6640	543.9 081.8	Mt. Killabahk	Comendite
6641	543.5 081.8	Mt. Killabahk	Comendite
6643	544.8 082.4	Mt. Killabahk	Comendite
6644	546.9 075.7	Mt. Goonook	Comendite
6645	550.0 077.3	Mt. Coxcomb	Comendite
6646	549.8 077.5	Mt. Coxcomb	Comendite
6647	555.8 077.8	Big Nellie	Comendite
6648	555.9 077.9	Big Nellie	Comendite
6650	556.5 076.5	Little Nellie	Comendite
6652	553.5 078.5	Mt. Oliver	Comendite
6653	540.4 064.7	Wingham	Comendite
6654	541.8 059.9	Wingham	Comendite
6658	552.0 084.8	Comboyne Plateau	Comendite
6659	556.4 078.5	Near Big Nellie	Comendite
6661	574.0 072.8	South Brother	Microgranite
6663	572.5 073.4	South Brother	Microgranite
6665	572.1 072.3	South Brother	Trachyte
6666	576.3 091.2	Lorne Forest	Diorite
6667	574.5 092.8	Lorne Forest	Diorite
6668	585.7 082.8	North Brother	Microdiorite xenolith
6669	585.7 082.8	North Brother	Microdiorite xenolith
6670	585.7 082.8	North Brother	Microdiorite xenolith
6671	585.7 082.8	North Brother	Microdiorite xenolith
6672	585.7 082.8	North Brother	Microgranite
6673	584.0 082.2	North Brother	Microgranite
6674	584.3 082.3	North Brother	Microgranite
6676	585.5 082.8	North Brother	Microgranite
6677	485.3 082.8	North Brother	Microgranite
6678	578.6 081.8	Ross Glen	Microgranite
6679	570.1 073.5	Juhle Mountain	Granodiorite
6680	570.0 074.0	Juhle Mountain	Granodiorite
6681	566.9 088.0	Lorne Forest	Granodiorite
6684	586.7 074.0	Diamond Head	Diorite
6686	586.8 073.6	Diamond Head	Dacite
6687	586.7 073.5	Diamond Head	Dacite
6689	586.5 073.3	Diamond Head	Basaltic Andesite
6690	571.2 071.9	Moorland	Rhyolitic Pitchstone
6692	579.3 085.3	Kew	Rhyolitic Pitchstone
6693	563.5 088.5	Hannon Vale	Rhyolite
6694	576.2 076.5	Middle Brother	Granite
6695	576.2 076.6	Middle Brother	Granodiorite
6696	574.8 077.8	Middle Brother	Granite

<u>Catalogue Number</u>	<u>Grid Reference</u>	<u>Location</u>	<u>Rock Type</u>
MU6697	576.8 077.8	Middle Brother	Granite
6698	576.5 077.5	Middle Brother	Granodiorite
6699	576.5 077.5	Middle Brother	Granodiorite
6700	574.2 077.2	Middle Brother	Granodiorite
6701	578.7 078.0	Middle Brother	Granodiorite
6702	578.7 078.0	Middle Brother	Granite
6703	563.5 079.2	Hannon Vale	Rhyolite
6704	574.2 092.8	Lorne Forest	Rhyolite
6705	570.0 074.0	Juhle Mountain	Granophyre
6706	576.2 076.4	Middle Brother	Granophyre

APPENDIX II - ANALYTICAL METHODS

Chemical Analyses

Material to be analysed was sawn into slabs, the slabs cleaned and reduced to 1.5 cm aggregate in a rock splitter, and then crushed in a tungsten carbide Siebtechnik Vibrating Disc Grinding Mill.

Whole rock analyses for the major oxides SiO_2 , TiO_2 , Al_2O_3 , total iron as Fe_2O_3 , MnO , CaO , MgO , K_2O and P_2O_5 and trace elements were determined by X-ray fluorescence following the methods of Norrish and Chappell (1967) on a Siemens S.R.S. spectrometer with a 4 kilowatt generator and a standard counting rack. A Cr anode tube was used in the determination of major oxides and V; a Au anode tube for Cr, Cu, Mn, Ni and Zn, and a Mo anode tube for Ga, Pb, Rb, Sr, Th, U, Y and Zr.

Matrix corrections for inter-element effects were made and backgrounds subtracted for major element analyses. Mass absorptions for Rb and Sr were measured, and those for Ga, Pb, Th, U, Y and Zr were calculated by factorising from the measured mass absorptions. Mass absorptions for Cr, Cu, Mn, Ni, V and Zn were derived from bulk rock analyses using published mass absorption tables; Count rates were adjusted where necessary for interfering spectral lines. For example, Y required major correction on the basis of the amount of Rb.

U.S. Geological Survey standard samples GSP1, AGV1, BCR1 and PCC1 were used as standards and values used for standardisation were kindly provided by Dr. B.W. Chappell. Comparisons between analyses obtained for the standard rocks during this study and those of Dr. Chappell are listed in Table A-2. Values used in machine calibration are underlined. Trace element values for the

TABLE A-2

ANALYSES OF U.S.G.S. STANDARD SAMPLES

Sample No:	GSP1	GSP1	AGV1	AGV1	BCR1	BCR1	PCC1	PCC1
Analysed At:	1	2	1	2	1	2	1	2
SiO ₂	68.16	67.40	59.10	59.16	54.10	54.38	42.04	41.68
TiO ₂	.66	.66	1.05	1.05	2.28	<u>2.25</u>	.004	.006
Al ₂ O ₃	14.96	15.00	17.10	<u>17.00</u>	13.31	13.51	.59	.62
Fe ₂ O ₃	4.22	4.23	6.59	6.66	13.14	<u>13.32</u>	8.22	8.23
MnO	.04	.04	.10	.10	.19	<u>.19</u>	.13	.13
MgO	1.04	.95	1.55	1.46	3.54	3.40	43.31	<u>43.24</u>
CaO	2.01	2.00	4.94	4.92	6.99	<u>7.01</u>	.55	.54
K ₂ O	5.38	<u>5.53</u>	2.88	2.92	1.70	1.71	.11	.003
P ₂ O ₅	.27	.28	.50	<u>.49</u>	.35	.37	-	.003

Sample No:	GSP1	GSP1	G2	G2	W1	W1	DT81	DT81
Analysed At:	1	2	1	2	1	2	1	2
Ba	1254	1200	-	-	165	165	-	-
Rb	-	-	-	170	22	22	-	-
Sr	-	-	475	478	177	187	-	-
Th	-	-	23	25	3	2.3	-	-
U	-	-	2	2	2	.57	-	-
V	43	53	24	35	260	245	8	10
Y	-	-	12	11	25	26	-	-
Zn	103	105	100	85	52	84	64	50
Zr	-	-	300	315	90	95	-	-
Cr	15	12	10	7.8	127	115	4040	4000
Cu	-	-	12	10	120	115	10	7
Ga	-	-	21	23	17	17	-	-
Ni	10	8	4	4	78	75	2385	2500
Pb	-	-	30	31	10	8	-	-

1. Macquarie University.

2. Dr. B.W. Chappell, Australian National University.

U.S. Geological Survey standard samples DTS1, G2, W1 and GSP1 determined in this study are also listed in Table A-2 and independent results of Dr. B.W. Chappell are listed for comparison purposes.

FeO was determined by the vanadate method of Wilson (1965), and Fe_2O_3 obtained by subtraction of FeO from total iron. Sodium determinations were carried out on an EEL photometer, using closely spaced artificial standards. Lithium sulphate was used to suppress inter-element interference. Checks on the accuracy of the results were regularly made using U.S. Geological Survey standard sample W1. Total water and carbon dioxide contents were measured by heating powdered samples in ceramic crucibles for one hour in a furnace at a temperature of 1300°C .

Mineral Analyses

Major element mineral analyses were determined following the methods of Reed and Ware (1973) at the Research School of Earth Sciences, Australian National University, using a Technisch Physische Dienst (TPD) microprobe fitted with an ORTEC Si (Li) detector and a Northern Scientific NS 710 multichannel analyser. Standards used for microprobe analyses were analysed natural silicate minerals.

X-Ray Diffraction

X-ray diffraction traces were run on a Siemens X-ray diffraction unit. $\text{Cu K}\alpha$ radiation and a Ni filter were used for all samples and SiO_2 was used as an internal standard.

APPENDIX III

Table A-3

MAJOR ELEMENT ANALYSES OF ALKALINE ROCKS

Rock Type:	Hawaiiite										Mugearite				
Sample No:	6571	6570	6572	6586	6565	6500	6499	6597	6562	6498	6607	6625	6608	6567	6578
SiO ₂	51.68	51.04	50.10	52.57	52.21	53.72	52.79	53.46	52.55	50.79	51.05	50.55	52.19	56.34	58.21
TiO ₂	2.30	2.70	2.62	2.14	2.14	2.14	2.25	1.93	2.23	2.34	2.24	2.24	2.05	1.66	1.34
Al ₂ O ₃	16.65	15.05	15.93	15.93	16.35	16.09	15.76	15.69	15.98	15.90	15.48	15.59	15.89	15.36	16.16
Fe ₂ O ₃	3.69	3.31	2.98	3.36	4.54	2.97	4.60	2.92	3.48	3.38	4.51	3.77	3.82	3.67	3.63
FeO	7.02	8.19	8.05	6.20	6.05	6.76	6.71	6.33	7.63	8.01	8.07	8.37	8.23	6.70	3.65
MnO	.17	.18	.16	.13	.17	.17	.18	.13	.18	.16	.19	.18	.20	.17	.17
MgO	4.75	3.85	3.72	4.12	3.82	3.89	3.89	4.03	3.81	4.14	2.38	2.67	2.32	2.18	2.11
CaO	7.82	7.19	7.26	6.30	6.57	6.09	6.71	5.84	6.59	6.78	5.45	5.63	5.14	4.55	4.55
Na ₂ O	4.04	4.10	4.29	4.34	4.27	4.47	4.41	4.32	4.50	4.60	4.63	4.65	4.82	4.93	4.70
K ₂ O	1.41	1.76	1.78	1.41	1.87	1.25	2.07	2.22	2.22	2.14	2.74	2.71	3.03	3.46	3.14
P ₂ O ₅	.65	.59	.57	.45	.60	.45	.69	.38	.87	1.15	1.27	1.28	1.21	.75	.30
Loss	.32	.87	2.16	1.26	1.19	1.19	.46	1.49	.40	.28	1.63	1.32	.53	.76	1.81
Total	100.70	99.83	99.62	98.21	99.79	99.79	100.52	98.74	100.49	99.67	99.64	98.96	99.43	100.53	99.77

Rock Type	Benmoreite							Trachyte		Anorthoclase Trachyte					
Sample No:	6575	6566	6564	6553	6505	6576	6507	6501	6565	6527	6531	6639	6635	6520	
SiO ₂	56.62	55.50	55.83	56.12	56.35	56.02	57.25	58.72	61.20	63.06	63.91	65.08	65.56	67.22	
TiO ₂	1.51	1.40	1.27	1.25	1.20	1.23	1.07	.61	1.04	.49	.46	.37	.30	.32	
Al ₂ O ₃	15.41	15.13	15.31	15.31	15.60	15.44	16.44	15.29	15.32	17.40	17.60	15.83	16.49	17.25	
Fe ₂ O ₃	4.50	3.05	6.02	6.86	7.10	4.85	4.08	5.42	2.87	3.17	2.62	2.37	1.56	1.53	
FeO	6.08	8.08	4.55	4.22	3.30	4.36	4.40	4.67	4.57	1.80	1.42	2.78	2.87	.78	
MnO	.16	.14	.23	.18	.17	.13	.17	.26	.34	.11	.07	.16	.10	.05	
MgO	1.68	1.42	1.11	1.10	1.07	1.15	1.48	.59	1.00	.47	.65	.16	.14	.10	
CaO	3.93	3.72	3.46	3.23	3.06	3.35	3.19	2.94	2.83	1.92	1.38	1.16	1.32	.69	
Na ₂ O	4.99	4.72	4.74	4.87	5.04	5.24	5.22	5.37	6.19	6.16	6.20	5.64	6.24	6.48	
K ₂ O	3.67	4.47	4.58	4.64	4.62	3.99	4.55	5.15	2.01	5.14	5.23	5.51	5.72	5.95	
P ₂ O ₅	.90	.87	.76	.77	.69	.56	.45	.31	.45	.15	.15	.10	.06	.04	
Loss	.76	1.29	1.91	1.07	1.28	1.13	1.47	.88	1.91	.32	1.31	.64	.22	.25	
Loss	100.11	99.79	99.82	99.62	99.48	99.45	99.90	100.21	99.73	100.19	99.11	99.80	100.58	100.66	

Table A-3 Cont'd

Rock Type:	Pitch	Comendite									
Sample No:	6636	6639	6637	6579	6638	6669	6506	6650	6659	6652	6640
SiO ₂	69.67	70.35	71.86	72.23	70.33	69.60	71.35	75.55	73.24	74.42	74.24
TiO ₂	.18	.20	.11	.13	.17	.26	.16	.13	.05	.10	.15
Al ₂ O ₃	12.77	14.10	13.54	11.22	14.75	14.34	12.10	11.46	13.26	11.85	10.93
Fe ₂ O ₃	.74	1.64	2.14	4.55	1.61	2.29	4.03	2.10	2.19	2.22	3.34
FeO	1.99	1.87	.91	1.26	1.90	2.12	1.55	1.35	.73	1.01	1.43
MnO	.08	.07	.05	.07	.05	.08	.06	.05	.08	.04	.05
MgO	.03	.10	.14	.13	.09	.07	.04	.05	.03	.08	.05
CaO	.47	.41	.08	.14	.16	.47	.26	.10	.08	.12	.20
Na ₂ O	4.87	6.18	4.21	4.49	5.56	6.15	5.78	4.80	5.81	4.96	4.98
K ₂ O	4.19	4.86	3.97	4.24	4.76	5.03	4.29	4.26	4.03	4.13	4.63
P ₂ O ₅	.02	.02	.02	.02	.02	.12	.02	.02	.01	.02	.02
Loss	4.59	.43	2.85	1.48	.54	.33	.74	.58	.55	.87	.36
Total	99.60	100.24	99.83	99.62	99.94	100.86	100.38	100.45	100.06	99.82	100.20

Rock Type:	Comendite										
Sample No:	6573	6643	6641	6645	6646	6647	6548	6644	6653	6664	6658
SiO ₂	71.33	70.25	73.79	73.42	73.60	74.49	72.21	71.47	74.72	73.74	70.33
TiO ₂	.15	.14	.15	.16	.17	.09	.09	.18	.09	.08	.14
Al ₂ O ₃	12.15	13.08	10.74	12.48	12.11	12.00	13.89	11.09	12.16	13.78	16.20
Fe ₂ O ₃	3.26	3.79	3.87	1.89	1.52	1.89	2.33	3.72	1.97	1.73	.98
FeO	1.62	.03	1.11	1.70	1.95	1.53	1.15	2.23	1.55	.79	.55
MnO	.07	.13	.06	.06	.08	.05	.05	.09	.06	.04	.04
MgO	.07	.29	.04	.15	.13	.04	.02	.07	-	.04	.05
CaO	.17	.08	.22	.39	.42	.09	.05	.34	.09	.09	.23
Na ₂ O	5.73	4.21	5.16	5.53	5.35	5.19	3.94	5.67	4.92	4.39	6.43
K ₂ O	4.36	4.51	4.24	4.42	4.34	4.33	3.50	4.47	4.32	4.06	5.18
P ₂ O ₅	.02	.02	.02	.05	.05	.01	.02	.01	.02	.01	.02
Loss	1.10	3.09	.51	.45	.34	.49	2.50	.40	.52	1.00	.31
Total	100.05	99.62	99.91	100.70	100.06	100.19	99.75	99.74	100.42	99.75	100.45

Table A-3

MAJOR ELEMENT ANALYSES OF SUB-ALKALINE ROCKS

Rock Type:	Tholeiitic Andesite		Basaltic Andesite	Diorite			Dacite				Dioritic Xenoliths				Granodiorite				
Sample No:	6577	6568	6569	6584	6565	6567	6586	6587	6581	6580	6670	6669	6671	6668	6702	6700	6698	6699	6681
SiO ₂	52.66	55.86	49.32	60.02	55.56	57.10	62.69	68.79	60.52	62.63	53.89	61.42	61.99	63.33	64.85	64.82	64.36	65.90	63.65
TiO ₂	2.04	1.46	2.41	1.00	1.12	.82	.79	.38	.89	.77	1.75	1.36	1.28	1.17	1.11	1.09	1.05	.85	1.08
Al ₂ O ₃	14.68	13.89	16.79	18.02	17.89	18.90	14.60	14.48	16.65	15.82	16.20	15.88	15.78	16.37	16.02	15.57	15.79	15.62	15.75
Fe ₂ O ₃	3.61	4.63	1.69	2.09	2.37	1.97	.69	.82	1.04	.90	2.20	1.27	.95	2.35	1.50	.54	1.46	1.28	1.38
FeO	7.58	7.84	10.00	3.99	4.34	3.61	7.95	2.95	3.75	2.95	6.82	5.28	5.20	3.36	4.26	4.17	3.86	3.71	3.13
MnO	.19	.23	.36	.11	.12	.14	.25	.38	.12	.09	.22	.16	.20	.14	.12	.10	.12	.13	.07
MgO	5.17	3.88	5.55	2.52	3.87	3.65	1.11	.40	3.45	2.26	3.55	2.15	1.86	1.52	1.69	1.21	1.63	1.30	1.81
CaO	8.34	5.21	7.29	5.28	7.93	8.13	.57	.23	3.18	3.63	5.19	4.76	3.07	2.40	4.18	3.08	4.02	3.27	2.75
Na ₂ O	3.29	5.33	3.11	3.89	3.74	3.66	3.83	4.53	5.64	4.18	5.92	4.86	5.47	5.09	4.47	4.86	4.70	4.94	4.77
K ₂ O	.96	.71	.35	1.30	1.14	.75	1.73	3.10	1.36	2.50	.67	1.97	2.72	2.82	2.40	2.55	2.39	2.75	2.26
P ₂ O ₅	.37	.71	.51	.25	.22	.19	.26	.11	.22	.20	.34	.39	.47	.49	.31	.23	.38	.31	.25
Loss	1.09	.66	2.29	1.41	.80	1.30	4.90	3.45	2.94	3.58	2.70	.47	.78	.72	.47	1.20	.67	.37	1.78
Total	99.98	100.41	99.67	99.88	100.37	100.12	99.83	99.68	99.76	99.51	99.45	99.97	99.77	99.81	101.28	99.42	100.43	100.44	98.68

Rock Type:	Granodiorite			Granite												Pitchstone		Rhyolite	
Sample No:	6679	6680	6695	6696	6697	6701	6694	6676	6674	6673	6677	6672	6678	6661	6653	6690	6692	6703	6693
SiO ₂	67.22	66.41	69.03	70.20	71.23	73.09	75.53	69.67	70.42	71.05	71.66	71.89	74.31	75.18	76.30	71.14	72.18	70.90	77.59
TiO ₂	.78	.68	.87	.47	.42	.37	.26	.33	.39	.41	.37	.34	.28	.09	.10	.26	.28	.56	.11
Al ₂ O ₃	15.69	15.35	15.58	15.05	15.38	14.16	14.11	14.06	14.79	14.89	14.57	14.41	14.12	13.03	12.98	12.96	13.07	15.47	12.03
Fe ₂ O ₃	2.14	1.85	1.24	.68	1.34	.44	.93	1.02	1.24	1.32	1.25	1.13	.58	1.11	1.41	.68	.63	.82	1.17
FeO	2.16	2.42	2.21	2.57	1.40	1.35	.33	1.54	1.73	1.71	1.51	1.41	1.23	.15	.17	1.11	1.07	.35	1.03
MnO	.07	.13	.12	.08	.06	.05	.04	.08	.09	.14	.09	.08	.06	.02	.01	.05	.06	.61	.07
MgO	.73	.90	.59	.66	.41	.33	.12	.44	.45	.51	.45	.36	.27	.10	.05	.24	.33	.43	.10
CaO	1.92	2.75	2.08	1.87	1.35	1.02	.69	1.28	1.60	1.33	1.12	1.15	.34	.08	.06	1.53	1.26	.26	.09
Na ₂ O	5.54	5.38	5.71	5.09	5.37	4.56	4.79	5.18	5.23	5.36	5.22	5.02	4.66	4.54	4.54	4.20	4.28	5.77	3.43
K ₂ O	2.87	2.88	2.51	3.45	2.92	4.34	3.56	3.68	3.58	3.55	3.58	3.86	4.56	5.06	4.62	2.36	3.44	3.90	4.76
P ₂ O ₅	.26	.22	.18	.09	.10	.08	.03	.08	.26	.10	.09	.08	.06	.01	.01	.04	.06	.08	.02
Loss	.51	.97	.25	.46	.55	.43	.30	1.83	.56	.34	.47	.59	.44	.42	.20	5.26	3.30	.72	.32
Total	99.89	99.94	100.14	100.67	100.53	100.21	100.89	99.19	100.34	100.77	100.38	100.32	100.91	99.79	100.45	99.83	99.96	99.87	100.77

APPENDIX IV

EXPERIMENTAL PROCEDURE

The natural hawaiite used in the experimental study (Chapter 5) was crushed, melted and quenched to glass. This glass was checked for homogeneity optically and by electron microprobe analysis of fragments, and then crushed to a fine powder. Iron content and oxidation state were checked by chemical analysis. The compositions of the experimental sample and the natural hawaiite (both corrected to 100%) are listed in Table 5.1.

Experimental techniques, using a piston-cylinder apparatus, follow those outlined by Green and Ringwood (1967) and Essene et al. (1970). In hydrous runs 2% and 5% by weight of H_2O was added using a micro-syringe. Some runs were conducted diluting the fluid phase with CO_2 by mixing 5% by weight of undried Analar oxalic acid with the rock powder. $Ag_{70}-Pd_{30}$ capsules were used in runs at temperatures of $1120^{\circ}C$ and less, and Pt capsules in runs at temperatures greater than $1120^{\circ}C$.

At the conclusion of a run the sample was studied optically and by X-ray diffraction. In a number of hydrous runs where crystal size was adequate, quantitative analyses of minerals present were obtained using a TPD electron microprobe following the method outlined by Reed and Ware (1973).

Plate 1a

Orthopyroxene megacryst in Lorne Forest hawaiite partially replaced by reaction rim of olivine and clinopyroxene.

MU6597, plane polarised light, X105.

Plate 1b

Orthopyroxene megacryst in Mt. Baldy hawaiite completely replaced by olivine and clinopyroxene, which in turn have undergone partial alteration to haematite.

MU6498, plane polarised light, X65.

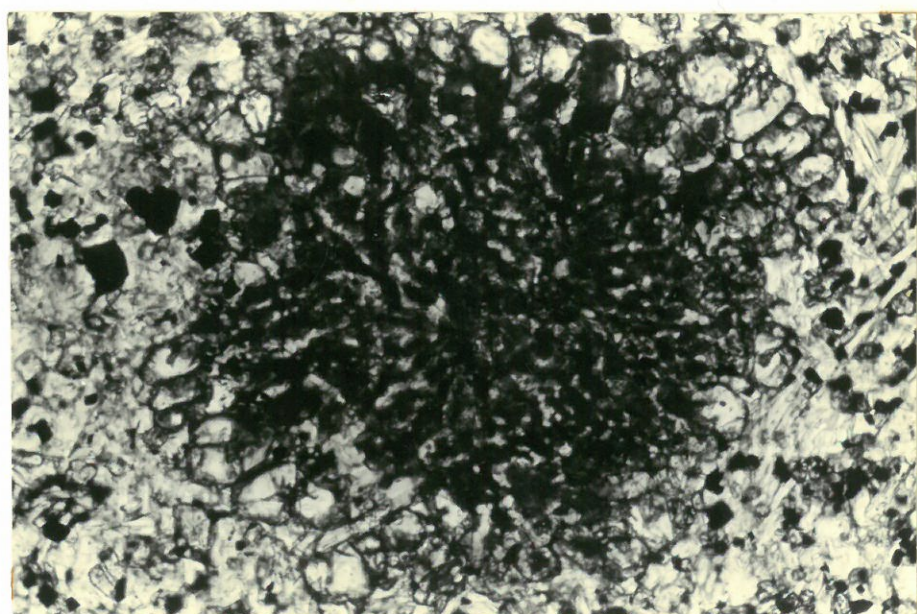
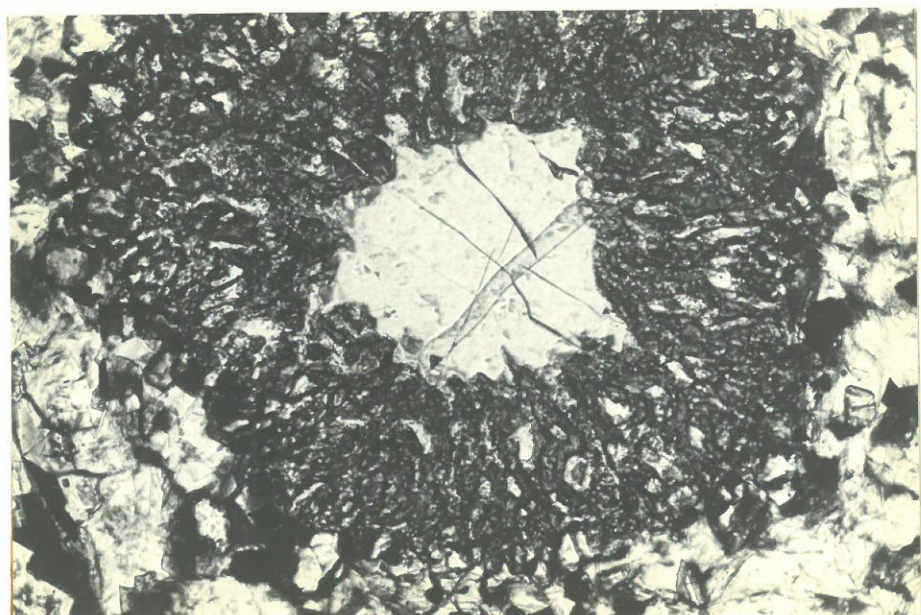


Plate 2a

Composite megacryst of olivine, orthopyroxene and plagioclase in Lorne Forest hawaiiite. Orthopyroxene has reaction rim consisting of olivine and clinopyroxene where in contact with host rock.

MU6597, crossed polars, X40.

Plate 2b

Crystal aggregate of plagioclase, clinopyroxene and orthopyroxene in Mount Baldy hawaiiite.

MU6498, crossed polars, X12.

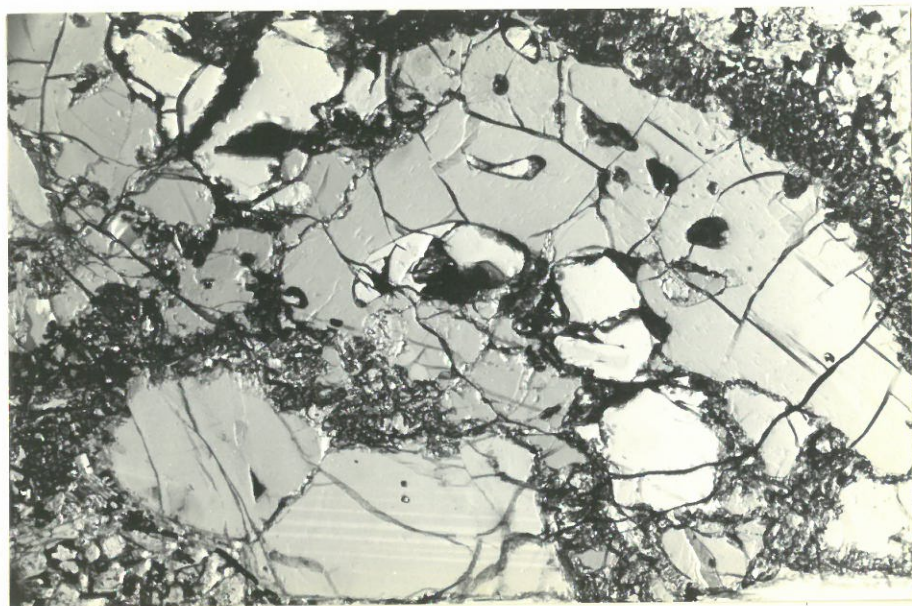


Plate 3a

Plagioclase poikilitically enclosing orthopyroxene
and clinopyroxene in Mount Baldy hawaiiite.

MU6498, crossed polars, X12

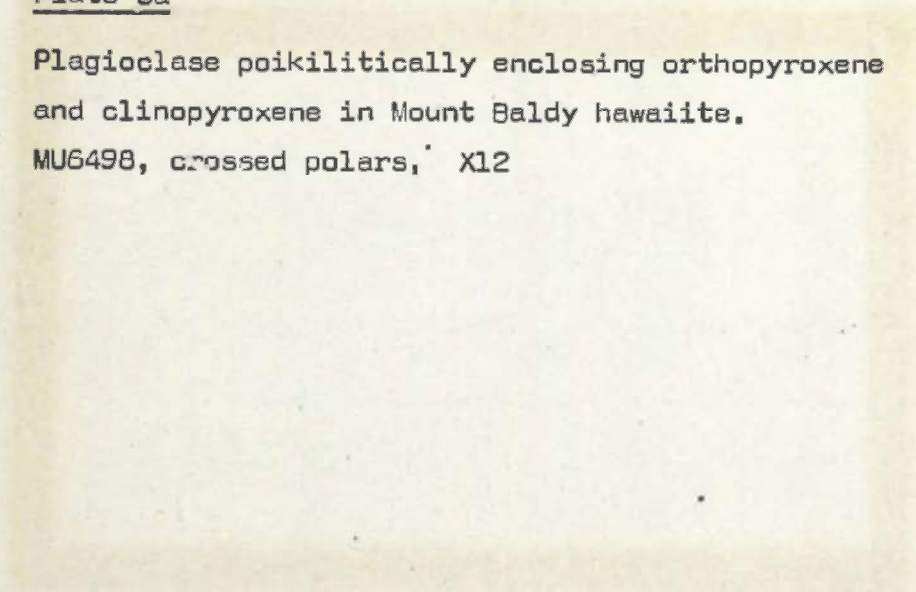
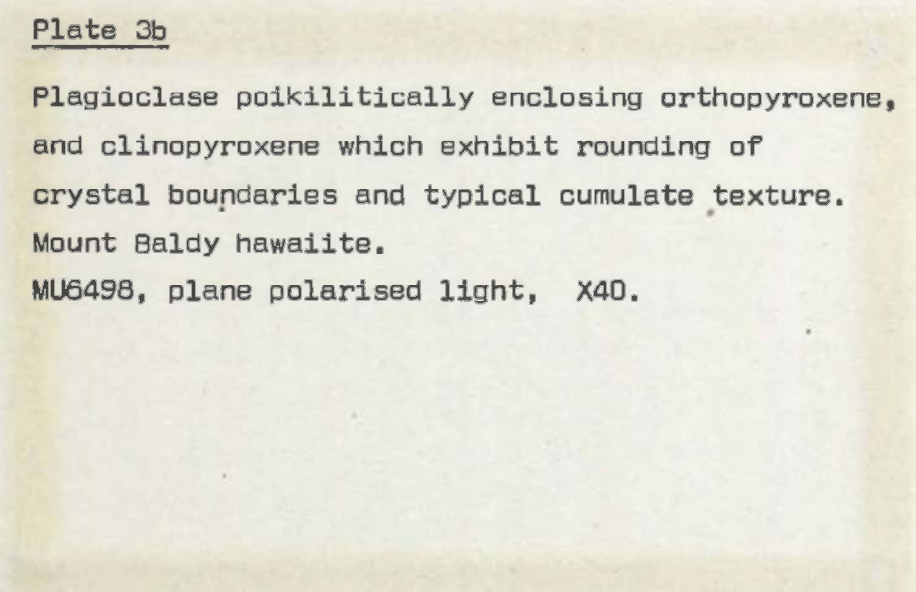


Plate 3b

Plagioclase poikilitically enclosing orthopyroxene,
and clinopyroxene which exhibit rounding of
crystal boundaries and typical cumulate texture.
Mount Baldy hawaiiite.

MU6498, plane polarised light, X40.



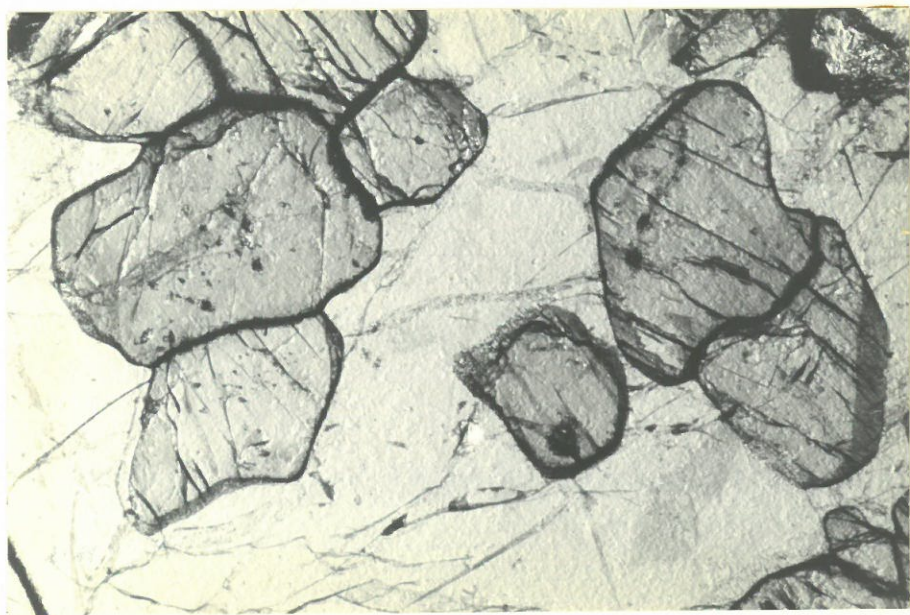
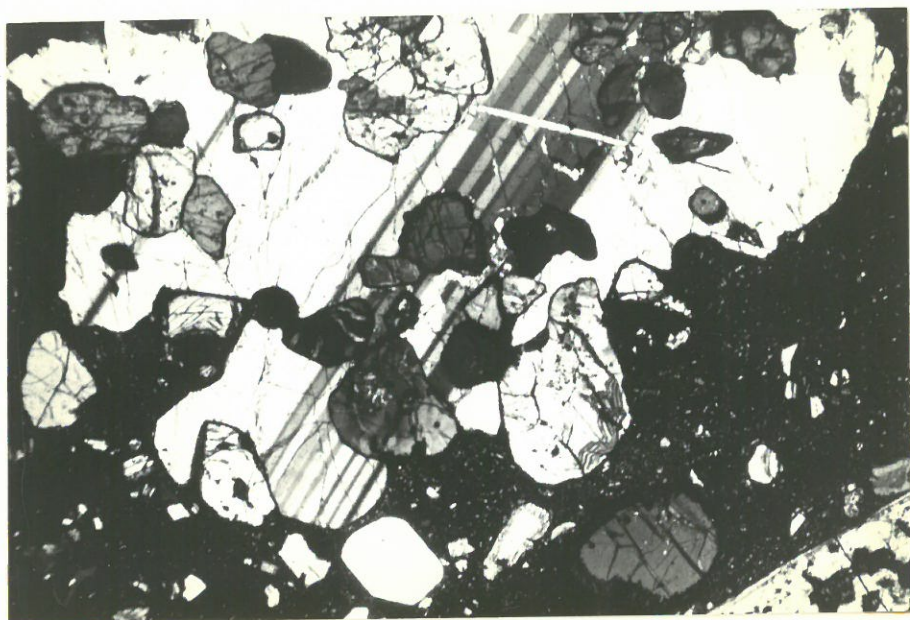


Plate 4a

Clinopyroxene megacryst in Lorne Forest hawaiite exhibiting sieved (fused) rim which in turn is enclosed within a reaction rim of olivine and clinopyroxene.

MU6597, plane polarised light, X105.

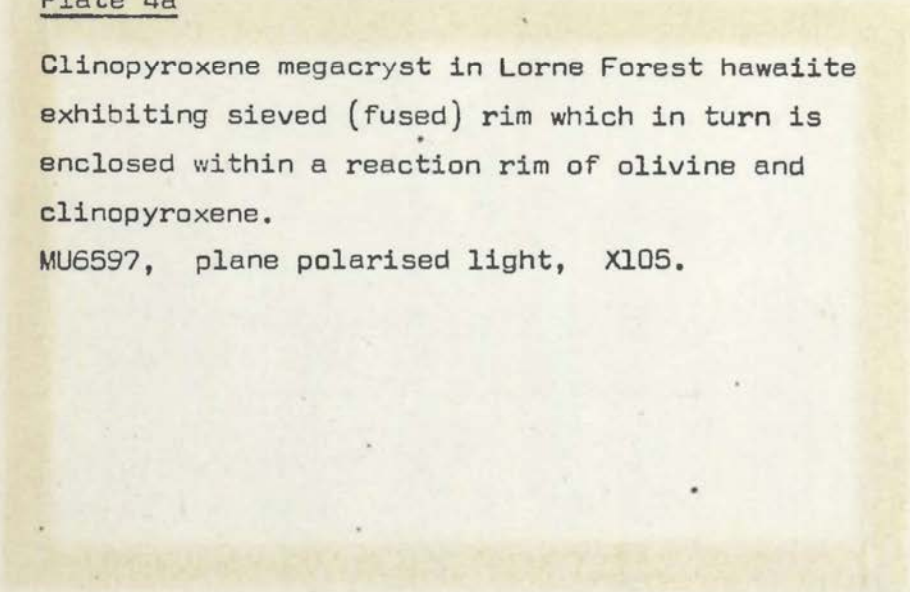
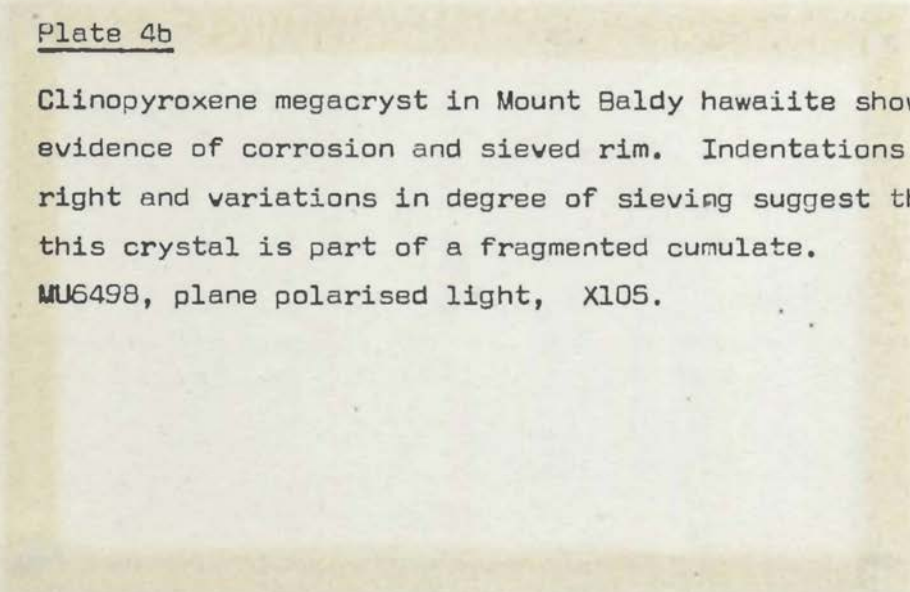


Plate 4b

Clinopyroxene megacryst in Mount Baldy hawaiite showing evidence of corrosion and sieved rim. Indentations to right and variations in degree of sieving suggest that this crystal is part of a fragmented cumulate.

MU6498, plane polarised light, X105.



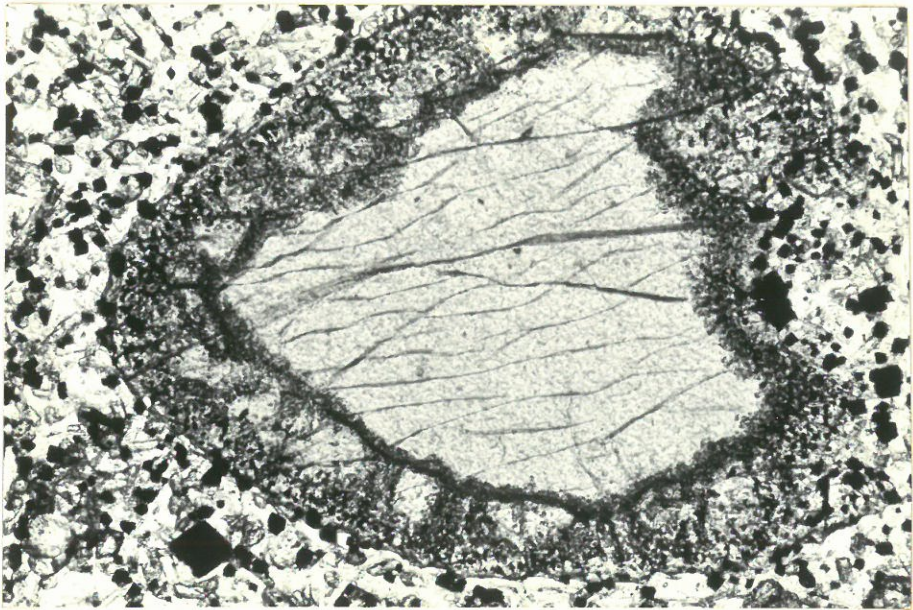
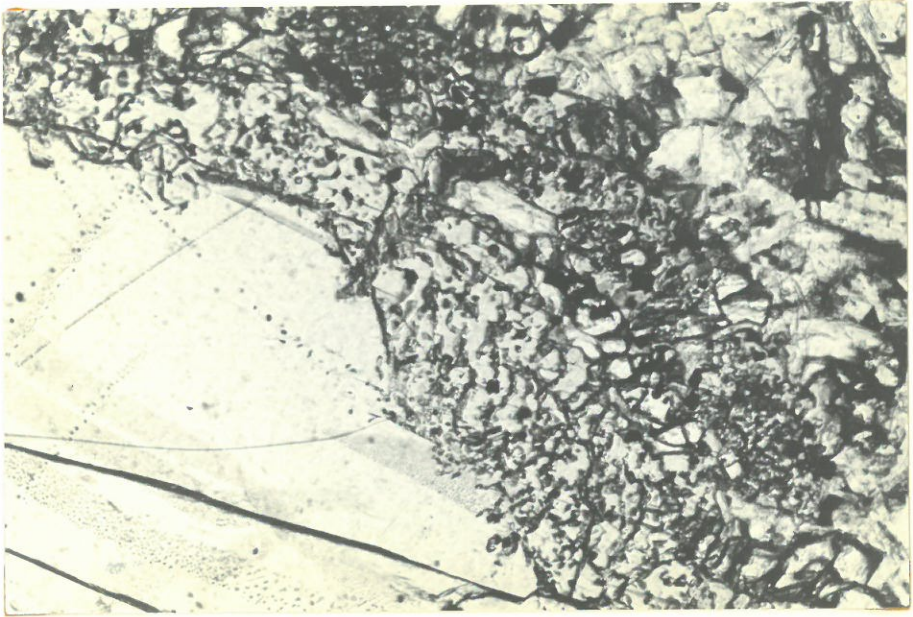


Plate 5a

Clinopyroxene megacryst with wide reaction rim of olivine and clinopyroxene in porphyritic mugearite. MU6578, crossed polars, X40.

Plate 5b

Clinopyroxene megacryst in Mount Baldy hawaiite with reaction rim of olivine and clinopyroxene. The megacryst also has inclusions of orthopyroxene which show reaction rims of olivine and clinopyroxene where in contact with the host rock. MU6498, crossed polars, X12.

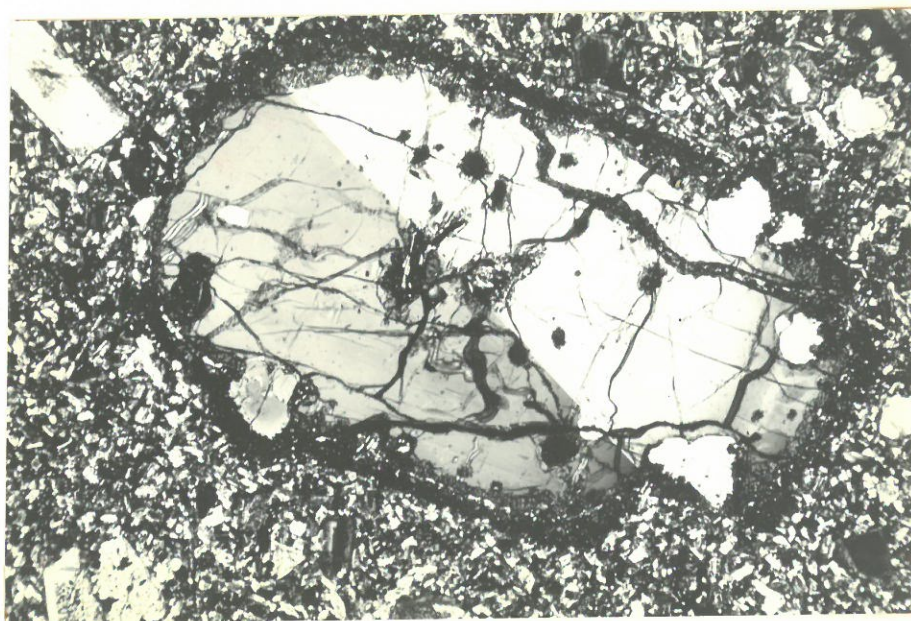
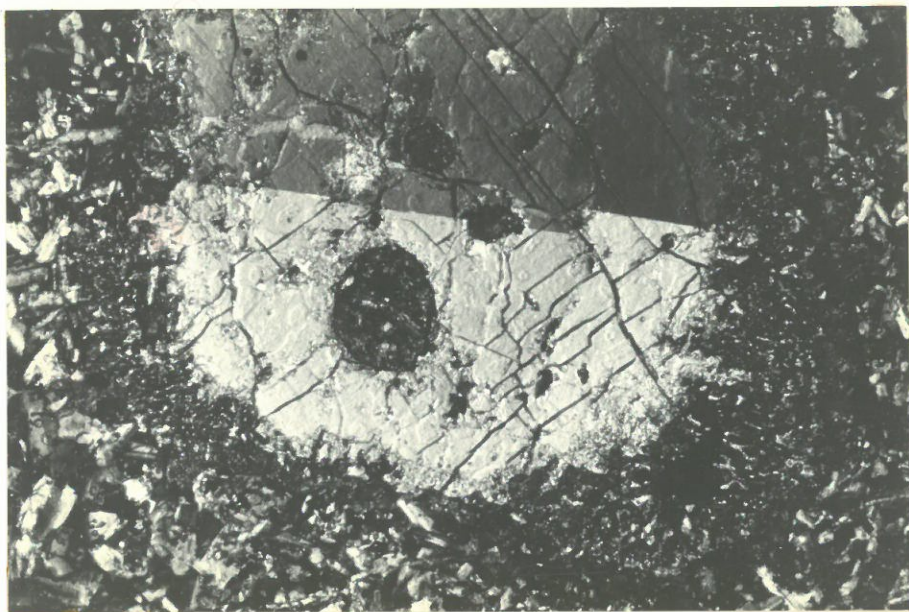


Plate 6a

Striated apatite crystal and strongly sieved clinopyroxene in Mount Baldy hawaiiite. The apatite has crystallographically aligned elongate rods of (?)Fe-Mg spinel.

MU6498, crossed polars, X105.

Plate 6b

Striated apatite and titanomagnetite in Mount Baldy hawaiiite. Both crystals show evidence of corrosion (rounding of grain boundaries).

MU6498, crossed polars, X40.

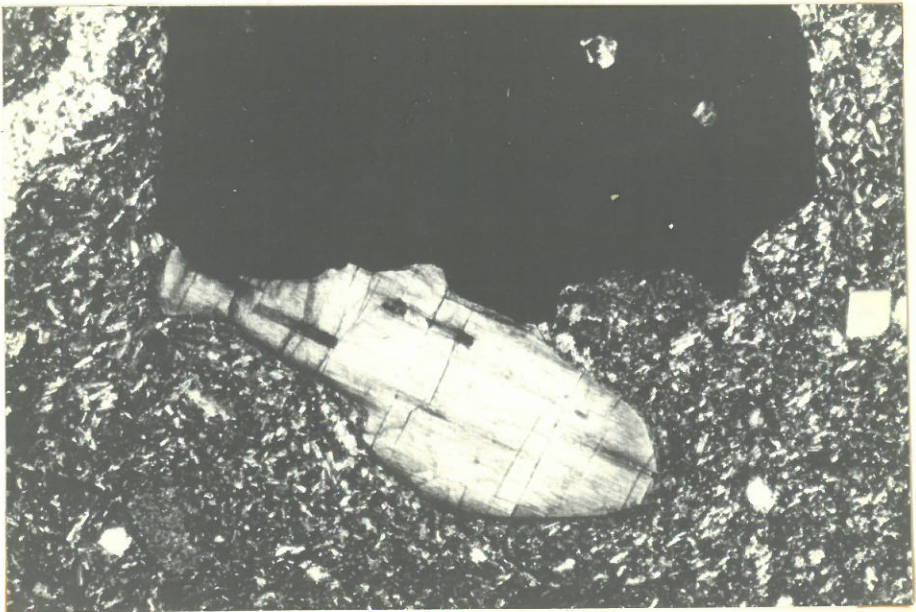
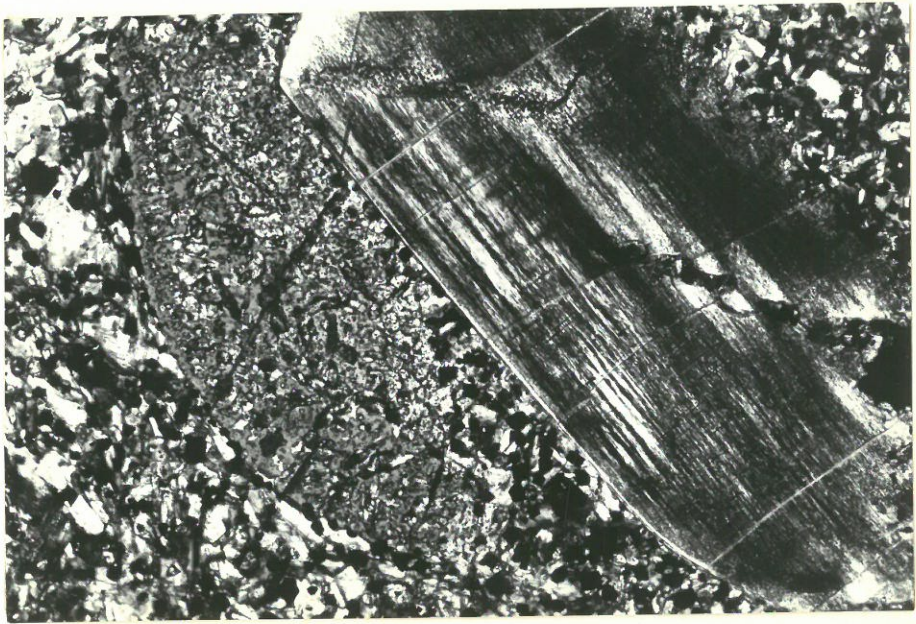


Plate 7a

Composite megacryst of orthopyroxene (with minor olivine) in the Comboyne tholeiitic andesite.

The orthopyroxene has a distinct rim of clinopyroxene.

MU6577, crossed polars, X40.

Plate 7b

Composite megacryst of orthopyroxene in Comboyne tholeiitic andesite, showing distinct rim of twinned clinopyroxene. Groundmass consists of laths of finely twinned plagioclase, augite, pigeonite, ilmenite and interstitial glass.

MU6577, crossed polars, X105.

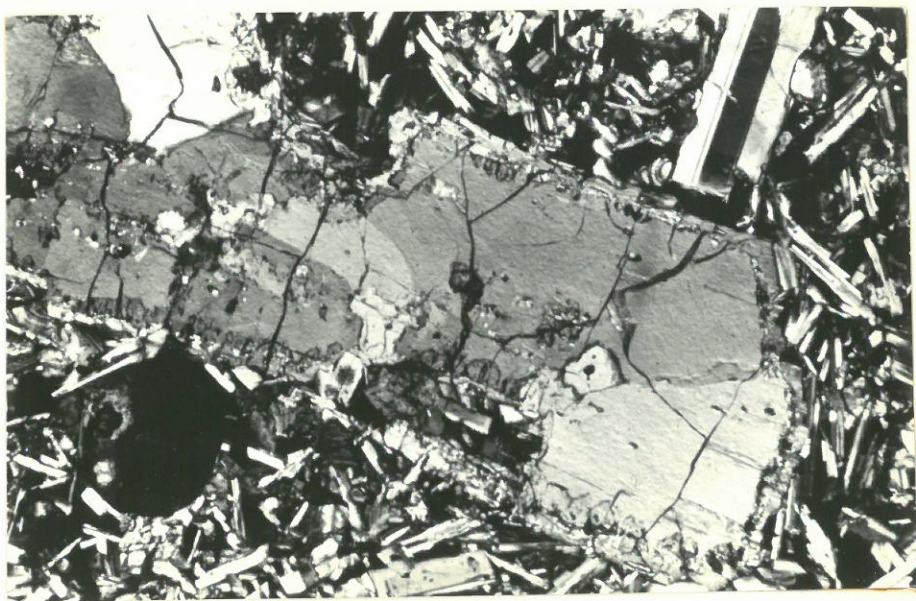


Plate 8a

Gabbroic aggregate in Comboyne tholeiitic andesite consisting of clinopyroxene, olivine, ilmenite and interstitial glass.

MU6577, crossed polars, X12.

Plate 8b

Gabbroic aggregate in tholeiitic andesite showing plagioclase with variable degrees of zoning, minor clinopyroxene, olivine and ilmenite.

MU6577, crossed polars, X30.

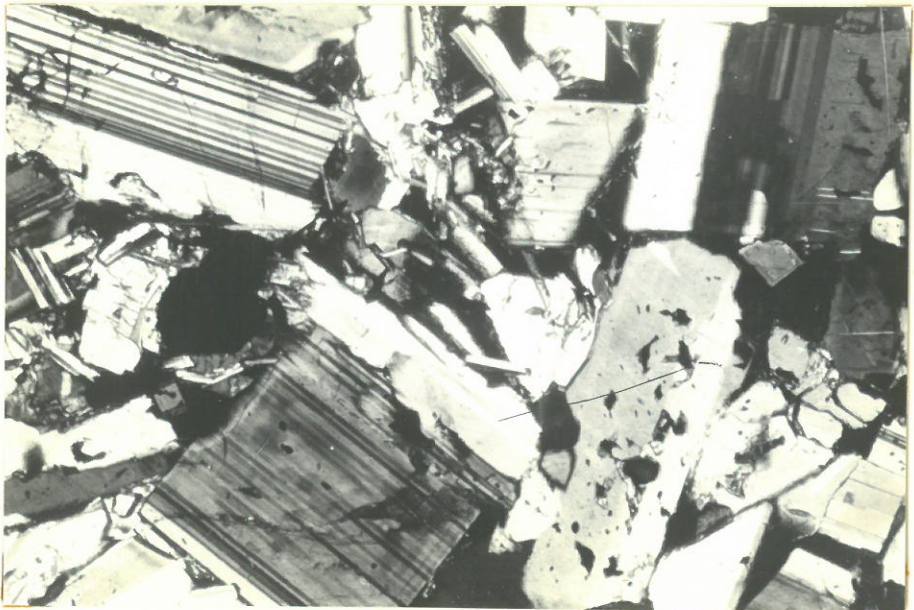


Plate 9a

Strongly corroded plagioclase (?) megacryst enclosing euhedral unsieved plagioclase crystal in Mount Baldy hawaiite.

MU6498, crossed polars, X12.

Plate 9b

Plagioclase (?) megacryst in Mount Baldy hawaiite showing corroded grain boundary and abundant inclusions of rounded globules of pyrrhotite.

MU6498, crossed polars, X75.

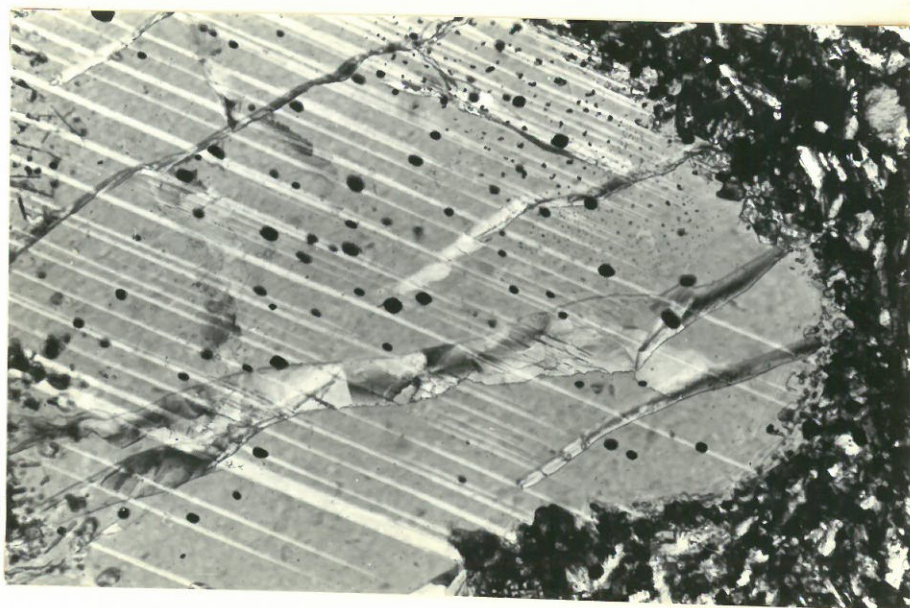
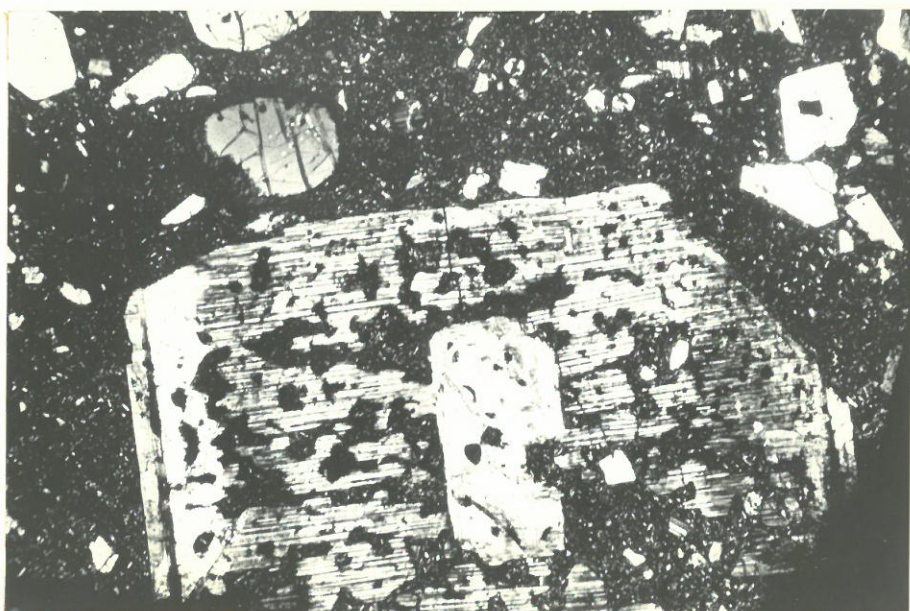


Plate 10a

Strongly corroded crystal of plagioclase with abundant inclusions of clinopyroxene, Fe-Ti oxides and clear apatite laths. Plagioclase crystal is rimmed by strongly zoned plagioclase. Mount Baldy hawaiiite.

MU6498, crossed polars, X105.

Plate 10b

Strongly corroded plagioclase and ilmenite crystals and a euhedral plagioclase phenocryst in the Mount Baldy hawaiiite.

MU6498, crossed polars, X40.

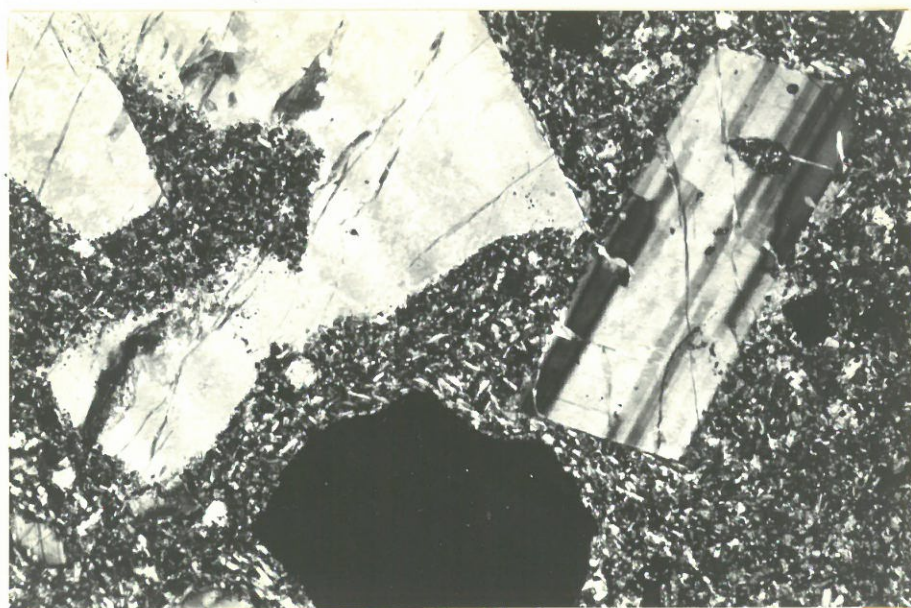
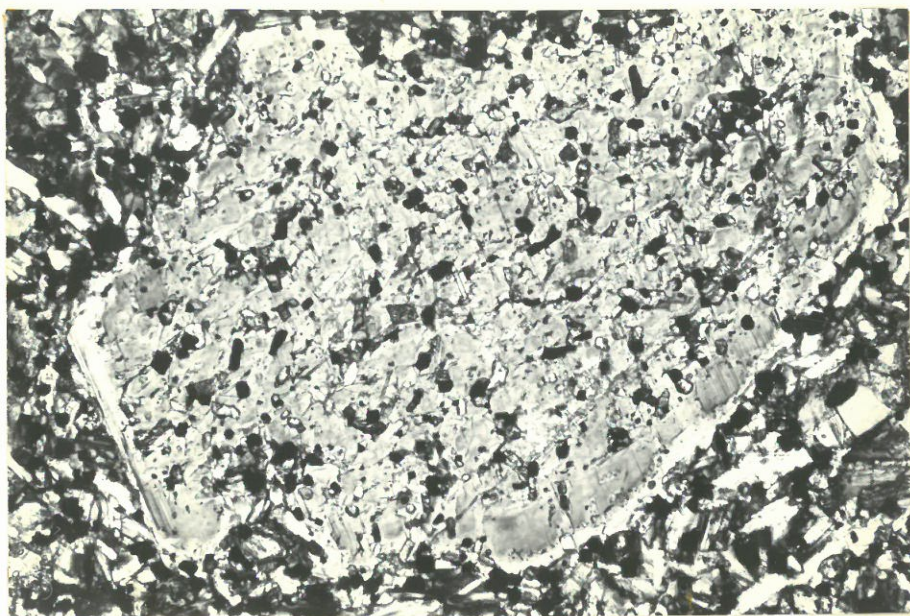


Plate 11a

Strongly sieved plagioclase (?) megacryst in the
Comboyne tholeiitic andesite.

MU6577, crossed polars, X65.

Plate 11b

Crystal aggregate in Kew rhyolitic pitchstone
consisting primarily of plagioclase, but with
minor orthopyroxene and clinopyroxene.

Devitrification of glass has occurred at
contact between glass and crystal phases.

MU6692, crossed polars, X40.

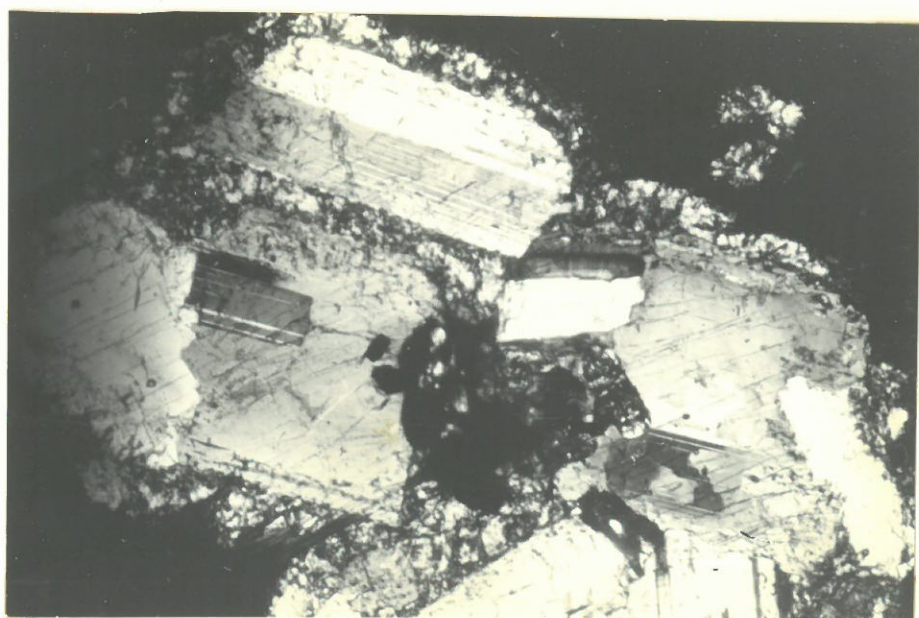


Plate 12a

Strongly sieved feldspar in Moorland rhyolitic pitchstone with microlites growing out into glass from crystal boundary.

MU6690, crossed polars, X105.

Plate 12b

Granophyric intergrowth interstitial to, and apparently replacing, feldspar phenocrysts in Juhle Mountain granophyre.

MU6705, crossed polars, X120.

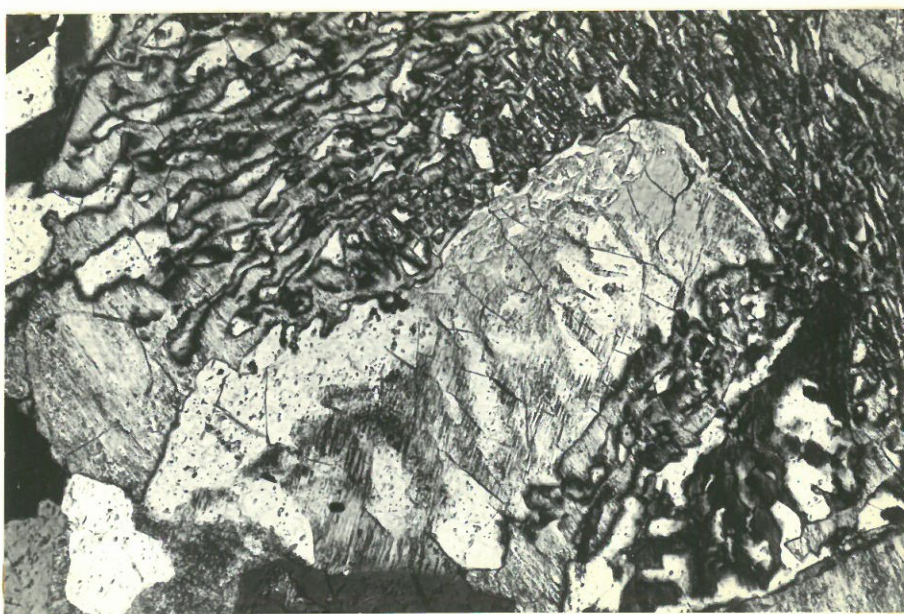


Plate 13a

Granophyric intergrowth apparently replacing large feldspar crystal in Juhle Mountain granophyre.

MU6705, crossed polars, X120.

Plate 13b

Coarse aggregate consisting of euhedral edenitic amphibole and apatite, and anhedral anorthoclase, in Comboyne benmoreite.

MU6564, plane polarised light, X105.

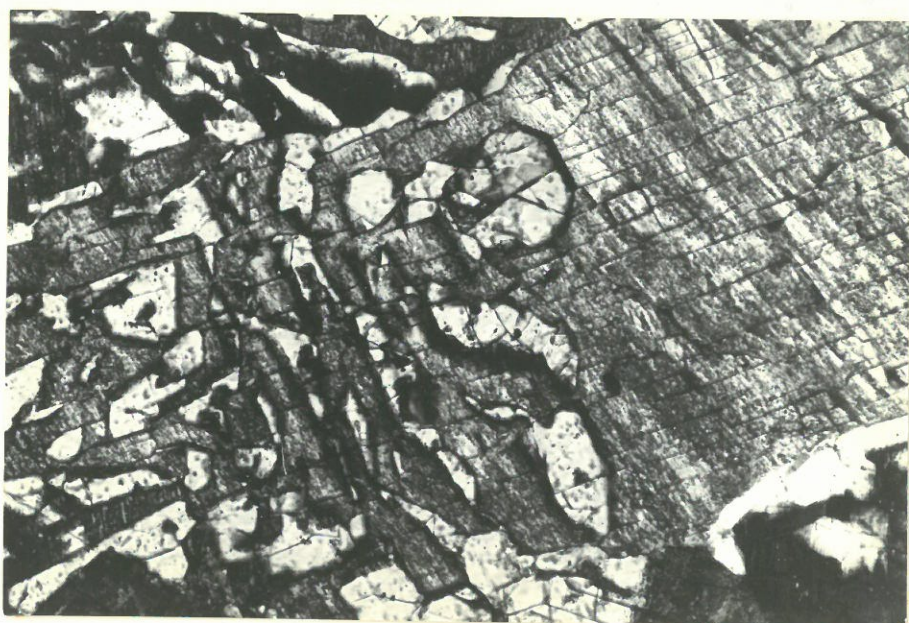


Plate 14a

Ilmenite megacryst partially rimmed by, and with lamellae of, titanomagnetite, in Mount Baldy hawaiiite. Note lamellae do not generally extend into rim.
MU6498, reflected light, X135.

Plate 14b

Curved interface between titanomagnetite and ilmenite in Fe-Ti oxide megacryst in Mount Baldy hawaiiite. Inclusions appear to be holding up grain boundary as ilmenite replaces titanomagnetite. Kingery (1960) notes that boundary energy decreases when a grain boundary reaches an inclusion and that this decrease in energy is proportional to the cross sectional area of the inclusion. The boundary energy must be increased again to pull away from the inclusion.
MU6498, reflected light, X725.

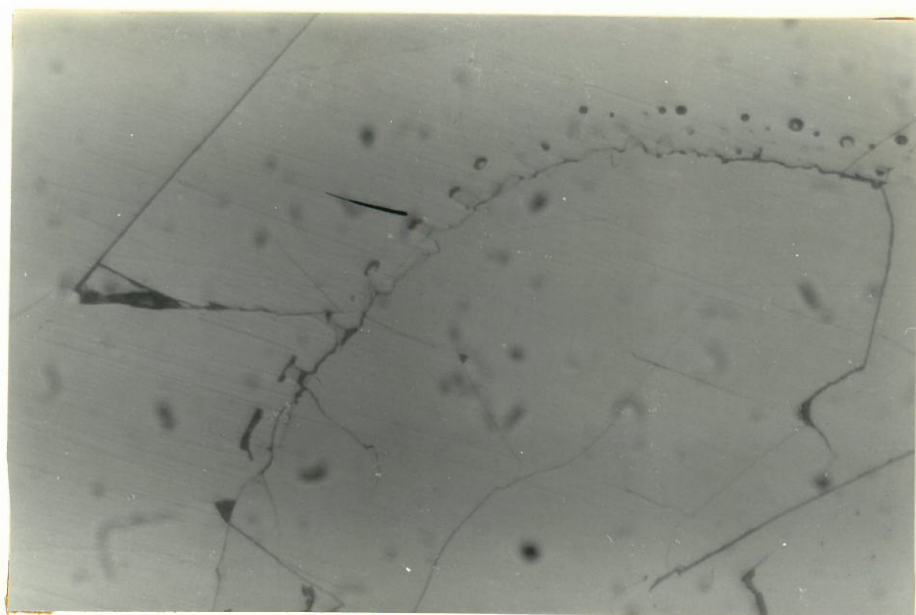
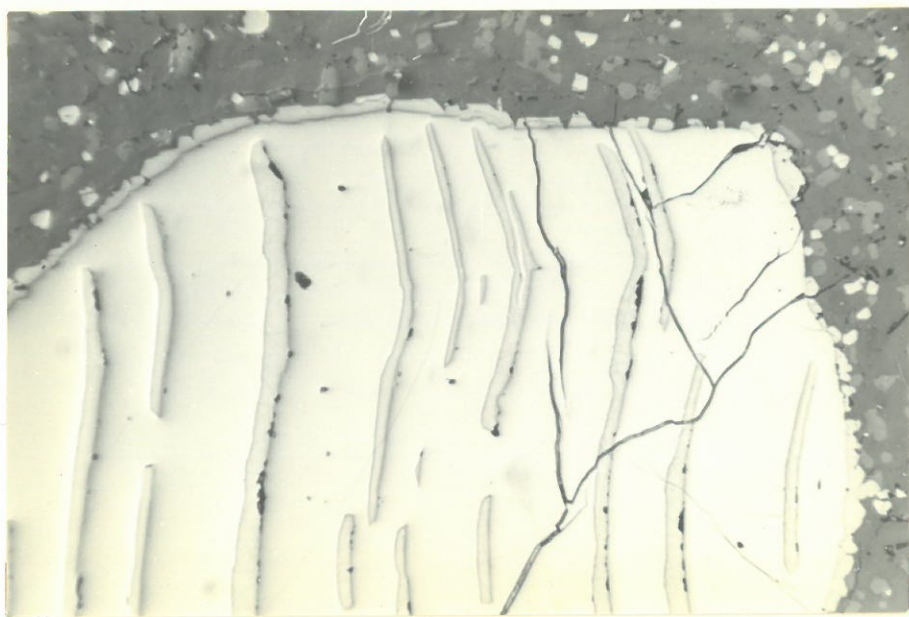


Plate 15a

Ilmenite in tholeiitic andesite showing dendritic outgrowths and partial rims of titanomagnetite indicating the titanomagnetite precipitated from the melt.

MU6577, reflected light, X1200.

Plate 15b

Fe-Ti oxide partially replaced by secondary minerals resulting in a residual lattice of ilmenite in a matrix of secondary minerals which include sphene. Diamond Head diorite.

MU6684, reflected light, X275.

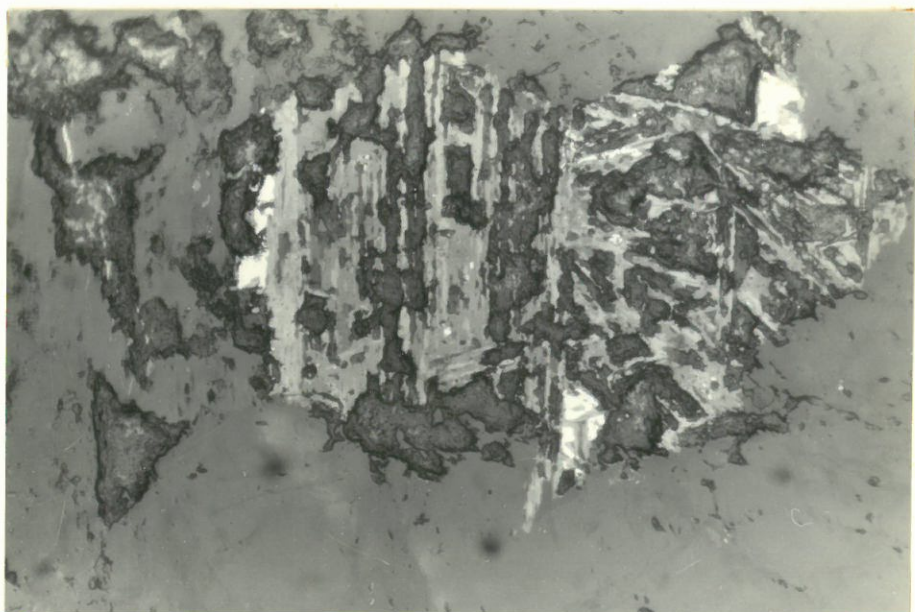
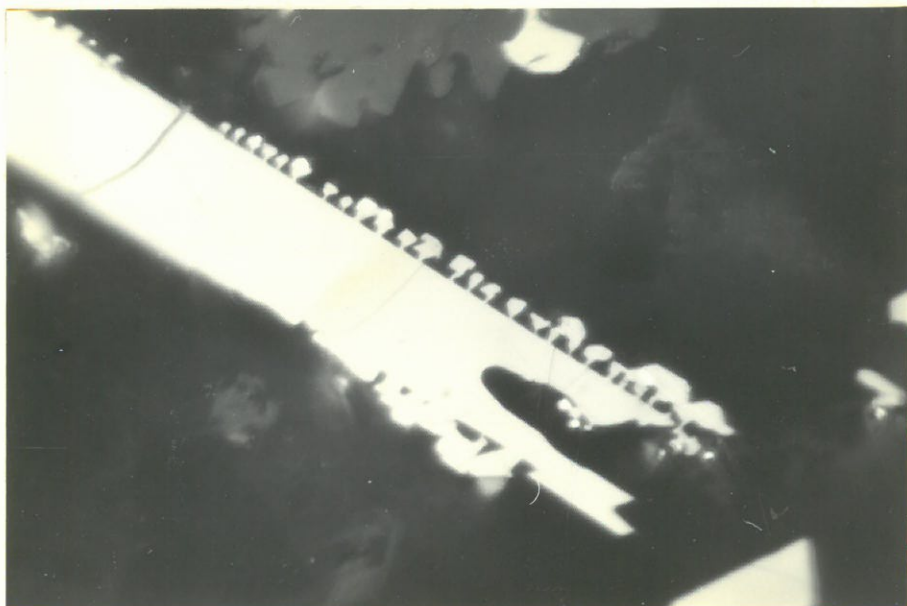


Plate 16a

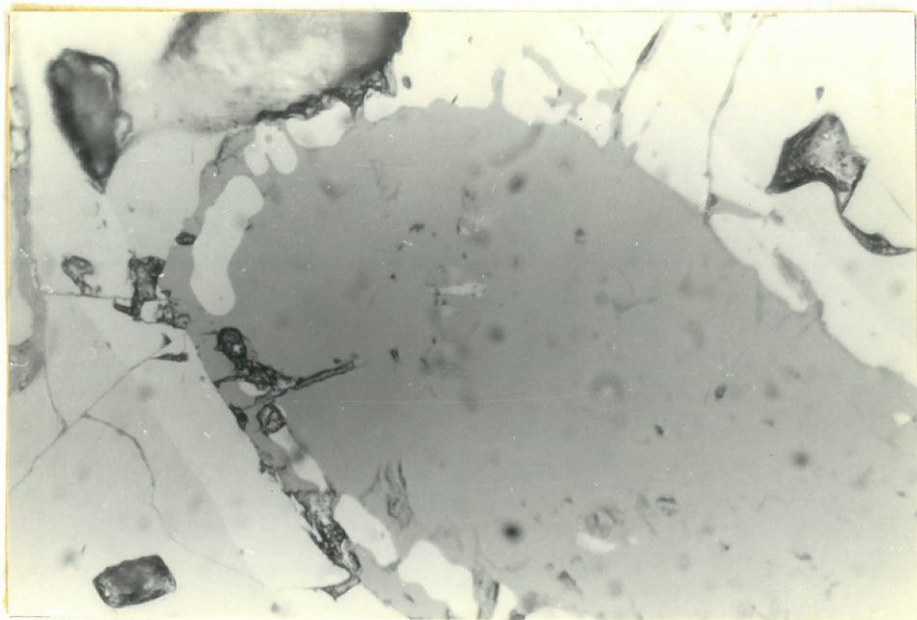
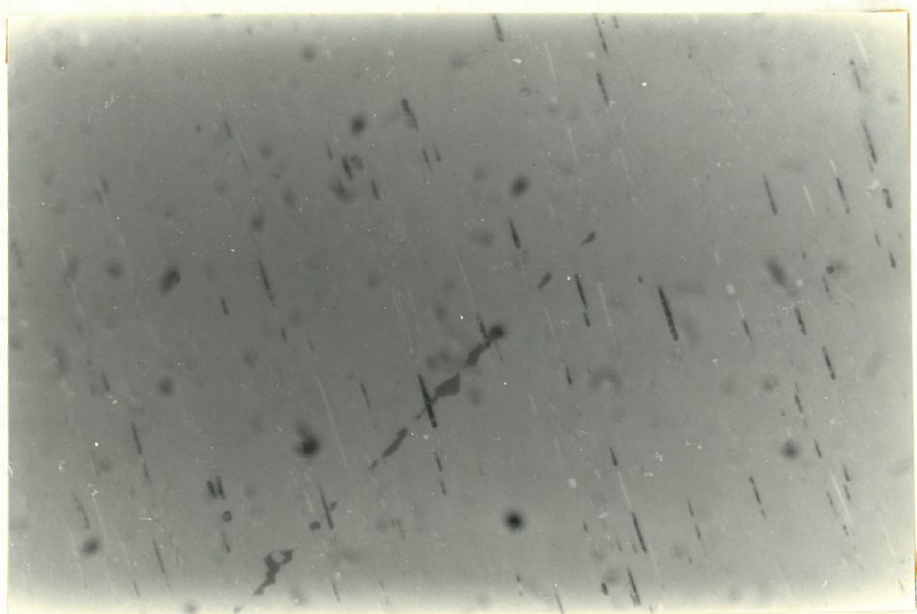
Apatite megacryst in Mount Baldy hawaiiite showing
crystallographically controlled (?)Fe-Mg spinel
(?)exsolution lamellae.

MU6498, reflected light, X725.

Plate 16b

Strongly corroded titanomagnetite megacryst rimmed
by ilmenite, which appears to have precipitated
from the melt. Mount Baldy hawaiiite.

MU6498, reflected light, X275.



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