

The Geochemistry, Mineralogy and
Igneous Petrology of some
Fracture Zone Basalts
from the Southern Mid-Atlantic Ridge

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degree
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A B S T R A C T

The compositional variation in a suite of oceanic basalts dredged from a fracture zone transecting the Mid-Atlantic Ridge at 54°S is quantitatively evaluated in terms of fractional crystallization and crystal accumulation models. Bulk rock chemistry and petrography are used to classify the samples into four distinct groups; an evolved sparsely plagioclase phyric suite (Group A), a less evolved sparsely plagioclase phyric suite (Group B), an olivine phyric suite (Group C) and a highly plagioclase phyric suite (Group D).

The Group A basalts are characterised by their evolved nature ($Mg\# = 52.8-59.4$) and by relatively high TiO_2 (1.89 - 2.15 wt%) and FeO (9.77 - 10.59 wt%) contents when compared to the Group B basalts. These are characterised by TiO_2 contents of between 1.08 and 1.63 wt% and by elevated Al_2O_3 (15.85 - 16.76 wt%) and CaO (11.43 - 12.27 wt%) contents. The Group C basalts are uniquely identified by their olivine phyric nature and consequently high MgO concentrations (8.39 - 15.54 wt%) and by having high to extremely high Cr and Ni contents (502 - 1079 ppm, 173 - 653 ppm respectively). The Group D basalt is identified by its plagioclase phyric nature, its high $Mg\#$ (70.0) and by the extremely high Al_2O_3 (19.35 wt%) and CaO (13.17 wt%) concentrations.

Whole rock major and trace element variations in conjunction with a mineralogical study of the major mineral phases present (olivine; plagioclase; spinel) are used to place constraints on the genetic relationships between the different basalt groups.

Trace element ratios and profiles, the distribution of NiO in olivines and Ba-Zr relationships are used to identify 4 distinct parental magmas.

Consideration of Fe-Mg partitioning between olivine and the host rock is used to identify possible liquid compositions. Such considerations indicate that the olivine phyric Group C sample represent olivine (+ minor Cr-spinel) enriched magmas, derived from a less MgO rich parental magma. Fe-Mg partitioning into plagioclase is used to infer that the highly plagioclase phyric Group D sample represents a partial plagioclase cumulate.

Least squares mixing calculations indicate that compositional variations in the Group A basalts cannot be explained by the simple closed system fractional crystallization of olivine and plagioclase from the least evolved sample. Group B compositional variation may be explained by up to 6.58% olivine, 22.20% plagioclase and 9.42% clinopyroxene fractionation. The Group C basalts may all be related to one of the least magnesian Group C samples by the accumulation of up to 17.05% olivine plus minor Cr-spinel.

It is proposed that basalts from this 54°S fracture zone evolved to their present compositions in a series of small, ephemeral crustal magma chambers. Unlike many basalts from this region of the Southern Mid-Atlantic Ridge (SMAR), their source regions show no evidence of geochemical enrichment, and these basalts have the trace element characteristics of normal depleted MORB. Within the model proposed by le Roex et al (1987) for the evolutionary history of this region of the SMAR, these basalts

are held to represent normal depleted MORB, unaffected by the either the Bouvet hot spot to the south or the Shona hotspot to the northwest.

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A C K N O W L E D G E M E N T S

The final product of a 2 1/2 year Masters degree - this thesis owes a vast amount to a great many people. Firstly, Elisabeth Brown must be commended for putting up with a year of long distance complaining. It is to her that I dedicate this work.

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Chapter One

I N T R O D U C T I O N

The African, South American and Antarctic plates meet at a triple junction near Bouvet Island ($54^{\circ}25'S$ $3^{\circ}25'E$) in the Southern Oceans (Figure 1). The Mid-Atlantic, American-Antarctic and Southwest Indian ridges radiate from this triple junction, with the Mid-Atlantic Ridge separating the South American and African plates. The spreading ridges are displaced in places by deep transform faults or fracture zones, and the basalts of this study were dredged from the southern wall of a fracture zone at $54^{\circ}S$, just north of the Bouvet triple junction.

The study of ocean floor basalts has developed concomitantly with the study of plate tectonics since the development of the theory of plate tectonics by Morgan and others since 1968 (Morgan, 1968; Le Pichon, 1968; Le Pichon et al, 1973; Wyllie, 1971; Dietz, 1961; Heirtzler & Bryan, 1975; Hess, 1965). The rise in interest in ocean floor volcanism is a result of the realisation that ocean floor basalts are integrally related to plate formation and movement at diverging plate boundaries. Study of the composition and evolutionary history of basalts increases the understanding of the nature of the sub-oceanic mantle, and processes operating within it.

This study was undertaken under the auspices of the Southern Oceans Lithosphere Program (SOLP), a collaborative program involving the Geology and Geochemistry departments of the University of Cape Town, Woods Hole Oceanographic Institution and the Bernard Price Institute for Geophysical Research.

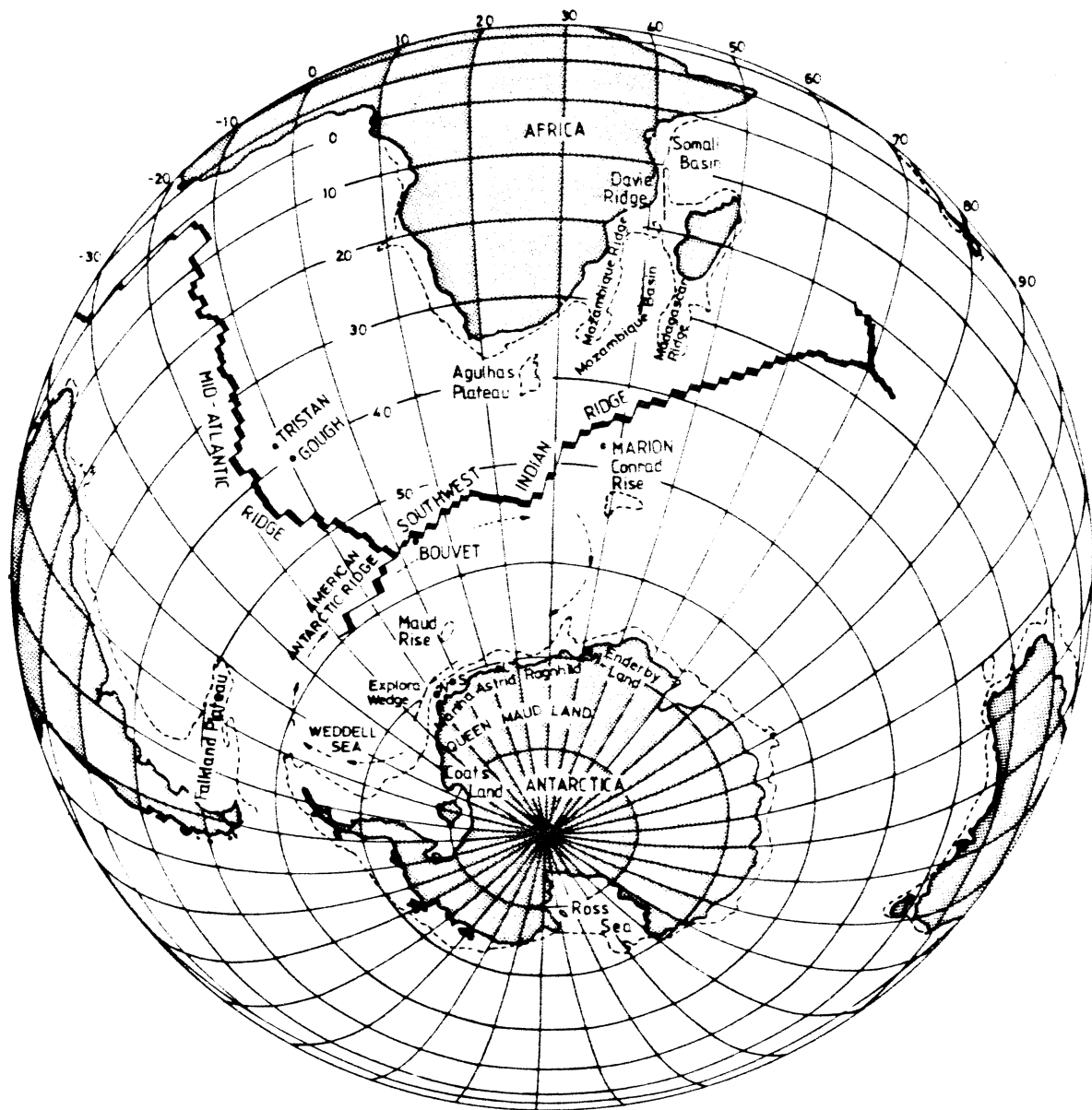


Figure 1. The Southern Oceans showing the Bouvet Triple Junction at which the Mid-Atlantic Ridge, The Southwest Indian Ridge and the American-Antarctic Ridge meet.

Primary aims of SOLP encompass the geochemical and geophysical investigation of mid-ocean ridge systems, their associated fracture zones and the oceanic islands and aseismic ridges of the Southern Oceans in order to study the evolution of the Southern Ocean Lithosphere.

Samples from 6 cruises (Islas Orcadas, Atlantis II, Vulcan 5, Agulhas 22, Agulhas 32 and Protea 5; see Figure 2) have formed the basis of SOLP. The study reported here concentrates on samples obtained from a single dredge (no. 3) of the Agulhas 32 cruise from the southern Mid-Atlantic Ridge (SMAR).

Work on SOLP is nearing completion and has resulted in a number of publications (i.a. le Roex, 1985; le Roex & Dick, 1981; le Roex et al, 1982, 1983, 1985, 1987; Dick et al, 1982; Lawer et al, 1982; Shimizu & le Roex, 1982). However, a number of important points still need to be addressed. For example, a suite of highly olivine phyric basalts with elevated MgO contents (10 to 15% MgO) was amongst those dredged on cruise 32 of the S.A. Agulhas. Such olivine rich lavas are comparatively rare on the ocean floor and carry important implications with regard to the occurrence and petrogenesis of picritic primary magmas in the oceanic environment. There is still contention as to whether average MORB (7 to 9% MgO) is derived from a high pressure picritic precursor by fractional crystallization, as proposed by O'Hara (1968) and others (Green et al, 1979; Jacques and Green, 1980; Stolper, 1980; Elthon, 1986; Elthon and Scarfe, 1980, 1984), or whether least evolved MORB (10 - 12% MgO) MORB can be produced as a primary melt by partial melting of sub-oceanic mantle at low pressure (8 - 11 Kbar) (e.g. Presnall

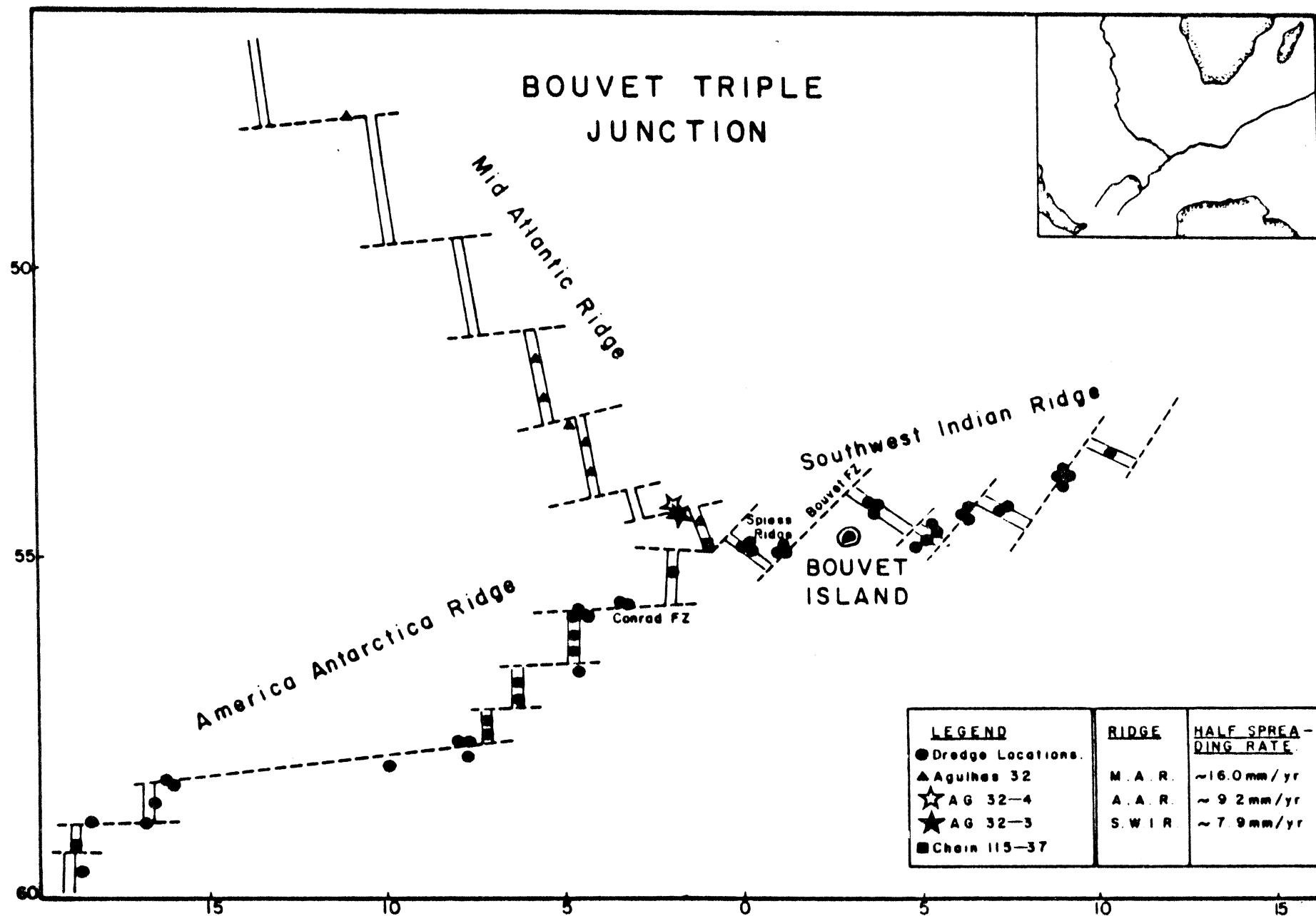


Figure 2. Sampling sites in the vicinity of the Bouvet Triple Junction. Agulhas 32 dredged the Southern Mid-Atlantic Ridge, Vulcan 5 the American-Antarctic Ridge and Agulhas 22 the Southwest Indian Ridge. Half spreading rates from Sclater et al (1976).

et al, 1979; Green and Ringwood, 1967; Kushiro, 1973; Fuji and Bougault, 1983; Takahashi and Kushiro, 1983; Presnall and Hoover, 1984, 1986).

A detailed study of the mineralogy and geochemistry of these high MgO basalts, in order to determine which of the two hypotheses presented above is most applicable for MORB from this region of the SMAR, forms part of this thesis. In addition, this study examines samples obtained from the same dredge haul (Agulhas 32, dredge 3) as the picritic suite, but more similar to average MORB (Bryan, 1983).

Early work on transform fault basalts (Melson and Thompson, 1971; Thompson and Melson, 1972; Melson et al, 1972) noted the more common occurrence of alkali basalts in these tectonic settings. Since then many other differences have been noted between fracture zone basalts and ridge basalts. A number of authors have pointed out that fracture zone basalts tend to be more evolved or differentiated than ridge axis basalts (Hekinian & Thompson, 1976; le Roex et al, 1983; Langmuir & Bender, 1984) and many postulated that this results from cooler thermal environments at the fracture zone (Schilling & Sigurdsson, 1979; Christie & Sinton, 1981; Sinton et al, 1983; le Roex et al, 1983).

A transform fault effect (Langmuir & Bender, 1984; le Roex et al, 1983) with decreased melting occurring in and near fracture zones, has been invoked by some authors to explain these compositional differences and discontinuities across the fracture zones (e.g. Langmuir & Bender, 1984; Schilling et al,

1983; Bougault & Treuil, 1980; Schilling et al, 1982, Machado et al, 1982). It must be noted, however, that a large number of well studied fracture zones show no evidence of a transform fault effect, with no alkali basalts or chemical discontinuities on either side of the fracture zone (Bryan et al, 1981; Bender et al, 1984; Campsie et al 1973; Hekinian & Aumento, 1973).

The principle aims of this project are (1) the detailed study of the amounts and effects of crystal fractionation and accumulation and partial melting in a spatially restricted suite of lavas and (2) the comparison of these fracture zone basalts with those from the adjacent ridges, to see whether any compositional differences exist, or whether a transform fault effect occurs in this vicinity.

G E O L O G I C A L S E T T I N G A N D S A M P L I N G

The samples used in this study were obtained during cruise 32 of the S.A. Agulhas, which took place in September, 1983. They were dredged from the steep south wall of an east-west trending fracture zone cutting the Mid-Atlantic Ridge at 54°S (Figure 2). This currently active (Sclater et al, 1976), dextrally shearing fracture zone is ~20 Km wide and has a maximum depth of 2800 m. A half spreading rate of 16.0 mm per year has been recorded for the Mid-Atlantic Ridge in this vicinity (Sclater et al, 1976).

The ship was positioned at 54°7.37'S, 1°57.6'W using a combination of bathymetry, magnetic stratigraphy and satellite navigation and 122 samples (having a total mass of 126 Kg) of moderately weathered porphyritic and aphyric pillow basalts were recovered from the wall of the transform between 2500m and 2200m depth. The dredge position was monitored with an O.R.E. pinger.

Preliminary hand specimen description and labelling was completed on board ship. In order to select a range of samples for detailed analysis, a subset covering all petrographic types was chosen. These samples were slabbed and a 23 mm disk cut from each was directly analysed by X-ray fluorescence spectroscopy (XRF) for Zr, Nb, Y, Sr and Rb. Mo Compton peak mass absorption coefficient determinations were used for all samples. The relative trace element abundances and interelement ratios thus obtained were utilised to select

a number of samples (26) for detailed study. Whole rock XRF analyses for major and trace elements were obtained from Dr A.P. le Roex and an electron microprobe was used by the author for glass and mineral analyses. Analytical procedures are detailed in Appendix 1.

Chapter Two

P E T R O G R A P H Y

The primary aims of the petrographic study of rocks are concerned not only with the identification of minerals and examinations of grain size and texture, all of which play important roles in the identification and classification of samples, but also with the elucidation of petrogenetic history for any particular rock or suite of rocks. In volcanic rocks texture and phenocryst assemblage are the most obvious characteristics. The phenocryst assemblage is defined as the group of solid phases in equilibrium with the magma prior to quenching (Cox et al, 1979) and provide information on the physio-chemical conditions experienced by the lava. Thermal and compositional disequilibrium and changing physical environments may also be indicated by resorbed, zoned or mantled phenocrysts.

The quenching of volcanic magmas, identified by characteristic textures and recognisable modes of mineral crystallization, preserves the phenocryst assemblage and also forms the fine grained volcanic groundmass. Extremes of quenching, such as those found in pillow basalt rinds, produces a volcanic glass, the composition of which represents a liquid composition. The chemistry of this glass, when unaltered, can be a valuable petrogenetic indicator.

On the basis of petrographic analysis and whole rock chemistry, different basalt types can be identified in the Agulhas 32,

dredge 3 samples (see Table 1). Each different basalt type is dealt with individually in the following sections. Descriptions of individual rocks are presented in appendix 2.

The Sparsely Plagioclase Phyric Suite

The basalts that comprise this suite (Groups A & B) are all sparsely porphyritic pillow basalts, and total phenocryst content of each sample is < 10%. The phenocryst assemblage is dominated by plagioclase which constitutes 60-100% of total phenocryst content. Olivine phenocrysts do occur, and in some samples (AG32-3-105, AG32-3-113) can comprise up to 40% of the phenocryst assemblage. Titanomagnetite microphenocrysts occur only in sample AG32-3-94 and only in trace amounts.

Plagioclase phenocrysts ($An_{75} - An_{90}$) range in length from < 1mm to 3mm and they are equant or stubby crystals that form euhedral or subhedral zoned grains that may be altered. Most grains show evidence of resorption, particularly at the (001) grain terminations, however resorption does occur at all grain boundaries. Most grains are rimmed by a distinct quench selvage (Figure 3) considerably more sodic than the core of the grain ($An_{55} - An_{70}$). This selvage is seldom broader than 0.01 mm but in a number of grains, where the core is extensively resorbed, the mantle is considerably broader than 0.01 mm (see Figure 4). Rapid growth of quench plagioclase frequently results in irregular quench mantles giving grain boundaries a fuzzy appearance.

GROUP A (EVOLVED, SPARSELY PLAGIOCLASE PHYRIC BASALTS)

AG32-3-2	AG32-3-18	AG32-3-32
AG32-3-50	AG32-3-72	AG32-3-74
AG32-3-75	AG32-3-83	AG32-3-86
AG32-3-97	AG32-3-100	AG32-3-109
AG32-3-112		n = 13

GROUP B (LESS EVOLVED, SPARSELY PLAGIOCLASE PHYRIC BASALTS)

AG32-3-63	AG32-3-94	AG32-3-99
AG32-3-105	AG32-3-108	AG32-3-111
AG32-3-113		n = 7

GROUP C (OLIVINE PHYRIC BASALTS)

AG32-3-30	AG32-3-36	AG32-3-39
AG32-3-40	AG32-3-46	n = 5

GROUP D (HIGHLY PLAGIOCLASE PHYRIC BASALT)

AG32-3-5		n = 1
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Table 1: Samples comprising the different basalt groups. Grouping is on the basis of petrography and whole rock geochemistry. See text for details.



Figure 3. Plagioclase phenocryst in microlitic groundmass showing the narrow quench selvage. 40X magnification. Field of view = 3.6 mm. From AG32-3-18 (Group A).

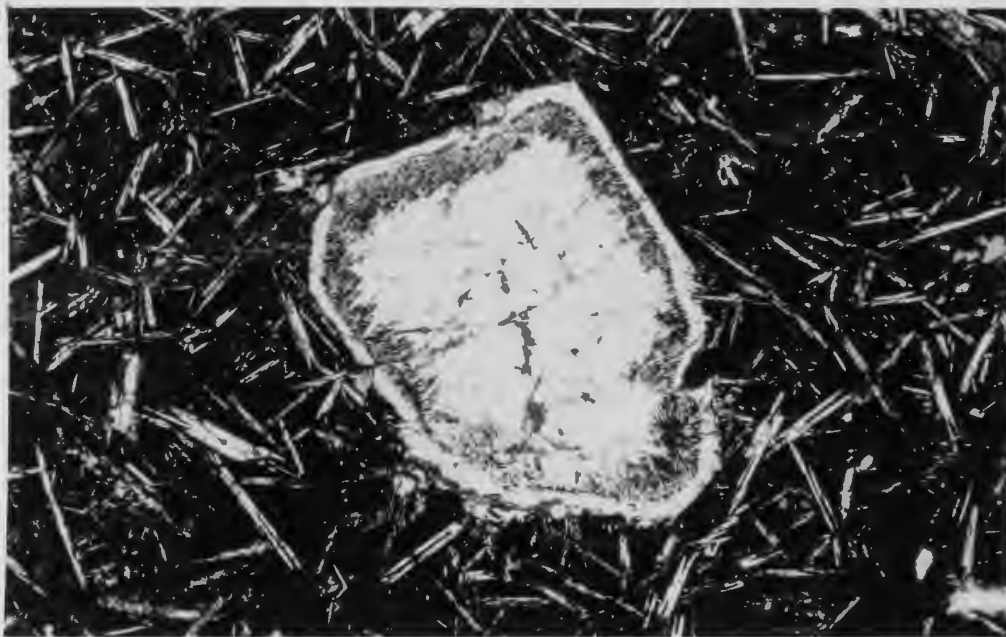


Figure 4. Resorbed plagioclase phenocryst rimmed by a distinct quench selvage, set in microlitic groundmass. Field of view = 3.6 mm. 40X magnification. AG32-3-63.

Trapped melt inclusions are common in many plagioclase phenocrysts. They occur as small rhombic inclusions < 0.1 mm in size. These inclusions are often concentrated in the core zones of crystals. Larger melt inclusions do occur, and they are prominent in some plagioclase phenocrysts from sample AG32-3-100. Most plagioclase phenocrysts are normally zoned, the maximum range being 24% An, from a phenocryst core of composition An_{85} , to the quench selvage of composition An_{61} (Sample AG32-3-18). Zoning in more evolved basalts is usually more pronounced (8 - 24% An) than in the less evolved samples (3-15% An) and a few phenocrysts are reversely zoned. The phenocrysts are commonly twinned (albite twinning), and they occur as individual grains and in glomeroporphyritic aggregates.

Olivine phenocrysts and microphenocrysts are rare. Phenocrysts occur as hopper crystals (Donaldson, 1976) up to 1mm in size, the compositions of which range from Fo_{76} to Fo_{91} . Inclusions of trapped melt are common. Microphenocrysts are more common than phenocrysts. They are about 0.3mm * 0.3mm in size and range from euhedral to anhedral grains and are relatively fresh although many exhibit some fine alteration products along fractures. Melt inclusions are rare in microphenocrysts. Microphenocrysts occur as individual grains, but also occur in the cores of plagioclase spherulites or clustered together in glomeroporphyritic aggregates.

Titano-magnetites occur in all samples. They occur as minute (<0.001 mm) quench grains. They are disseminated throughout the groundmass, and are occasionally concentrated within

segregation vesicles. A single sample (AG32-3-94) contains very rare microphenocrysts of titano-magnetite. These 0.03 mm blebs are all adjacent to a single olivine grain.

Segregation vesicles (Figure 5) usually about 1mm in diameter, are randomly scattered throughout all samples. They are commonly infilled with mesostasis, leaving only a relict vapour bubble. These vesicles range from opaque to transparent and are commonly a locus of alteration. Many are rimmed by yellow-green smectites. Trapped melt also occurs interstitially and much has devitrified. Vitreous volcanic glass occurs in pillow rinds.

Groundmass phases are olivine ($\text{Fo}_{75} - \text{Fo}_{85}$), plagioclase ($\text{An}_{64} - \text{An}_{74}$), clinopyroxene and titano-magnetite. The samples are olivine deficient in terms of Kirkpatrick's (1976) classification. Quench crystal morphologies are illustrated in figure 6. Olivine occurs as spinifex or H shaped quench crystals. Clinopyroxene is very rare, and it occurs as square, totally skeletal hopper crystals that have a melt filled core. Plagioclase microlites dominate the groundmass. They are about 1mm long, with 1:10 aspect ratios. They occur as spherulites, fan or sheaf shaped aggregates or as individual crystals. The latter are normally zoned (up to 7% An), skeletal and contain rare melt inclusions. Rounding of (001) terminations is common, indicating resorption.

Distinct zones can be recognised in the groundmass of pillow basalts (Kirkpatrick, 1976). A summary of these zones is presented in Table 2. The complete range of zones from 1 to 6

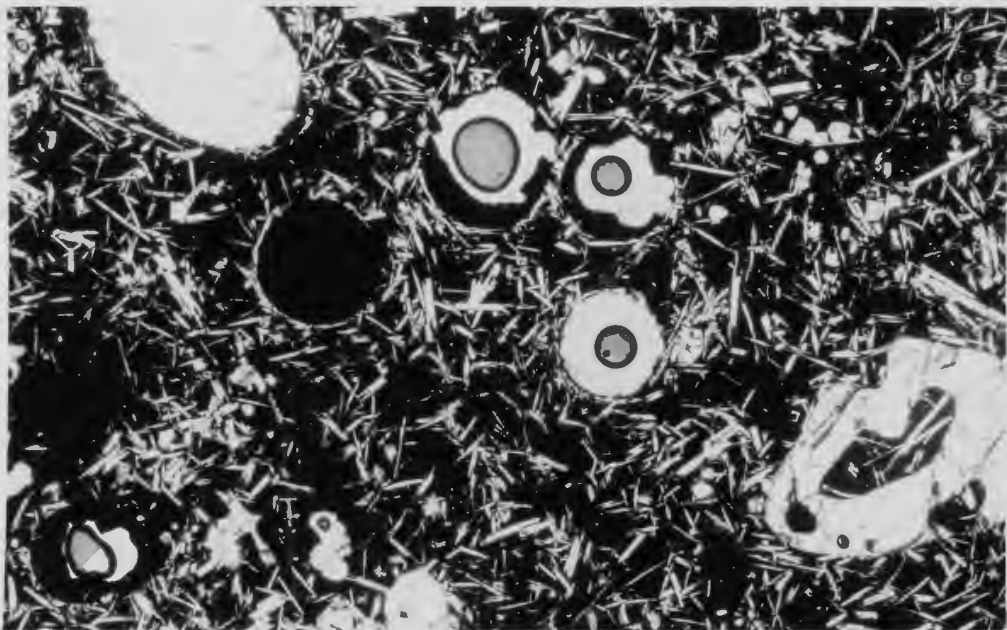


Figure 5. Segregation vesicles (center) in microlitic groundmass. Both opaque and transparent varieties occur. A melt filled hopper olivine is visible in the lower right hand corner. 20X magnification. Field of view = 7.2 mm. Vesicles are approximately 1 mm in diameter. AG32-3-46 (Group C).

**6a**

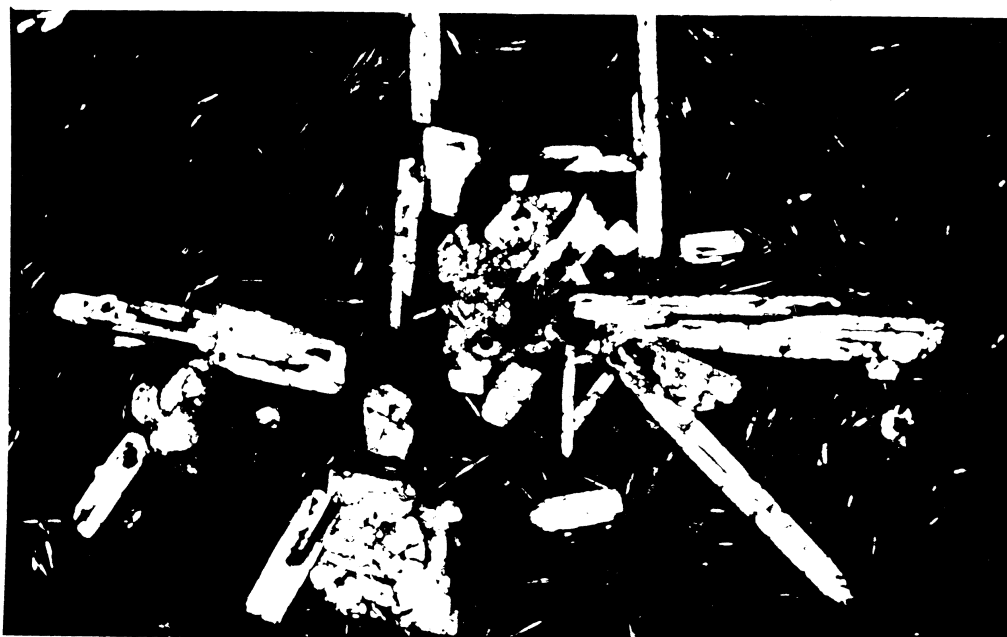
(AG32-3-36)

**6b**

(AG32-3-39)

**6c**

(AG32-3-83)

**6d**

(AG32-3-50)

**6e**

(AG32-3-113)

Figure 6 a. Hopper olivines in microlitic groundmass. 20X magnification. Field of view = 7.2 mm. Olivine in the top left hand corner is about 2mm across.

Figure 6 b. Spinifex olivine, hopper olivine and microlitic plagioclase in glass. 200 X magnification. Field of view = 0.72 mm.

Figure 6 c. Needle like plagioclase crystals and a melt filled suspected clinopyroxene (square at top left). 200 X magnification. Field of view = 0.72 mm.

Figure 6 d. Hopper olivines and microlitic plagioclase in a glomeroporphyritic intergrowth in glassy groundmass. 40 X magnification. Field of view = 3.6 mm.

Figure 6 e. Quench olivine crystals amongst glass and plagioclase spherulites. 400 X magnification. Field of view = 0.36 mm. These H shaped olivines are less than 0.01 mm long.

ZONE 1	- Quench glass zone - Some skeletal- euhedral olivine and plagioclase - No spherulites
ZONE 2	- Isolated plagioclase spherulites - Few olivine dendrites - No pyroxene - No oxides
ZONE 3	- Partially coalesced plagioclase spherulites - Few olivine dendrites - No identifiable pyroxene - No oxides
- - - - - GRADATIONAL CONTACT - - - - -	
ZONE 4	- Fully coalesced spherulites - Groundmass olivine as equant anhedral crystals or "lantern chain" skeletons - Grain size increased away from pillow rim - No identifiable pyroxene
- - - - - GRADATIONAL CONTACT - - - - -	
ZONE 5	- Spherulites become bow-tie shaped - Groundmass olivine in conspicuous. May be altered - Pyroxene occurs between plagioclase microlites - Titano-magnetite may be present
- - - - - GRADATIONAL CONTACT - - - - -	
ZONE 6	- Individual plagioclase laths become prominent (microlitic zone) - Spherulites disappear towards core - Pyroxene may be present. It occurs progressively more commonly as fan shaped dendrites between plagioclase laths - Titano-magnetite grain size increases towards pillow center.

Table 2: Zones defined by Kirkpatrick (1976), as occurring between rind and core of pillow basalts.

occurs in the groundmass of the sparsely plagioclase phyric basalts. The visible zones are: (1) glass zone, (2) zone of glass with isolated plagioclase spherulites and a few olivine dendrites, (3) zone comprising partially coalesced plagioclase spherulites and a few olivine dendrites, (4) zone of fully coalesced spherulites with well defined boundaries, (5) zone of bow tie plagioclase spherulites possessing poorly defined edge and (6) an interior plagioclase microlitic zone. Representative examples are illustrated in figure 7. The greatest range is exhibited by sample AG32-3-105 (zones 1-5).

The Olivine Phyric Basalts

Basalts from this group range from sparsely porphyritic (< 5% phenocrysts) to highly porphyritic (> 20% phenocrysts); all contain olivine comprising more than 95% of the total phenocryst content. Both phenocrysts and microphenocrysts of olivine occur, with phenocrysts ranging in size from 2-3 mm while microphenocrysts are considerably smaller (0.1 - 1.5 mm across). The olivine phenocrysts are normally zoned, skeletal hopper crystals. The maximum range in composition is from Fo₉₀ (core) to Fo₈₀ (rim). Trapped melt and Cr-spinel inclusions are common and are usually < 0.1 mm across.

Olivine microphenocrysts (Fo_{78.5} - Fo_{80.4}) are euhedral or subhedral, and are normally zoned. They are compositionally distinct from olivine phenocrysts in terms of forsterite content. They contain rare melt inclusions. Olivine

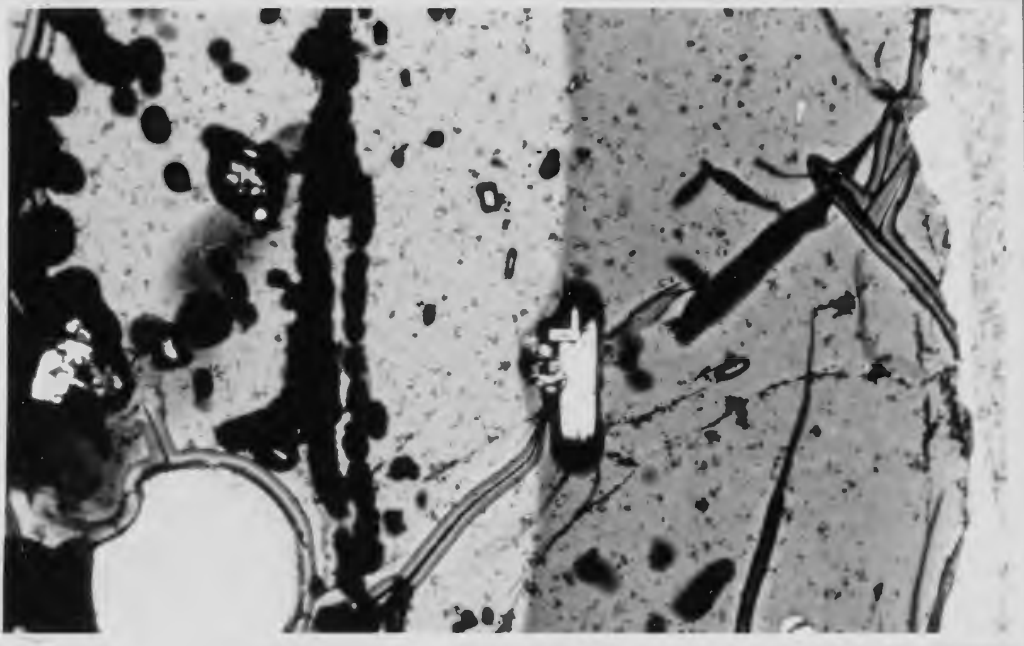


Figure 7 a. Kirkpatrick's (1976) zones 1 (on right) and 2 (on left) moving from the outer quench glass rim on a pillow basalt inwards. 40 X magnification. Field of view = 3.6 mm. From AG32-3-105 (Group B).



Figure 7 b. Zone 3 (on right) grading into zone 4 (on left). This clearly illustrates the increasing proportion of plagioclase spherulites. 40 X magnification. Field of view = 3.6 mm. From AG32-3-39 (Group C).



Figure 7 c. Zone 5. This illustrates the coalesced plagioclase spherulites comprising the groundmass. Also visible are needle like plagioclase microlites and quench olivine. 100 X magnification. Field of view = 1.45 mm. From AG32-3-113 (Group B).



Figure 7 d. Zone 6, the microlitic core of a pillow basalt. Also visible are segregation vesicles (left margin) and an olivine microphenocryst (center bottom). 20 X magnification. Field of view = 7.2 mm. From AG32-3-46 (Group C).

phenocrysts and microphenocrysts usually occur as individual crystals, but glomeroporphyritic aggregates do occur.

Plagioclase phenocrysts are rare (< 5% of total phenocrysts). They occur as large (2 - 3 mm) rounded, unzoned grains (An_{80} - An_{90}), that in some cases, exhibit complex overgrowth patterns (Figure 8). Embayed edges are common, as are small rhombic glass inclusions. These glass inclusions define concentric zones in some crystals. All plagioclase phenocrysts are rimmed by narrow (< 0.01 mm) quench selvages that are compositionally and optically distinct from the mantle and core of the grain (see Figure 3). These selvages are considerably more sodic (An_{60} - An_{70}) than phenocryst cores, and have similar compositions to the quench plagioclase crystals and plagioclase microlites.

Red chromium spinel is a rare microphenocryst phase (Figure 9), occurring as euhedral and subhedral grains < 1 mm in size. These microphenocrysts are very dark red or nearly opaque and may contain scattered inclusions of trapped melt. Cr-spinel also occurs as rounded grains < 0.1 mm in diameter, enclosed within olivine. Spinel appears to be the only non-silicate phase present in this group, save for a single, minute sulphide bleb in AG32-3-30.

All the textural zones defined by Kirkpatrick (1976) (see Table 2) are represented in the olivine phyric basalts. Sample AG32-3-39 is the only example in this group exhibiting zones 1 to 5. The remainder of the group have groundmass that can be classified into zones 4 to 6.

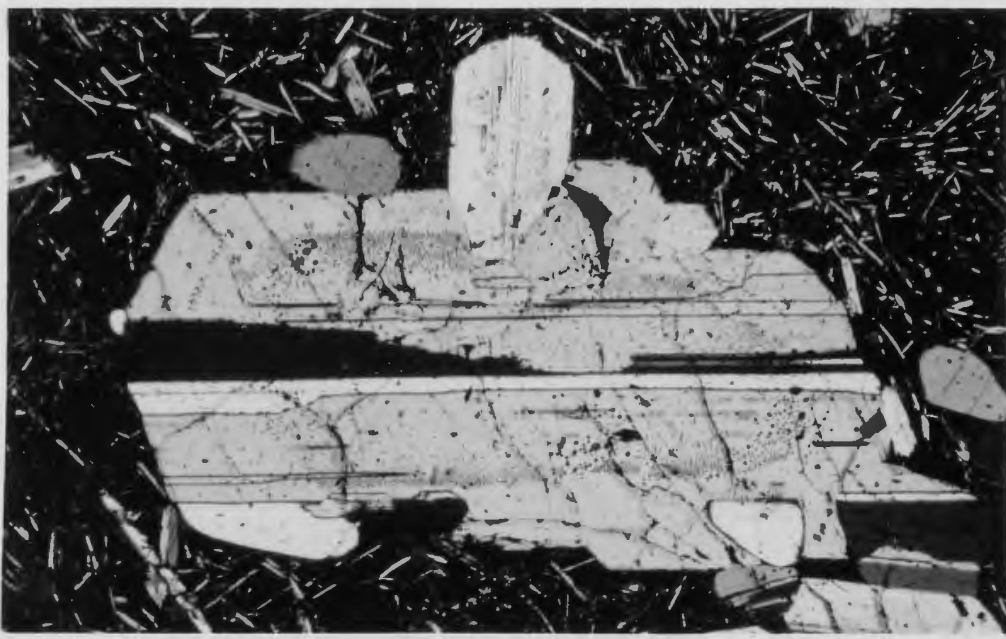


Figure 8. Complexly overgrown plagioclase phenocryst in microlitic groundmass. This phenocryst contains abundant melt inclusions, forming concentric zones within the crystal. 20 X magnification. Field of view = 7.2 mm. From AG32-3-18 (Group A).



Figure 9. Cr-spinel microphenocrysts at the edge of a thin section. Also visible is a dark, glassy groundmass and a portion of an olivine phenocryst (right). Microphenocrysts are fractured and the large central example contains a trapped melt droplet. 200 X magnification. Field of view = 0.72 mm. AG32-3-36 (Group C)

Groundmass olivine ($\text{Fo}_{79.2}$ - $\text{Fo}_{80.0}$) in the olivine phyric basalts is rare. Quench oxide phases (titano-magnetites) do occur but these are rare. They form skeletal crystals and are up to 0.01mm long, and are disseminated throughout the groundmass. Segregation vesicles (up to 1% modal volume) are scattered throughout the groundmass of all samples. They are smaller than 0.5 mm in diameter and relict vapour bubbles occur within the cores of some of these vesicles. The interior of the vesicle has devitrified to cryptocrystalline or microcrystalline mesostasis.

The Highly Plagioclase Phyric Basalt - Group D

Group D comprises a single sample, AG32-3-5, which contains 20% phenocrysts, of which plagioclase contributes 95%, and olivine 5%. Plagioclase phenocrysts (An_{84} - An_{88}) are commonly about 2mm * 0.75mm in size, but reach a maximum size of 3mm * 4mm. They are euhedral or subhedral and are always rimmed by a more sodic (An_{81}) quench selvage. The cores of smaller grains are more commonly zoned than larger grains. Trapped melt inclusions are ubiquitous and occur as small, black blebs about 0.01mm across, and are found in the core regions of phenocrysts. Larger inclusions (1mm * 0.3mm) also occur. Slight resorption is evident along (001) grain terminations and along fractures within the plagioclase grains.

Olivine phenocrysts range from hopper to euhedral or subhedral crystals about 1mm in size. They contain significant amounts of trapped melt, are optically unzoned, are unaltered, and have restricted compositional range (Fo_{86} - Fo_{88}). They usually occur as individual crystals, but may occur in glomeroporphyritic aggregates. Euhedral olivine microphenocrysts ($0.1\text{mm} \times 0.1\text{mm}$) are rare.

The groundmass is olivine deficient in Kirkpatrick's (1976) classification and zone 6, the microlitic core zone, is most obvious. Plagioclase microlites are about $0.2\text{mm} \times 0.02\text{mm}$ in size. Bow tie spherulites and fan or sheaf like aggregates are also common. Groundmass olivine occurs as spinifex and H shaped quench grains. Skeletal clinopyroxene is present and is disseminated throughout the groundmass along with quench titano-magnetites and segregation vesicles.

Chapter Three

M I N E R A L O G Y

The relationship between a lava and its postulated primary magma is of critical importance in an evaluation of the evolutionary history of the magma. Magmatic processes such as fractional crystallization and crystal accumulation leave their imprint on the erupted lava, although the latter can be ignored if the lavas unequivocally represent liquid compositions. Fresh aphyric lavas and basaltic glasses are conventionally assumed to represent liquids. Careful evaluation is required in the case of porphyritic lavas to distinguish between processes such as crystal accumulation, which may change the lava composition considerably, or processes such as compensated crystal settling (Cox and Bell, 1972; Krishna et al, 1977), which could allow a lava to become highly porphyritic, while retaining its original composition.

The possible occurrence of high - MgO precursors to the voluminous ocean floor basalts is still a contentious issue (O'Hara, 1965, 1968; Green & Ringwood, 1967; Hart & Davis, 1978, 1979; Jakobsen et al, 1978; Dungan & Rhodes, 1979; Elthon, 1979; Clarke & O'Hara, 1979; Elthon & Ridley, 1979). The possibility that high-MgO picritic lavas such as those of the highly olivine phyric suite (Group C) represent high MgO liquids, is therefore important in understanding ocean floor basalt petrogenesis.

It is to be expected that consanguinous lavas should show uniform variations in mineral chemistry, particularly with respect to minor elements. Compositional variations amongst such lavas caused by low pressure fractional crystallization are revealed by systematic chemical variations in the composition of liquidus minerals (e.g. Bryan & Moore, 1977; Hekinian et al, 1976). Similarly, discontinuities in mineral compositions may be indicative of more than one parental magma (e.g. Bougault & Hekinian, 1974; Nisbet & Fowler, 1978; Langmuir et al, 1977; le Roex et al, 1981), and thus the study of the mineralogy and mineral chemistry of a suite of lavas can provide valuable insight into the processes governing the formation and evolution of a magma. Such a mineralogical study should indicate whether the highly magnesian, olivine phyric basalts represent a high MgO liquid, or are the products of a process such as crystal accumulation. All mineral analyses obtained are presented in appendix 2 (microfiche).

In the following chapter, the naming convention

Group A.....Evolved sparsely plagioclase phyric basalts

Group B.....Sparsely plagioclase phyric basalts

Group C.....Olivine phyric basalts

Group D.....Highly plagioclase phyric basalt

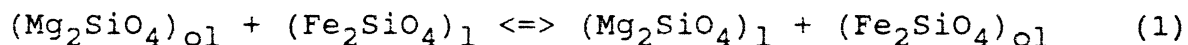
has been used.

Olivine

Olivine phenocrysts are ubiquitous in the AG32, dredge 3 samples. Representative analyses are presented in Table 3. Groundmass olivine occurs in the highly olivine phyric suite (Group C), the sparsely plagioclase phyric suite (Group B), and in the evolved sparsely plagioclase phyric suite (Group A). No groundmass olivine occurs in the highly plagioclase phyric basalt (Group D). From the whole rock geochemistry (see Figure 26), it is apparent that olivine is the liquidus phase, or at least a co-liquidus phase, for all groups at less than 10Kbar, and thus a study of the distributions of Fe, Mg and Ni in olivine should reflect the nature of the liquids at the onset of crystallization.

1: Fe - Mg distributions

Ferrous iron and magnesium distribution between olivine and coexisting liquid are commonly used to assess whether olivine phenocrysts are in equilibrium with the host rocks in which they occur. The distribution can be represented by the equation:



	GROUP A				GROUP B				GROUP C				GROUP D		
SiO ₂	40.02	39.96	39.74	39.*08	39.84	40.07	39.52	38.*89	40.95	40.23	39.93	39.*16	40.02	40.10	39.72
Al ₂ O ₃	0.06	0.05	0.04	0.06	0.06	0.06	0.04	0.03	0.09	0.08	0.04	0.06	0.05	0.07	0.09
Cr ₂ O ₃	0.03	0.03	0.02	0.03	0.05	0.04	0.05	0.03	0.07	0.07	0.06	0.05	0.05	0.06	0.07
FeO*	11.11	13.06	15.48	18.97	11.90	14.46	17.21	21.66	9.78	13.16	14.83	18.59	11.20	12.75	12.72
MnO	0.16	0.18	0.23	0.28	0.18	0.21	0.25	0.36	0.17	0.18	0.22	0.26	0.20	0.20	0.02
MgO	47.87	46.51	44.77	41.47	46.51	45.00	43.22	39.36	48.07	46.02	44.59	42.09	46.97	46.39	46.08
CaO	0.29	0.29	0.30	0.30	0.30	0.32	0.31	0.29	0.27	0.27	0.29	0.30	0.31	0.36	0.36
NiO	0.24	0.18	0.18	0.17	0.26	0.18	0.16	0.11	0.30	0.27	0.29	0.18	0.20	0.15	0.16
TOTAL	99.78	100.26	100.76	100.36	99.10	100.34	100.76	100.75	99.70	100.28	100.18	100.69	99.00	100.08	99.40

ATOMIC PROPORTIONS CALCULATED ON THE BASIS OF 4 OXYGENS

Si	0.991	0.993	0.994	0.998	0.997	1.001	0.996	1.001	1.007	1.000	1.001	0.995	0.999	0.997	0.995
Al	0.002	0.001	0.001	0.002	0.002	0.002	0.001	0.001	0.002	0.002	0.001	0.002	0.002	0.002	0.003
Cr	0.001	0.001	-	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001
Fe	0.230	0.272	0.324	0.405	0.249	0.302	0.363	0.466	0.201	0.273	0.311	0.395	0.234	0.265	0.266
Mn	0.003	0.004	0.005	0.006	0.004	0.004	0.005	0.008	0.003	0.004	0.005	0.006	0.004	0.004	0.004
Mg	1.767	1.723	1.669	1.578	1.735	1.675	1.577	1.510	1.762	1.705	1.666	1.594	1.748	1.719	1.720
Ca	0.008	0.008	0.008	0.008	0.008	0.009	0.008	0.008	0.007	0.007	0.008	0.008	0.008	0.010	0.010
Ni	0.005	0.004	0.004	0.004	0.005	0.004	0.003	0.002	0.006	0.005	0.004	0.001	0.004	0.003	0.003
Sum	3.007	3.006	3.005	3.002	3.001	2.998	2.974	2.997	2.989	2.997	2.997	3.002	3.000	3.001	3.002
Fe	88.4	86.2	83.5	79.3	87.3	84.6	81.3	76.1	89.6	86.0	84.1	79.9	88.0	86.5	86.4
Fe	11.6	13.8	16.5	20.7	12.7	15.4	19.7	23.9	10.4	14.0	15.9	20.1	12.0	13.5	13.6

Table 3: Representative Olivine analyses from the AG32-3 basalt groups, showing compositional range. All iron as FeO. Olivines marked with a * are quench grains. The remainder are phenocrysts.

The distribution coefficient (K_D) for this reaction can be written as:

$$K_D = \frac{\overset{\text{ol}}{X} \underset{\text{FeO}}{\text{FeO}}}{\underset{\text{X}}{1} \underset{\text{FeO}}{\text{FeO}}} * \frac{\overset{\text{l}}{X} \underset{\text{MgO}}{\text{MgO}}}{\underset{\text{X}}{\text{ol}} \underset{\text{MgO}}{\text{MgO}}} \quad (2)$$

Where X represents mole fractions of oxides in the respective phases.

Roeder and Emslie (1970) investigated this equilibrium for a variety of basaltic compositions over a temperature range from 1150° - 1300° C, and a range in oxygen fugacities from 10^{-68} to 10^{-12} atmospheres, at one atmosphere total pressure. They concluded that the equilibrium constant (K_D) was essentially constant at 0.30 ± 0.03 and independent of temperature and oxygen fugacity. These results have been experimentally confirmed over a wide variety of compositions (Roeder, 1974; Longhi et al, 1978), although some authors have predicted slightly higher average K_D values of 0.33 for high MgO liquids (Cawthorne et al, 1974; Thompson, 1975; O'Hara, 1977).

Systematic increases in K_D with increasing pressure have been noted (Bender et al, 1978; Longhi et al, 1978; Mysen, 1975), as has a small but significant dependence of K_D on silica concentration (Longhi et al, 1978). In view of the implied low pressure control on whole rock - glass variations, K_D values of 0.30 ± 0.03 will be regarded as representing equilibrium conditions.

The vast majority of analysed olivine phenocrysts and microphenocrysts from the sparsely plagioclase phyric basalts, the evolved sparsely plagioclase phyric basalts and the highly olivine phyric basalts, show normal zoning with maximum decreases in forsterite content from core to rim of 3.8% Fo (Group A), 7.9% Fo (Group B) and 9.5% Fo (Group C). Based on the assumption that the core of the most Mg-rich olivine should be in equilibrium with a liquid with an FeO/MgO ratio of the whole rock, phenocryst and microphenocryst histories can be evaluated. Should Fe-Mg K_D 's fall within the equilibrium range (i.e. 0.27 - 0.33), then significant phenocryst redistribution is unlikely to have taken place. However, should calculated K_D 's be too low, indicating the presence of olivines too forsteritic to have crystallized from a melt with the whole rock FeO/MgO ratio, alternative explanations will have to be found. Two such alternatives, mentioned by le Roex (1980) are that (1) the olivines are remnants, left over after the bulk of liquidus olivines have been lost by settling from a more primitive magma, or (2) that the rocks are enriched in Mg rich xenocrystic olivines. K_D 's that are too high, when calculated using whole rock FeO/MgO ratios indicate that the rock was too Mg rich to have crystallized olivine with observed compositions.

Whole rock-olivine Fe-Mg distribution coefficients for the different basalt groups are presented in Table 4. Glass-olivine microphenocryst distribution coefficients at various glass $\text{Fe}_2\text{O}_3/\text{FeO}$ ratios are presented in Table 5.

GROUP A	Mean $K_D^{\text{Fe-Mg}}$	=	0.271	$\sigma = 0.081$ $n = 67$
GROUP B	Mean $K_D^{\text{Fe-Mg}}$	=	0.306	$\sigma = 0.053$ $n = 41$
GROUP C	Mean $K_D^{\text{Fe-Mg}}$	=	0.238	$\sigma = 0.079$ $n = 25$
GROUP D	Mean $K_D^{\text{Fe-Mg}}$	=	0.340	$\sigma = 0.019$ $n = 5$

Table 4: Distribution Coefficient (K_D) of Fe, Mg between Olivines and Whole Rocks. Equilibrium $K_D = 0.30 \pm 0.03$ (Roeder & Emslie, 1971)

		All Fe as FeO	$\text{Fe}_2\text{O}_3/\text{FeO} = 0.10$	$\text{Fe}_2\text{O}_3/\text{FeO} = 0.15$	$\text{Fe}_2\text{O}_3/\text{FeO} = 0.20$
GROUP A	AG32-3-2	$K_D = 0.27$	$K_D = 0.30$	$K_D = 0.31$	$K_D = 0.33$
	AG32-3-18	$K_D = 0.27$	$K_D = 0.30$	$K_D = 0.32$	$K_D = 0.34$
	AG32-3-32	$K_D = 0.20$	$K_D = 0.22$	$K_D = 0.23$	$K_D = 0.25$
	AG32-3-50	$K_D = 0.30$	$K_D = 0.33$	$K_D = 0.35$	$K_D = 0.37$
GROUP B	AG32-3-63	$K_D = 0.25$	$K_D = 0.28$	$K_D = 0.31$	$K_D = 0.32$
	AG32-3-105	$K_D = 0.24$	$K_D = 0.26$	$K_D = 0.29$	$K_D = 0.30$
GROUP C	AG32-3-36	$K_D = 0.24$	$K_D = 0.27$	$K_D = 0.29$	$K_D = 0.30$
	AG32-3-39	$K_D = 0.26$	$K_D = 0.29$	$K_D = 0.31$	$K_D = 0.33$
GROUP D	AG32-3-5	$K_D = 0.30$	$K_D = 0.33$	$K_D = 0.35$	$K_D = 0.37$

Table 5: Olivine - glass Fe-Mg Distribution Coefficients (K_D) at various $\text{Fe}_2\text{O}_3/\text{FeO}$ ratios for selected samples from each group. Quench olivine compositions used for Groups A & C. Olivine phenocryst compositions used for Groups B and D, due to lack of quench olivines in samples AG32-3-63, AG32-3-105 & AG32-3-5.

The olivine phyric basalts (Group C) have olivine - whole rock K_D 's too low (0.24) for the olivines to have formed in equilibrium with liquids having the FeO/MgO ratio of the whole rock. These results are in accord with the textures of these rocks, which are consistent with an origin due to the accumulation of olivine. Further support for such a process is obtained from the whole rock geochemical variations in Ni and other elements (see trace element geochemistry section).

Groups A and B, the evolved sparsely plagioclase phyric basalts and the sparsely plagioclase phyric basalts have mean olivine - whole rock K_D 's falling within the range indicating equilibrium crystallization of olivine (0.27 - 0.33). This suggests that phenocryst redistribution has not been a significant process affecting these rocks. The large scatter in K_D 's from these rocks can be ascribed to changes in magma composition with time, and to minor differences in the whole rock $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio for different samples. Following Brooks (1976) a $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio of 0.15 has been assumed for all whole rock analyses, however this ratio will be affected by varying physical parameters such as magma temperature, oxygen fugacity and the relative proportions of crystallizing phases.

The olivine - whole rock K_D 's calculated for the highly plagioclase phyric basalt (Group D) are consistently higher than 0.33 when a whole rock $\text{Fe}_2\text{O}_3/\text{FeO}$ of 0.15 is assumed. Assuming a whole rock $\text{Fe}_2\text{O}_3/\text{FeO}$ of 0.10 brings K_D 's into the equilibrium range (i.e. 0.27 - 0.33; Roeder & Emslie 1970). The nature of the olivines in this rock, i.e. unzoned, often

skeletal microphenocrysts, is consistent with relatively rapid crystallization during quenching.

Olivine - glass distribution coefficients can be used to determine whether the parental magma at eruption was in equilibrium with the olivine phenocrysts, microphenocrysts or quench olivine crystals. This is most clearly illustrated in the highly olivine phyric basalts (Group C). In these, skeletal quench olivines are in equilibrium with the glasses (K_D 's = 0.29 - 0.31 at $\text{Fe}_2\text{O}_3/\text{FeO} = 0.15$), while the phenocryst core - glass K_D 's (0.15) clearly indicate disequilibrium. This is further evidence for the olivine phenocrysts in the Group C basalts being cumulate crystals.

Olivine - glass K_D 's from the evolved sparsely plagioclase phyric basalts (Group A) and the sparsely plagioclase phyric basalts (Group B) indicate equilibrium crystallization, save for olivines from sample AG32-3-32. This sample has quench olivine - glass K_D 's of less than 0.23 at $\text{Fe}_2\text{O}_3/\text{FeO}$ of 0.15. Assuming a $\text{Fe}_2\text{O}_3/\text{FeO}$ of >0.20 brings olivine - glass K_D 's into the equilibrium range (0.27 - 0.33). Olivine - glass K_D 's for the highly plagioclase phyric (Group D) basalt fall within Roeder & Emslie's (1970) equilibrium range (0.27 - 0.33) when a glass $\text{Fe}_2\text{O}_3/\text{FeO}$ of ≤ 0.10 is assumed. As mentioned above, olivine - whole rock K_D 's also indicate equilibrium at this low $\text{Fe}_2\text{O}_3/\text{FeO}$, which is taken to reflect the primitive nature of this basalt ($\text{Mg\#} = 70.0$).

In summary; Group A and Group B phenocryst, microphenocryst and quench olivines are generally in equilibrium with both

whole rock and glass compositions in which $\text{Fe}_2\text{O}_3/\text{FeO}$ is assumed to be 0.15. It is assumed that these rocks have undergone no significant olivine redistribution. In contrast, Group C olivine phenocrysts are clearly out of equilibrium with both whole rock and glass compositions. These olivine phenocrysts are too forsteritic to have crystallised from the host whole rock compositions, implying phenocryst accumulation. Quench olivines are however in equilibrium with the quench glass, and this indicates that they grew during the final stages of magma evolution and during the eruption of the Group C lavas.

The Group D lavas contain abundant olivine microphenocrysts that are in equilibrium with both whole rock and glass composition at $\text{Fe}_2\text{O}_3/\text{FeO}$ of 0.10 or less. This low ratio is consistent with the primitive nature of the basalt (whole rock $\text{Mg\#} = 70.0$ at $\text{Fe}_2\text{O}_3/\text{FeO} = 0.15$). As the olivine microphenocrysts in this basalt are in equilibrium with both whole rock and glass, and have a limited compositional range ($\text{Fo}_{86} - \text{Fo}_{88}$), significant amounts of olivine redistribution is unlikely to have occurred.

2: Minor elements

In this section the minor elements (i.e. Al, Cr, Mn, Ca and Ni) in olivines from all basalts groups are presented and compared. In order to minimise any effects of olivine - groundmass diffusion and entrapment during cooling, only core analyses have been used.

2.1: Al_2O_3

Al_2O_3 ranges from below detection limit (0.02%) in Group A olivines, up to a maximum of 0.10 wt% in the sparsely plagioclase phyric (Group B) basalts. Most olivines in Groups A, C & D contain between 0.04 and 0.07 wt% Al_2O_3 , while the Group B basalts have slightly higher Al_2O_3 contents of 0.06 to 0.08 wt%. Figure 10 illustrates the distribution of aluminium in olivine phenocryst cores from the different basalt groups, and shows the lack of any systematic correlations between Al_2O_3 and Fo content of olivines.

2.2: Cr_2O_3

Chromium occurs at very low levels in olivines, ranging from below detection limit (0.03 wt%) in the evolved sparsely plagioclase phyric (Group A) basalts, up to 0.09 wt% in the olivine phyric (Group C) basalts. Cr shows a poor, but positive, correlation with the Fo and NiO contents of olivines (Figure 11). Deer et al (1966) attribute the minute amounts of Cr_2O_3 found in olivine analyses to exsolved chromic oxide octahedra, and plates exsolved parallel to the (001) and (100) planes. Charge balance and crystal chemical considerations would tend to preclude the substitution of Cr for other elements such as Ni in the olivine structure.

Figure 12 shows the distribution of Cr_2O_3 content of olivines for Groups A, B, C and D.

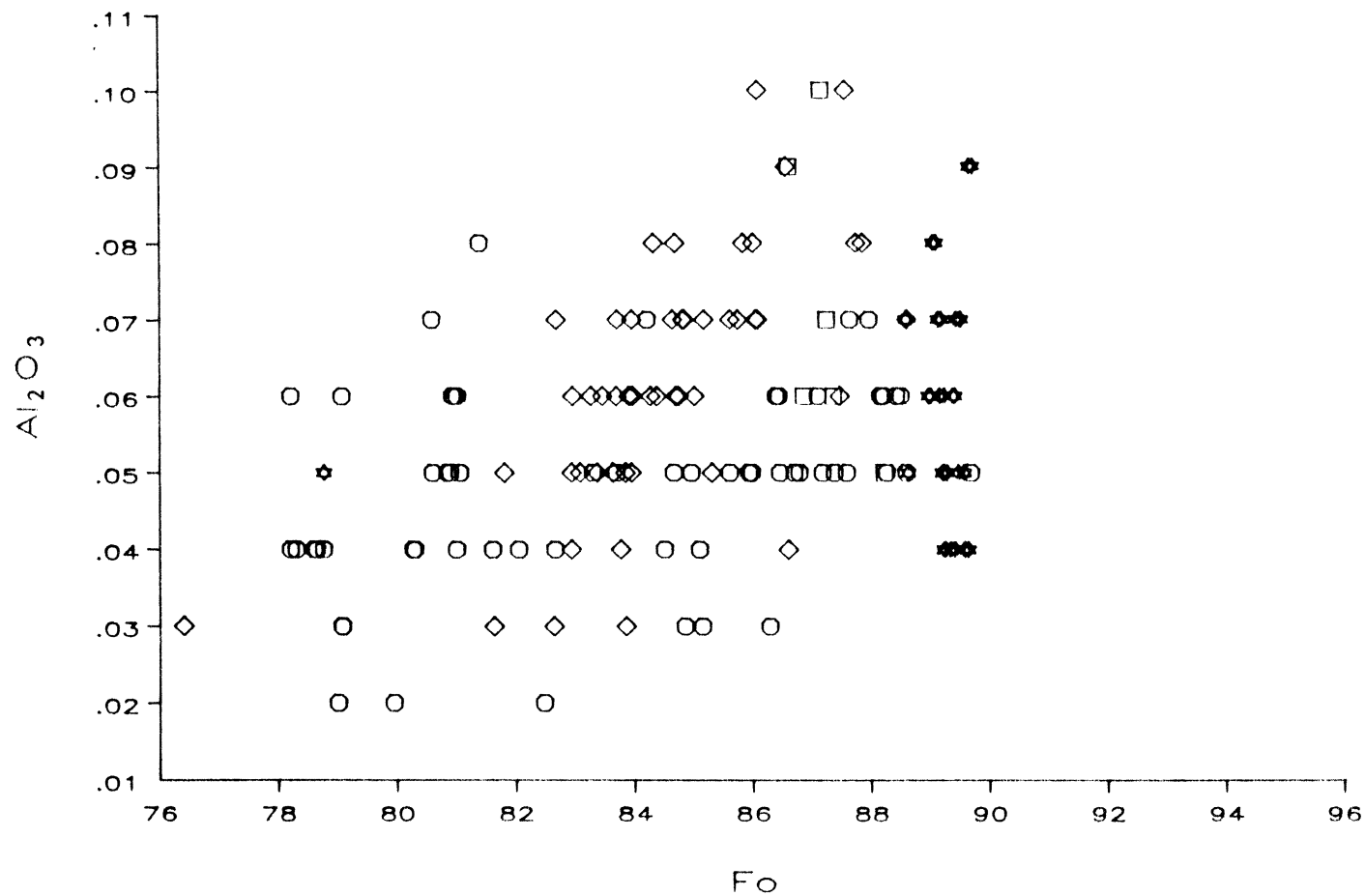


Figure 10. Al_2O_3 contents of olivine phenocryst cores plotted against Forsterite content. The poor, positive correlation is evident. Circles = Group A, Diamonds = Group B, Stars = Group C and Squares = Group D.

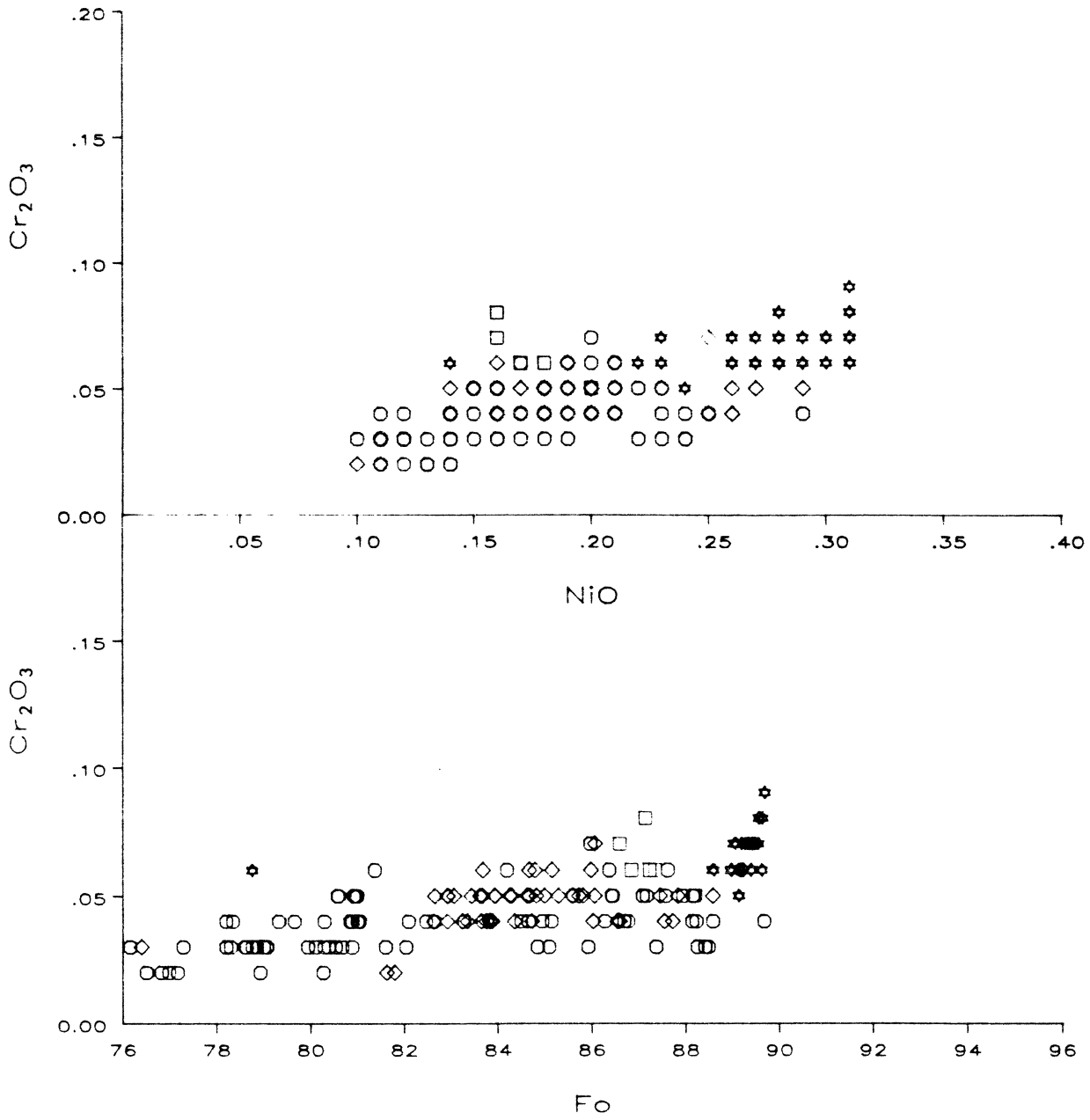


Figure 11. Cr_2O_3 vs NiO and Cr_2O_3 vs Forsterite content of olivines showing the general, positive correlation. Symbols as per Figure 10.

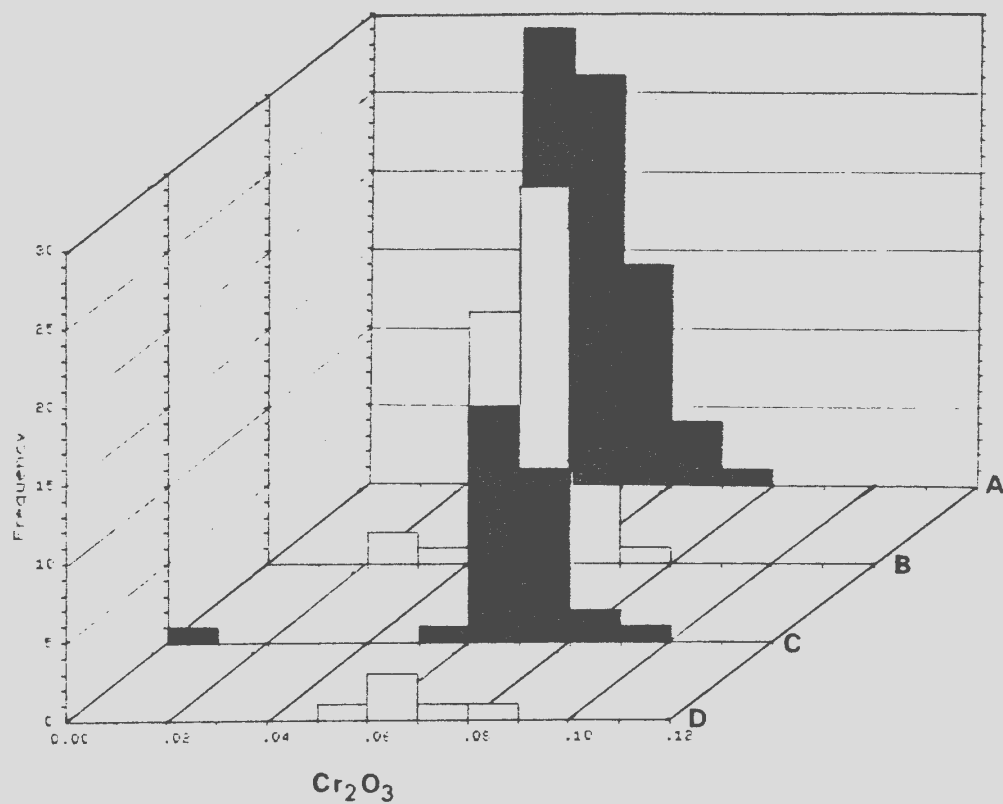


Figure 12. Distribution of Cr_2O_3 in olivine phenocryst cores. This shows that the Group C and Group D olivine phenocrysts have slightly higher Cr_2O_3 contents than olivine phenocrysts from Groups A and B.

2.3: MnO

MnO shows a broad negative correlation with the Fo content of olivines (Figure 13). This is explained by the easy substitution of Mn for Fe^{2+} (ionic radii; $\text{Fe}^{2+} = 0.074$, $\text{Mn}^{2+} = 0.080$ nanometers). It is thus not surprising that the most magnesian olivines, found in the olivine phyric (Group C) basalts, have the lowest MnO contents (0.12 - 0.19 wt% MnO). Olivines in the evolved plagioclase phyric (Group A) and plagioclase phyric (Group B) basalts have high MnO contents with maxima at 0.30 - 0.32 and 0.20 - 0.22 respectively.

2.4: CaO

Calcium contents of olivines have been linked to the cooling history and crystallization environment by Simkin & Smith (1970), who found that low pressure extrusive rocks contained olivines with comparatively high CaO contents (>0.10 wt% CaO) relative to those in intrusive rocks (<0.10 wt% CaO). The olivines of this study are no exception and contain between 0.22 and 0.40 wt% CaO (Figure 14). The olivine phyric (Group C) basalts have olivines with the lowest CaO content (0.24 - 0.32 wt% CaO). The evolved plagioclase phyric (Group A) basalts contain between 0.24 and 0.34 wt% CaO and olivines from the sparsely plagioclase phyric (Group B) basalts have a mean CaO content of 0.31 wt%. Olivines from the highly plagioclase phyric (Group D) basalt contain more CaO with a maximum between 0.35 and 0.36 wt% (Figure 14).

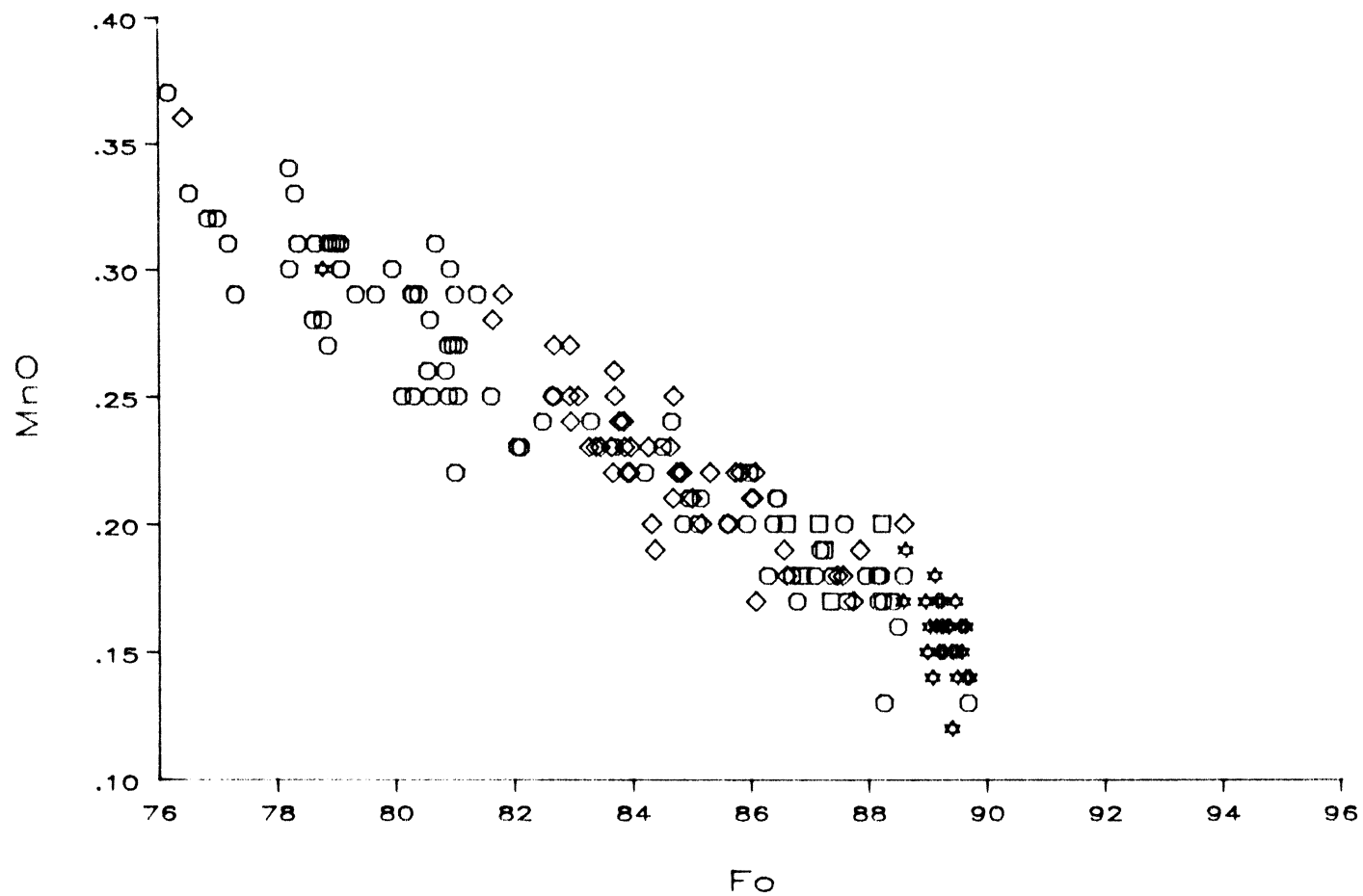


Figure 13. MnO vs Forsterite content of olivine phenocryst cores, showing strong negative correlation. Symbols as per Figure 10.

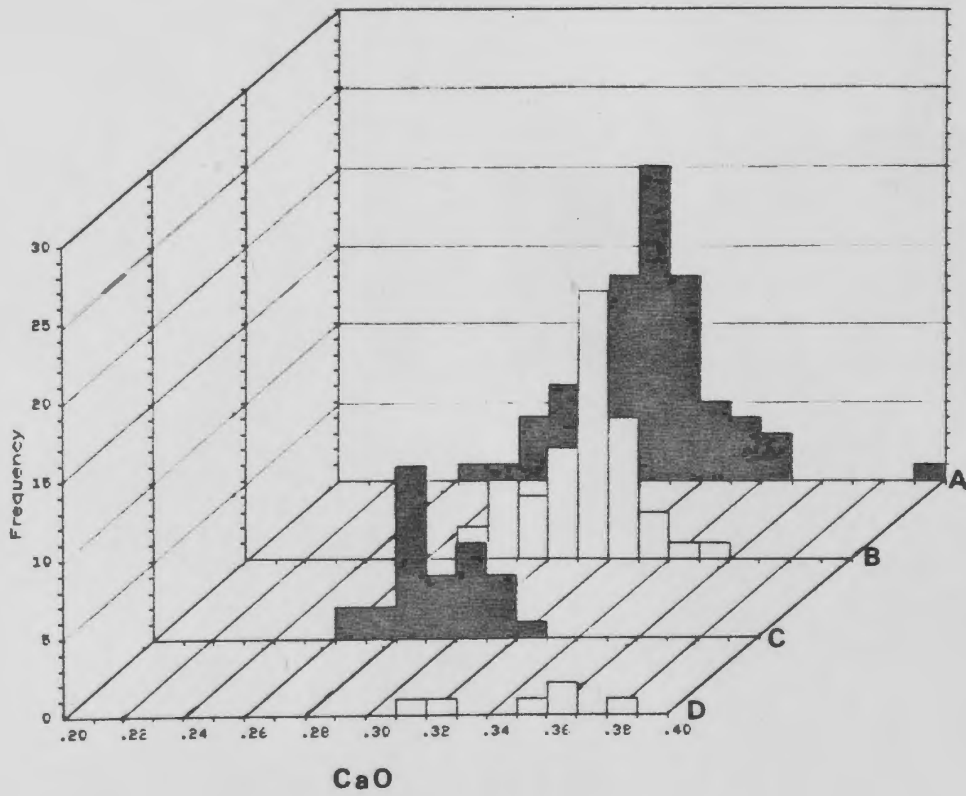


Figure 14. The distribution of Ca in olivine phenocryst cores. This illustrates the high Ca content of the Group D olivines.

2.5: NiO

NiO is detectable in all analysed olivines from all basalt groups and ranges in concentration from 0.1 wt% to 0.31 wt%. Group C olivines are clearly distinguished from other basalt groups by their high NiO content (0.22 - 0.31 wt%) while Group D olivines contain 0.16 - 0.20 wt% NiO. Group A and B olivines are very similar with respect to NiO contents (0.10 - 0.29 wt%). Figure 15 shows the distribution of NiO in olivines.

The high distribution coefficient of Ni between olivine and melts (see Table 6) results in the strong partitioning of this minor element into early forming olivine phenocrysts. Thus the distribution of NiO in these phenocrysts should closely reflect the composition of the parental melt. It is for this reason that Ni is the most studied of all the minor elements in olivine, and partitioning of Ni between olivine and silicate melts has received considerable attention since the pioneering work of Hakli and Wright (1967). It has been used to argue that andesites can (Hedge, 1971) and cannot (Taylor et al, 1969) be products of fractional crystallization of basalt; to identify primitive basaltic liquids (Kay et al, 1970; O'Hara et al, 1975; Sato, 1977); as a geothermometer (Leeman, 1974; Leeman & Scheidegger, 1977); to test crystallization models (Irving, 1974) and to constrain mantle composition (Allegre et al, 1977; Minster et al, 1977).

It was originally suggested that the Mg-Ni K_D between olivines and liquids was constant at between 2.0 and 2.6 (Sato, 1977), from the exchange reaction;

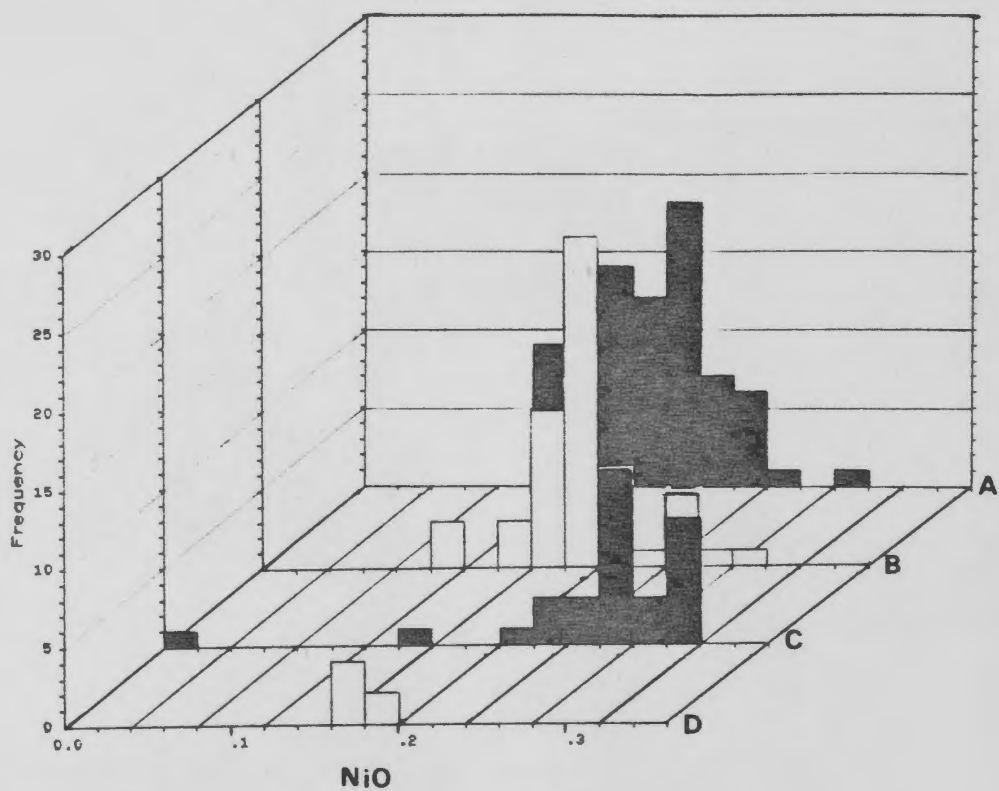
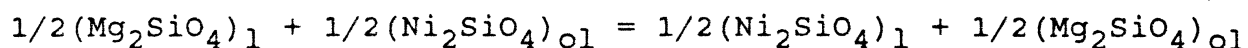


Figure 15. NiO distribution in olivine phenocryst cores. This illustrates the high NiO content of the Group C olivines.

	Sample	Predicted $\frac{\text{Ni}}{\text{D}_{01}}$	Calculated $\frac{\text{Ni}}{\text{D}_{01}}$ (whole rocks)
GROUP A	AG32-3-2	14.60	14.94
	AG32-3-18	18.51	11.96
	AG32-3-32	18.53	17.59
	AG32-3-50	15.92	16.35
	AG32-3-72	16.23	20.95
	AG32-3-74	16.81	16.67
	AG32-3-75	16.79	13.68
	AG32-3-83	17.34	17.70
	AG32-3-86	17.10	16.16
	AG32-3-97	18.21	12.37
	AG32-3-100	16.89	17.48
	AG32-3-109	17.28	20.19
	AG32-3-112	16.94	18.63
GROUP B	AG32-3-63	14.92	13.39
	AG32-3-94	13.17	10.81
	AG32-3-99	15.35	15.04
	AG32-3-105	15.12	16.28
	AG32-3-108	14.47	10.81
	AG32-3-111	15.77	13.71
	AG32-3-113	13.35	14.00
GROUP C	AG32-3-30	12.64	11.67
	AG32-3-36	8.20	5.34
	AG32-3-39	11.38	8.39
	AG32-3-40	7.11	4.45
	AG32-3-46	14.39	16.28
GROUP D	AG32-3-5	13.90	12.50

Table 6: $\frac{\text{Ni}}{\text{D}_{01}}$ calculated from $\frac{\text{Ni}}{\text{D}_{01}} = \frac{\text{Ni}^{01}}{\text{Ni}^t}$. Predicted $\frac{\text{Ni}}{\text{D}_{01}}$ from Hart & Davis' (1978) formula : $D = 124/\text{MgO} - 0.9$. Close agreement is generally found between observed and predicted values. Consequently, unless otherwise stated, Hart & Davis (1978) has been used in petrological modelling.



for which

$$K_D = \frac{\text{MgO}}{\text{NiO}}_1 \bigg/ \frac{\text{MgO}}{\text{NiO}}_{\text{ol}}$$

It has since been shown that there is a strong dependence of Mg-Ni K_D on the MgO content of the liquid (le Roex, 1980; Hart & Davis, 1978; Arndt, 1977, Irving & Kushiro, 1976; Takahashi, 1978). This relationship is illustrated in figure 16.

In this study Mg-Ni K_D 's were calculated for olivine-whole rock pairs known to be in equilibrium from Mg-Fe K_D data, and are presented in Table 7.

For basalt groups A and B, K_D 's fall within the equilibrium range of 2.0 - 2.6 (Sato, 1977; Hart & Davis, 1978; le Roex, 1980) for their MgO contents (6.5 - 9.0 wt%). Olivines from basalt Group C however produce K_D 's that are below the equilibrium range (mean K_D = 1.45) even at Hart and Davis' (1978) highest MgO content. This reinforces the suggestion that the Group C basalts are enriched in cumulus olivine. Figure 17 shows NiO vs Fo content. Olivines from groups A, B and C fall along a single (diffuse) trend, while those from Group D define a similar trend, but with lower NiO at a given forsterite content. This suggests a distinct parental magma for the Group D basalt, while the Group A, B and C basalts were conceivably derived from the same or similar parental magma.

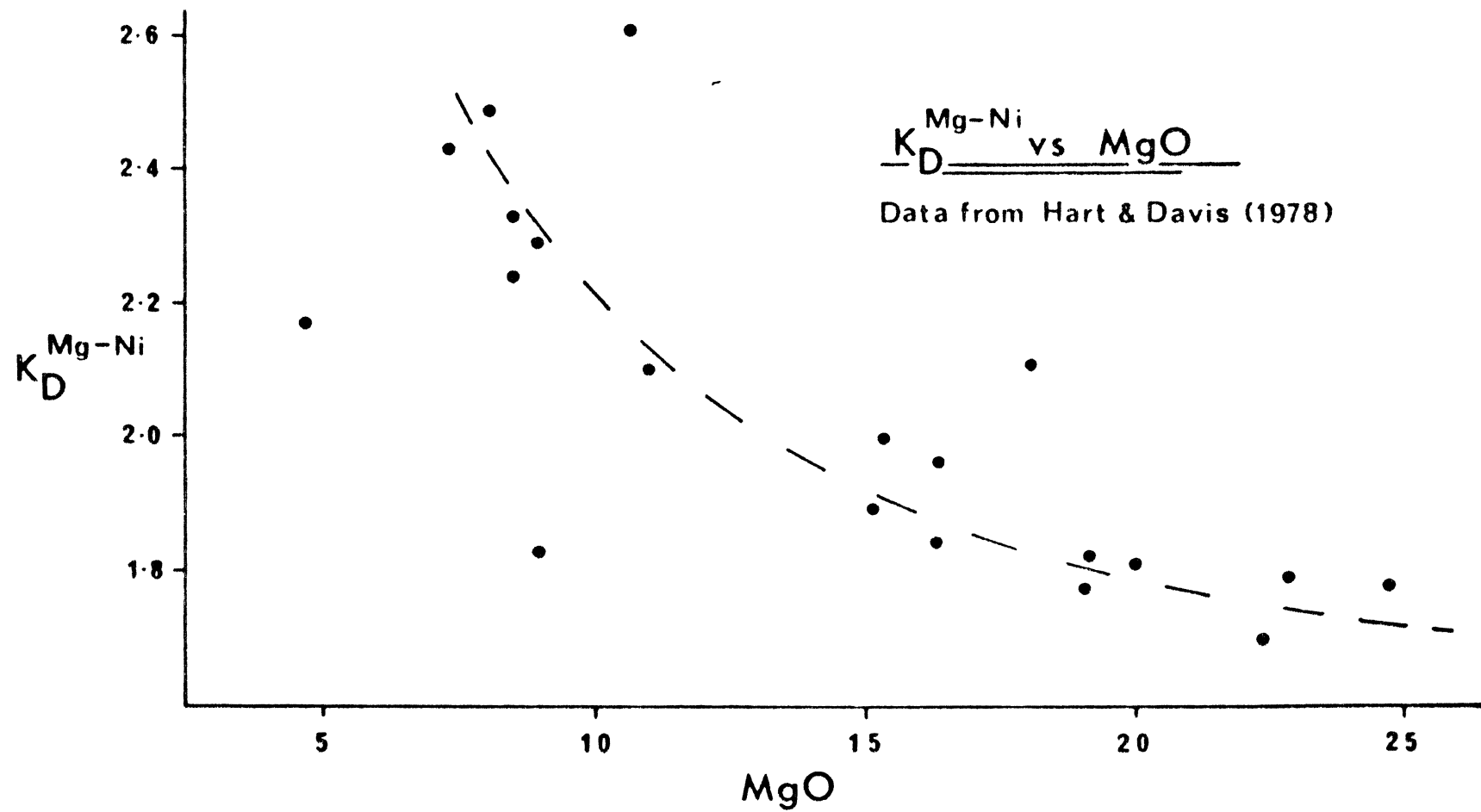


Figure 16. Distribution coefficient (K_D) of Mg - Ni in olivines vs MgO content of magma, showing the dependence of K_D on MgO. Data from Hart and Davis (1978). From le Roex (1980).

	Sample	Mg-Ni Distribution Coefficient between Olivine and Melt $\pm 2\sigma$.	
GROUP A	AG32-3-18	1.91 $\pm$ 0.32	
	AG32-3-50	2.83 $\pm$ 0.72	
	AG32-3-74	2.69 $\pm$ 0.26	
	AG32-3-75	2.38 $\pm$ 0.24	Mean $K_D^{Mg-Ni} = 2.46$
	AG32-3-86	2.61 $\pm$ 0.38	
	AG32-3-97	2.20 $\pm$ 0.34	
	AG32-3-100	2.59 $\pm$ 0.92	
GROUP B	AG32-3-63	2.41 $\pm$ 0.22	
	AG32-3-94	2.23 $\pm$ 0.35	
	AG32-3-99	2.57 $\pm$ 0.18	
	AG32-3-105	2.54 $\pm$ 0.28	Mean $K_D^{Mg-Ni} = 2.33$
	AG32-3-108	1.85 $\pm$ 0.36	
	AG32-3-111	2.34 $\pm$ 0.40	
	AG32-3-113	2.40 $\pm$ 0.81	
GROUP C	AG32-3-36	1.49 $\pm$ 0.30	
	AG32-3-40	1.41 $\pm$ 0.20	Mean $K_D^{Mg-Ni} = 1.45$
GROUP D	AG32-3-5	2.62	

Table 7: Calculated Mg-Ni distribution coefficients between olivines in equilibrium with whole rock composition and the whole rock composition. Samples with K_D^{Mg-Ni} values of 2.0 - 2.6 fall within the equilibrium range defined by Sato (1977) and by Hart and Davis (1978).

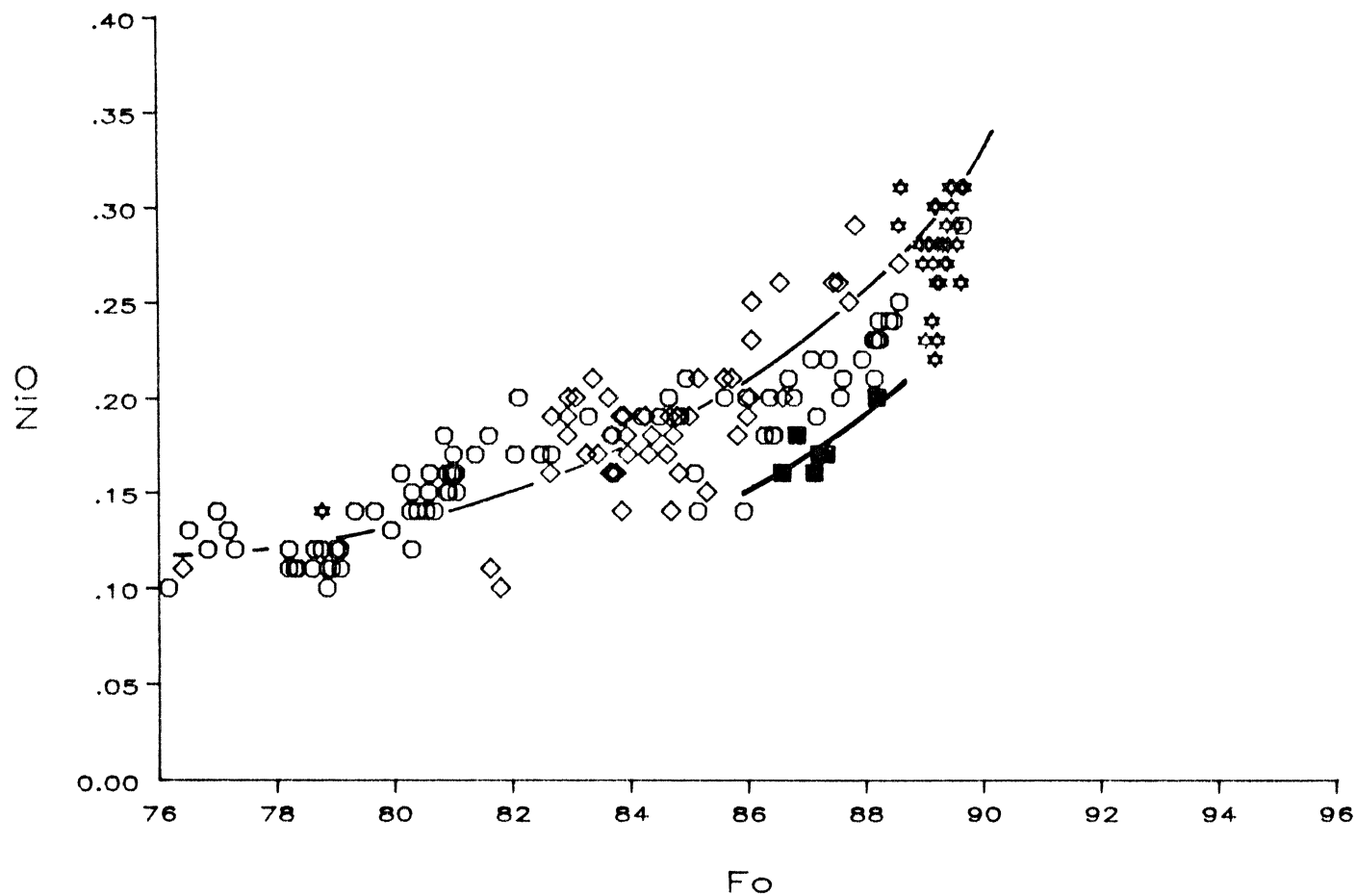


Figure 17. NiO content of olivine phenocryst cores vs Fo content. Symbols as per Figure 22 except that Group D are filled squares for clarity. The top trend is for basalt groups A, B and C, while the lower trend is for the Group D olivines alone. This is illustrative of a distinct parental magma for Group D.

However, the scatter in the data is such that a single parental magma for all basalts is not negated.

Plagioclase

Four modes of crystallization of plagioclase in the Agulhas 32, dredge 3 samples exist. These are phenocrysts or microphenocrysts; quench rims on phenocrysts; as microlites, and as highly skeletal quench grains embedded in the groundmass of the basalt. Representative plagioclase analyses are presented in Table 8. The slowly cooled phenocrysts are considerably more calcic than rapidly cooled quench rims and microlites. This compositional distinction is evident in all basalt groups and is illustrated in figure 18.

Phenocrysts are generally more calcic than An_{75} , with the most calcic phenocrysts ($An_{90.5}$) occurring in Group A (the evolved, sparsely plagioclase phyric) basalts. The majority of phenocrysts from groups A, B and D are compositionally similar, with cores of An_{85} . Phenocryst cores from Group C basalts are slightly less anorthitic ($An_{81} - An_{86}$).

Despite the similarity of phenocryst core compositions, there are large differences in the mean quench plagioclase composition. The most albitic quench feldspars (An_{65}) are found in Group A basalts, closely followed by quench plagioclase from Group C basalt (mean $An = 67\%$) and Group B basalt (mean $An = 73\%$). The most calcic quench plagioclase

	GROUP A				GROUP B				GROUP C				GROUP D		
	(1)	(2)	(3)	(4)	(1)	(2)	(3)	(4)	(1)	(2)	(3)	(4)	(1)	(2)	(3)
SiO ₂	57.40	47.58	53.64	49.18	56.04	46.50	52.10	48.26	55.30	47.43	53.99	49.11	48.76	47.03	47.79
Al ₂ O ₃	26.16	33.82	28.81	32.49	27.52	33.54	30.70	32.40	27.79	33.70	29.16	32.49	32.00	33.59	32.86
FeO*	1.39	.37	0.79	0.47	.61	.39	0.43	0.40	1.23	.36	0.87	0.35	.29	.24	.32
MgO	.19	.15	0.25	0.22	.16	.19	0.27	0.23	.33	.22	0.28	0.25	.23	.21	.26
CaO	9.84	17.56	12.64	16.32	11.09	17.52	14.06	16.59	11.45	17.47	12.93	16.45	16.29	17.87	17.27
Na ₂ O	5.21	1.07	3.97	1.81	4.68	1.16	2.84	1.60	4.29	1.19	8.75	1.74	2.13	1.34	1.57
K ₂ O	.10	ND	0.05	ND	.04	ND	0.04	ND	.08	ND	0.05	ND	ND	ND	ND
TOTAL	100.29	100.55	100.15	100.49	100.14	99.30	100.44	99.48	100.47	100.37	101.03	100.39	99.70	100.28	100.07

ATOMIC PROPORTIONS CALCULATED ON THE BASIS OF 8 OXYGENS

Si	2.577	2.171	2.429	2.283	2.520	2.153	2.354	2.221	2.489	2.169	2.424	2.229	2.239	2.157	2.193
Al	1.384	1.819	1.538	1.743	1.458	1.830	1.635	1.757	1.474	1.816	1.543	1.756	1.732	1.816	1.777
Fe	0.052	0.014	0.030	0.018	0.023	0.015	0.016	0.016	0.046	0.014	0.032	0.013	0.011	0.009	0.012
Mg	0.013	0.010	0.017	0.015	0.010	0.013	0.018	0.016	0.022	0.015	0.019	0.017	0.016	0.015	0.018
Ca	0.474	0.858	0.613	0.796	0.534	0.869	0.680	0.818	0.552	0.856	0.622	0.800	0.802	0.878	0.849
Na	0.453	0.094	0.348	0.160	0.408	0.104	0.249	0.143	0.374	0.105	0.327	0.153	0.189	0.119	0.140
K	0.006	-	0.003	-	0.002	-	0.002	-	0.005	-	0.003	-	-	-	-
An	50.79	90.02	63.58	83.21	56.57	89.28	73.04	85.05	59.29	88.97	65.33	83.95	80.84	87.97	85.80
Ab	48.61	9.89	36.13	16.73	43.21	10.72	26.75	14.84	40.20	10.96	34.34	16.05	19.09	11.91	14.11
Or	0.60	0.09	0.29	0.06	0.21	-	0.21	0.11	0.51	0.08	0.33	-	0.07	0.12	0.09

1 = Most albitic microlite
2 = Most anorthitic phenocryst

3 = Representative microlite
4 = Representative phenocryst

1 = Most albitic microlite
2 = Most anorthitic phenocryst

3 = Representative microlite
4 = Representative phenocryst

1 = Most albitic microlite
2 = Most anorthitic phenocryst

3 = Representative microlite
4 = Representative phenocryst

1 = Most albitic plagioclase
2 = Most anorthitic plagioclase

3 = Representative plagioclase analyses

Table 8: Representative plagioclase analyses from AG32-3 basalt groups. All iron as FeO

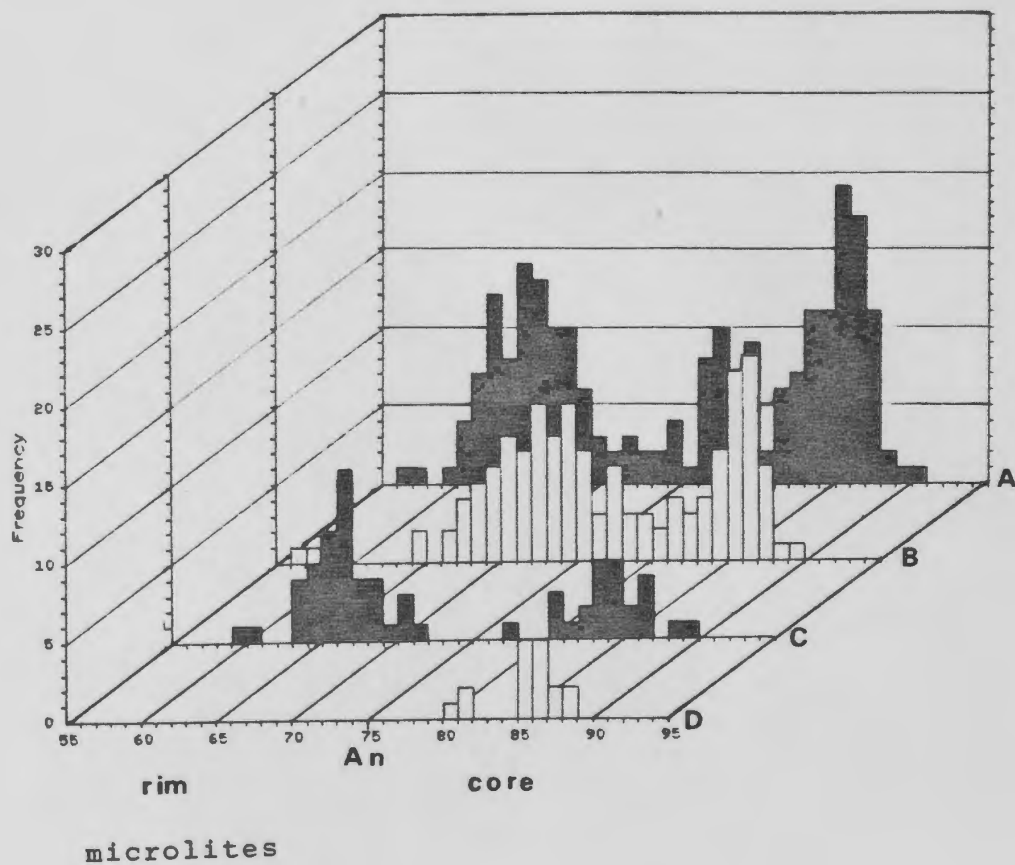


Figure 18. Plagioclase composition. Phenocryst cores are characterised by Anorthite contents $> 75\%$, while microlites and quench rims on phenocrysts are characterised by An contents $< 75\%$.

composition (An_{81}) occurs in Group D basalt. The mean orthoclase content is $<0.2\%$ in Groups A and C, and below detection limit in Groups B and D. Other minor elements which may be present in limited amounts in plagioclase include Fe, Mn, Mg and Sr.

The variation in quench plagioclase composition can be ascribed to the different quenching temperatures for different basalts. Quench temperatures were calculated for plagioclase microlite quench glass pairs for selected basalts, using the Mathez (1973) refinement of the Kudo and Weill (1970) geothermometer for basaltic rocks. These, as well as minimum estimates of the liquidus plagioclase temperature (calculated using the same geothermometer, from the most calcic plagioclase phenocryst core whole rock composition pair), are presented in Table 9. Quench temperatures range from 1103°C - 1125°C in Group A basalts; from 1112°C - 1114°C in the Group B basalts, and from 1049°C - 1083°C in the Group C basalts. A calculated quench temperature for the Group D basalt is 1161°C , which may (if the sample represents a primitive high alumina basalt), or may not (if the sample is a plagioclase cumulate) be meaningful. The greatest cooling range (between temperature of first plagioclase crystallization and quenching) of 98°C occurs in the Group C basalt AG32-3-36. A number of basalts show extremely limited intervals over which plagioclase was crystallizing (0°C in AG32-3-39; 3°C in AG32-3-50). Errors on these temperatures are estimated at $\pm 20^{\circ}\text{C}$ from the differences between calculated and experimental temperatures (Mathez, 1973).

	Sample No.	Liquidus Temperature	Quenching Temperature	Δ T°
GROUP A	AG32-3-2	1193°C	1103°C	90°
	AG32-3-18	1180°C	1107°C	73°
	AG32-3-32	1168°C	1125°C	43°
	AG32-3-50	1101°C	1104°C	-3°
GROUP B	AG32-3-63	1177°C	1113°C	64°
	AG32-3-105	1148°C	1114°C	34°
GROUP C	AG32-3-36	1147°C	1049°C	98°
	AG32-3-39	1080°C	1083°C	-3°
GROUP D	AG32-3-5	1210°C	1161°C	49°

Table 9: Plagioclase geothermometry: Temperatures are calculated using the Mathez (1973) refinement of the Kudo-Weill (1970) plagioclase thermometer. Liquidus temperatures calculated for the most anorthitic phenocryst core-whole rock pair; Quenching temperature calculated from plagioclase microlite - quench glass pair.

Bryan (1974) investigated the Fe-Mg relationships in plagioclase from ocean floor basalt. From stoichiometric considerations he concluded that Fe^{2+} and Mg^{2+} substitute for Al in the plagioclase structure, in a $\text{Ca}(\text{Mg},\text{Fe})\text{Si}_3\text{O}_8$ type formula unit. This substitution is reportedly favoured by decreasing temperature (Bryan, 1974). Similar conclusions have been reached by Crawford (1973), Neibuhr et al (1973), Wenk & Wilde (1973) and Longhi et al (1976). However, le Roex (1980) found that in plagioclases from the Islas Orcadas fracture zone, Fe and Mg substitute in different sites. In the latter case the use of Fe/Mg ratios in plagioclase as indicators of magma crystallization conditions (Longhi et al, 1984) is invalidated by the differing crystallo-chemical effects of the different sites in the plagioclase structure.

The variation in FeO and MgO with An content in plagioclase is shown in figure 19. Although there is considerable scatter in the data, there are definite trends of FeO and MgO decreasing with increasing anorthite content. The more rapid decrease in FeO than MgO with increasing An, results in a decreasing Fe/Mg ratio with increasing An (Figure 19). The Fe/Mg ratio in plagioclase is generally regarded as reflecting the Fe/Mg ratio of the magma at the time of crystallization, an interpretation consistent with the observed increasing ratio with increasing albite content.

To investigate Fe, Mg site occupancy in plagioclase, the approach used by Bryan (1974) and le Roex (1980) has been adopted, and is detailed below.

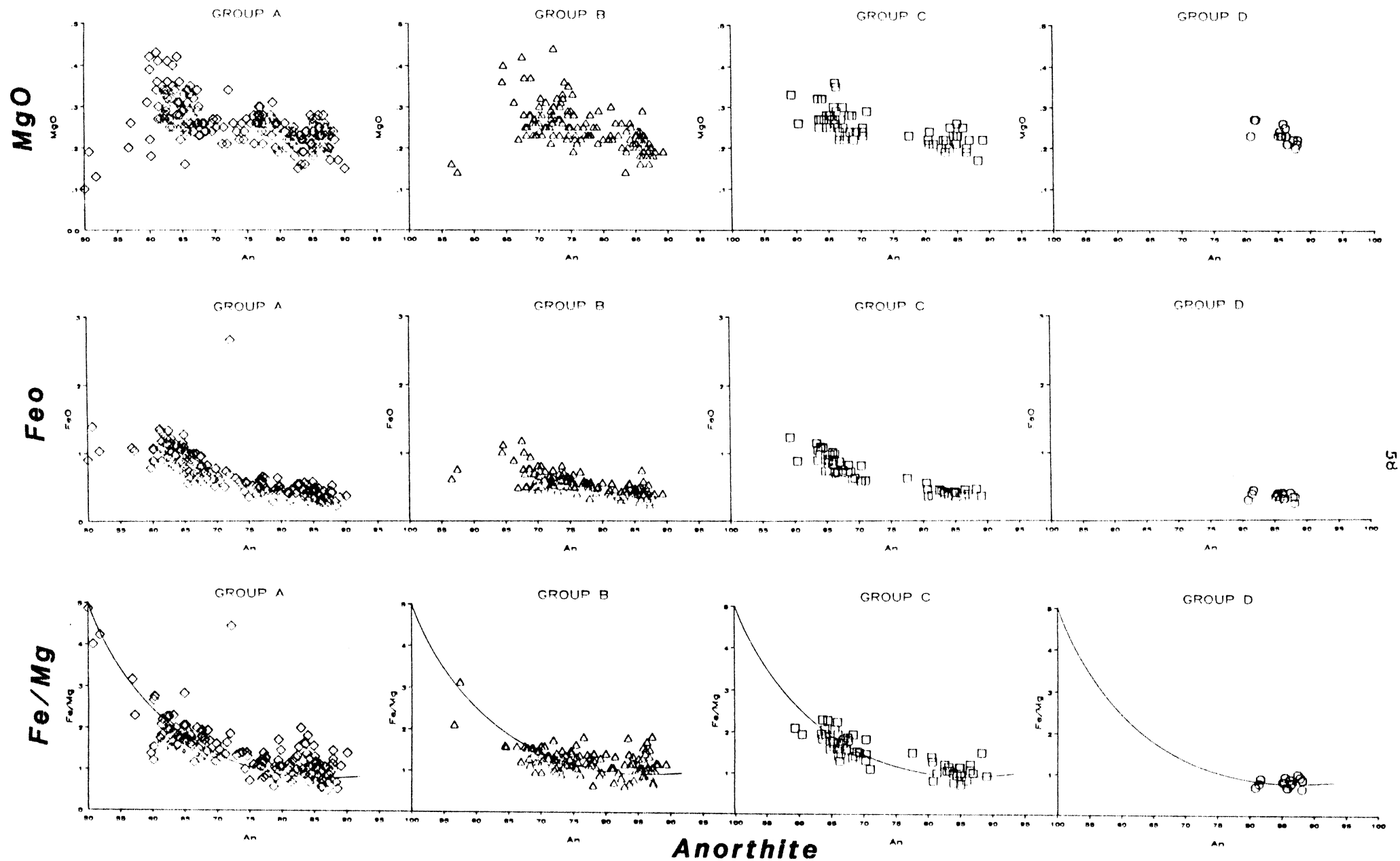


Figure 19. FeO, MgO and Fe/Mg in plagioclase vs Anorthite content. Fe, Mg calculated on the basis of 8 oxygens. FeO and MgO decrease with increasing An content, as does Fe/Mg which illustrates the more rapid depletion of Fe than Mg.

A variety of substitutions can be utilized to balance charge in the plagioclase structure. In the anorthite end-member the substitution is of the type $R^{2+}(R^{3+})_2Si_2O_8$, while in the albite end member the common substitution is of the $R^+R^{3+}Si_3O_8$ (or $R^{2+}R^{2+}Si_3O_8$) type.

Ideally if all Ca is combined in anorthite type formula units, the following two expressions will be fulfilled:

$$Ca = Al - 1.0 \quad Ca = -1.0 * Si + 3.0$$

Substitution of Fe and/or Mg can theoretically occur for either Ca in an anorthite type molecule, resulting in the following relations:

$$Ca + Mg + Fe = Al - 1.0 \quad Ca + Mg + Fe = -1.0 * Si + 3.0$$

or, for Al in an albite type molecule, where

$$Ca = (Al + 2Fe + 2Mg) - 1.0 \quad Ca = -1.0(Si - Fe - Mg) + 3.0$$

(Doubling the amounts of Fe and Mg is necessary to satisfy the expected relation for pure anorthite-albite mixtures (le Roex, 1980)).

Using the above relations, Bryan (1974) concluded that Fe and Mg substitute for Al in an albite type formula unit, satisfying the relations;

$$Ca = (Al + 2Mg + 2Fe) - 1.0 \quad Ca = -1.0(Si - Fe - Mg) + 3.0$$

Manipulation of the plagioclase data from all four basalt groups from Ag32, dredge 3 reveals possible Fe, Mg substitutions in these rocks. Figure 20 a & b shows variation in Al and Si relative to Ca for the four basalt groups. These are analagous to Bryan's (1974) findings in that there is an excess of Ca over Si and Al. Assuming all Fe to be divalent, and substituting with

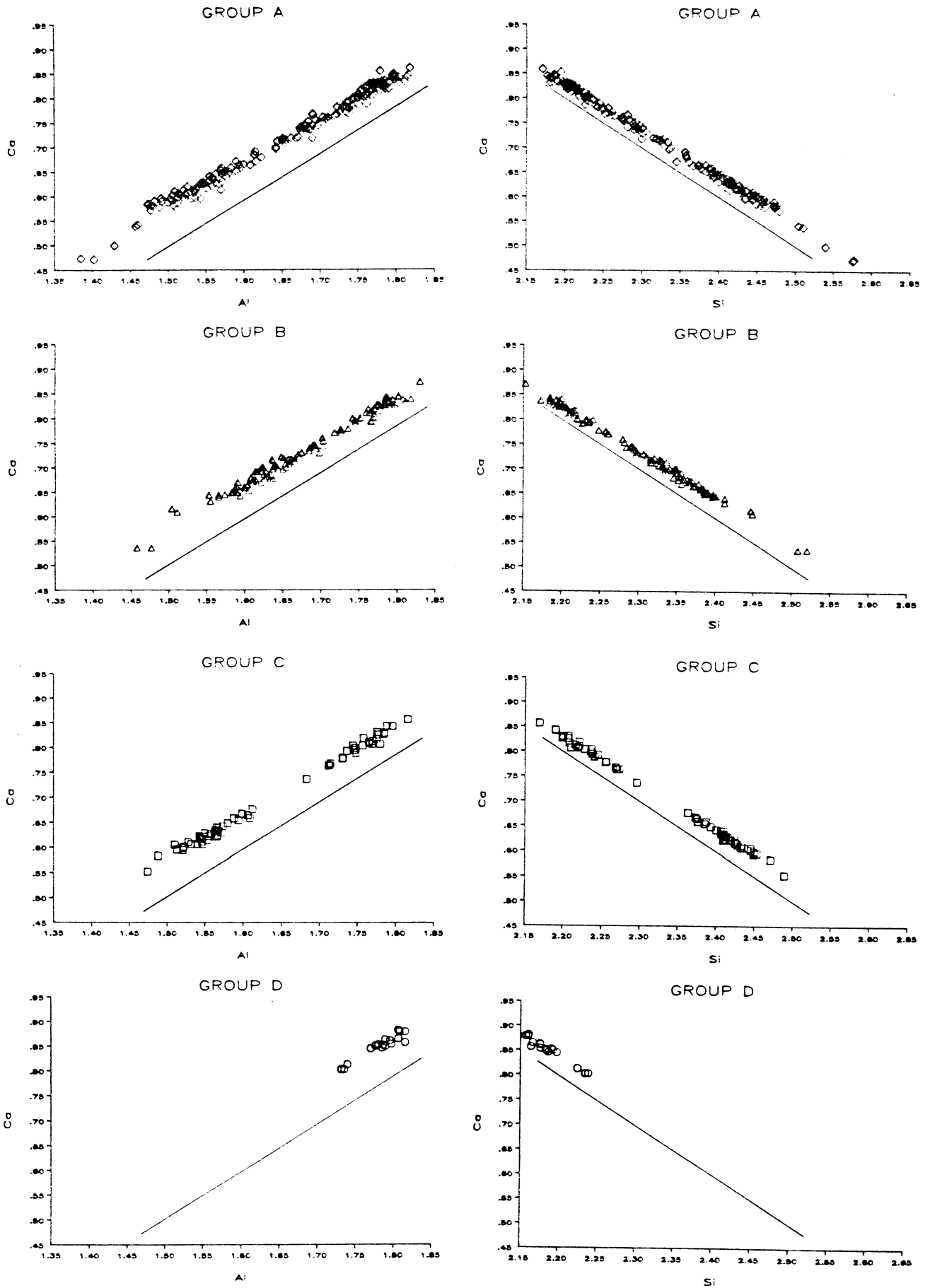


Figure 20 (a) Ca vs Al (b) Ca vs Si
for plagioclase crystals. Expected trends are those for pure Anorthite in which $Ca = Al - 1$ and $Ca = -Si + 3$.

Mg for Al into an albite type formula unit (Figure 20, c & d) shows that data from the Group D basalts fit the expected trend perfectly, while a less than perfect fit is found for groups A, B & C. Assuming all Fe to be divalent and substituting for Al, with Mg substituting for Ca only exacerbates the situation (Figure 20, e & f), with data plotting far off the expected trend.

If all Fe is assumed to be trivalent in groups A, B & C plagioclases, and substituting for Al, with Mg substituting into a $\text{CaMgSi}_3\text{O}_8$ molecule the following relation is very closely adhered to (Figure 20, g);

$$\text{Ca} = \text{Al} + \text{Fe} + 2\text{Mg} - 1.0$$

In order to test whether some trivalent Fe substituted directly for Al, with some divalent Fe substituting for Al in a substitution coupled with the Ca-Mg substitution, as found by le Roex (1980), $\text{Al} + 1.5\text{Fe}$ & $\text{Si} - 0.5\text{Fe}$ were plotted against Ca and Mg (Figure 20, h & i). This assumes half the Fe to be divalent, and half to be trivalent. It is clearly evident from figure 20(h & i) that the substitution of Fe as Fe^{2+} and Fe^{3+} is not likely to have occurred in any of the basalts from AG32, dredge 3.

From the above it is evident that only in plagioclase from the Group D basalt, do Fe and Mg substitute into the same site in the plagioclase structure, and hence only in this basalt may the Fe/Mg ratio found in the plagioclase ($1.861: \sigma = 0.222, n = 17$) be used to infer the magma Fe/Mg ratio at the time of plagioclase crystallization. Due to the different crystallochemical effects of the tetrahedral (Al) site and the 8

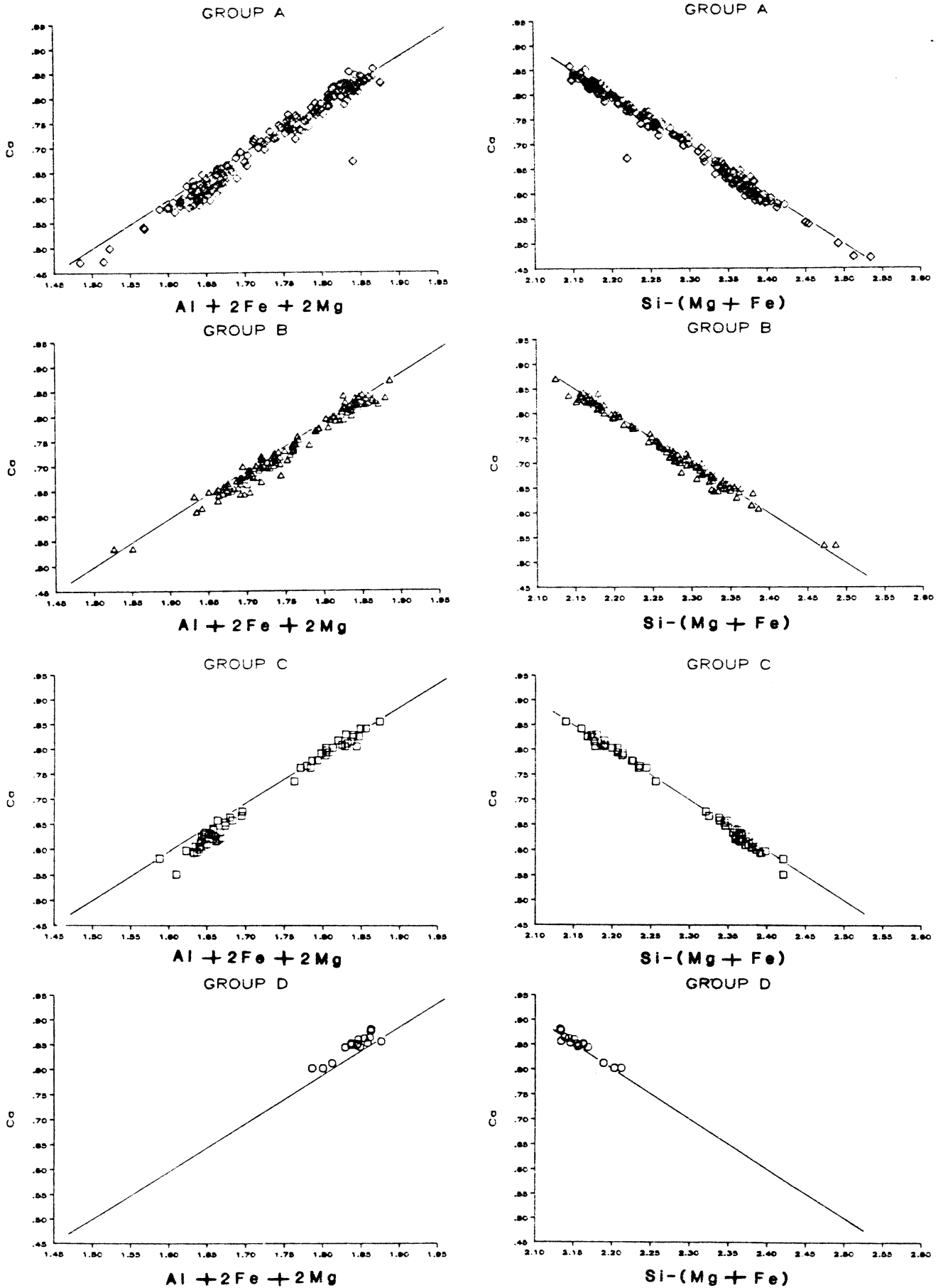


Figure 20 (c) Ca vs $Al + 2Fe + 2Mg$ (d) Ca vs $Si - (Mg + Fe)$
 Expected trends are those for all Fe and Mg substituting for Al in
 an Albite type formula unit.

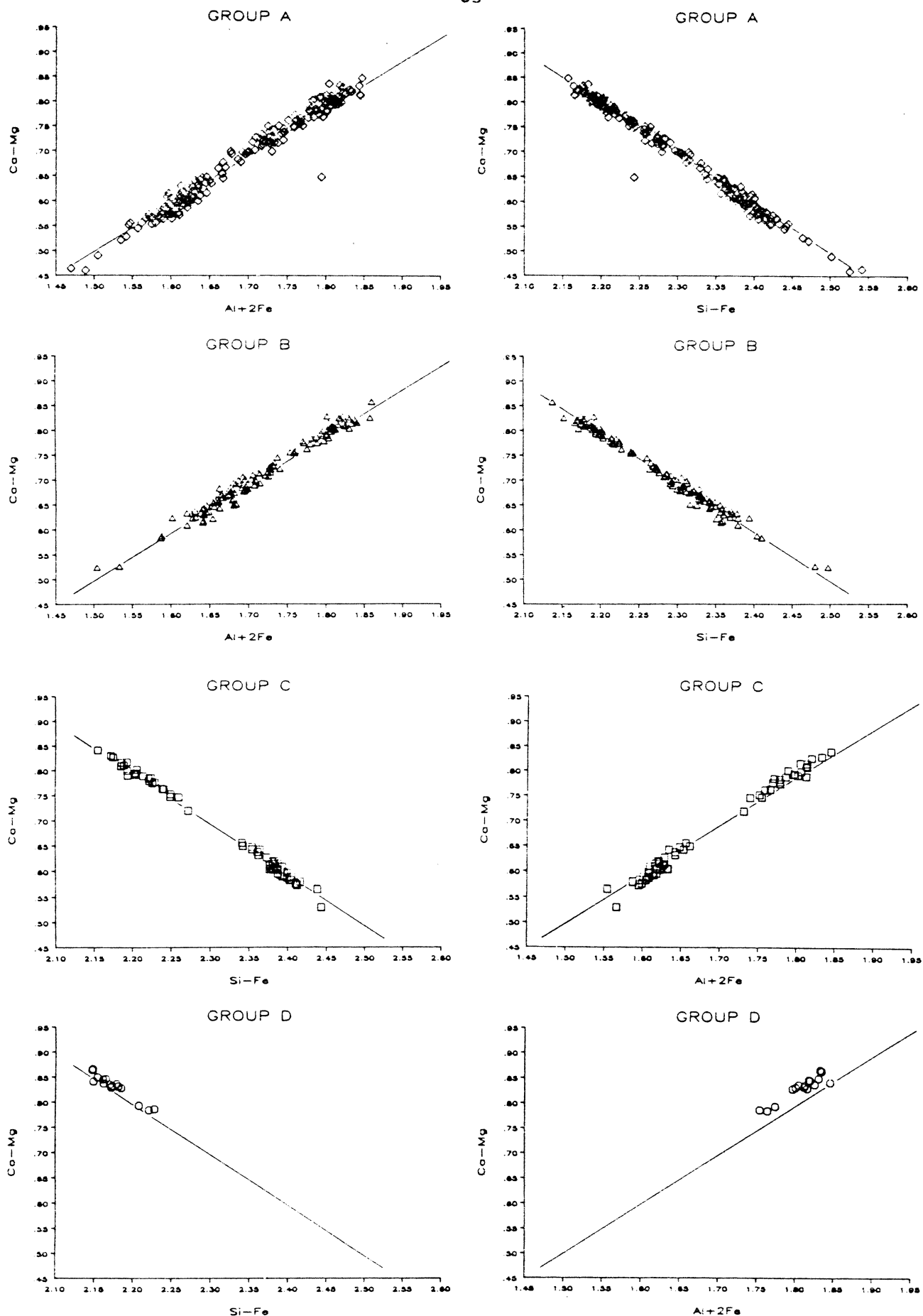


Figure 20 (e) Ca - Mg vs Si - Fe

(f) Ca - Mg vs Al + 2Fe

Expected trends are those for all Fe substituting for Al and all Mg substituting for Ca.

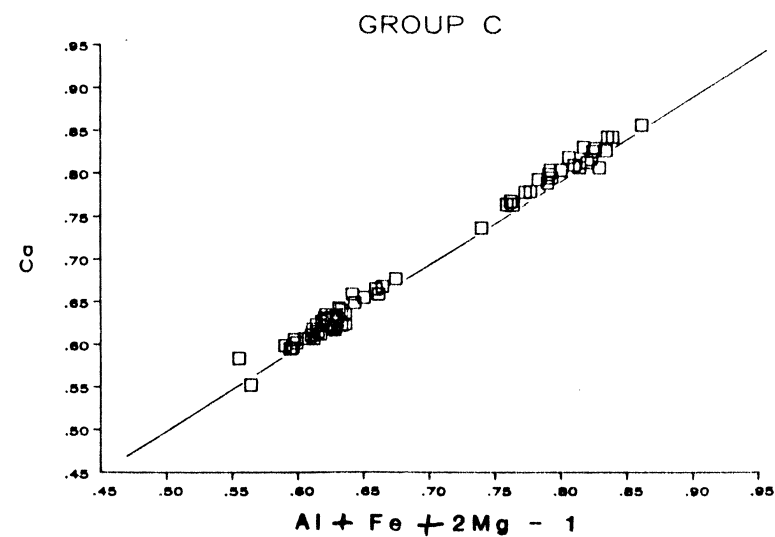
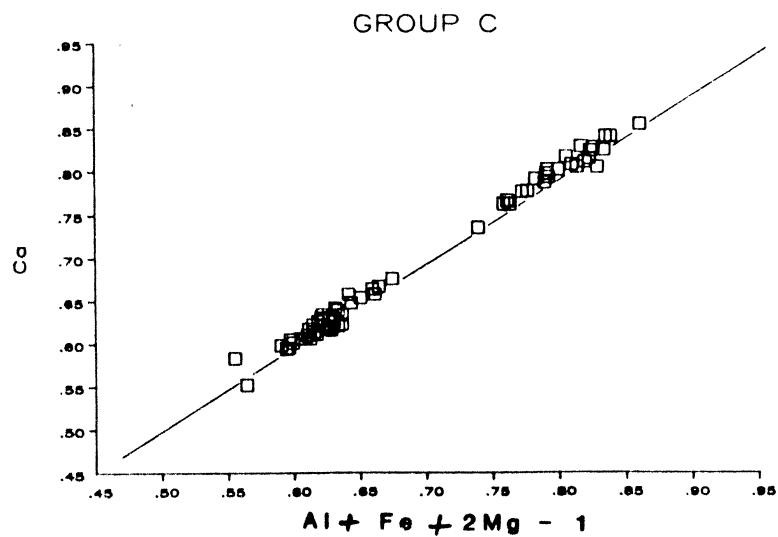
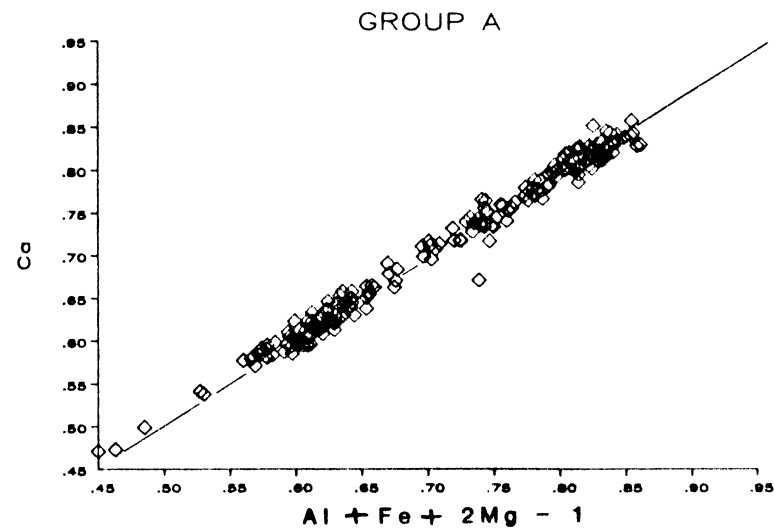
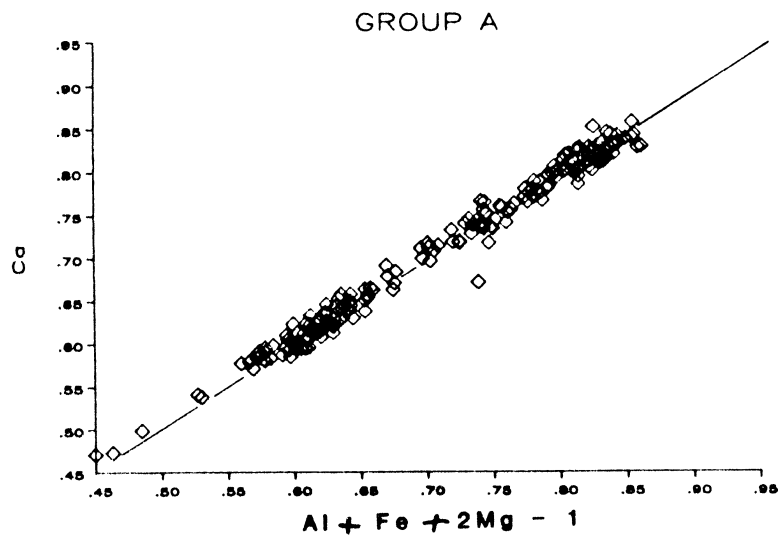


Figure 20 (g), Ca vs Al + Fe + 2Mg - 1.

Expected trend is that for all Fe being trivalent and substituting for Al,
while Mg substitutes into a $\text{CaMgSi}_3\text{O}_8$ molecule.

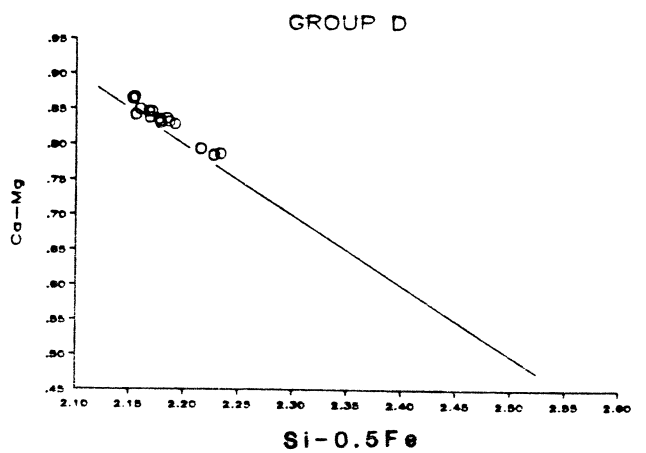
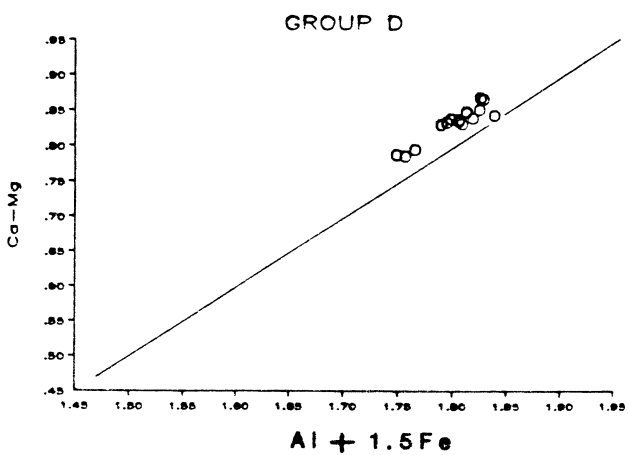
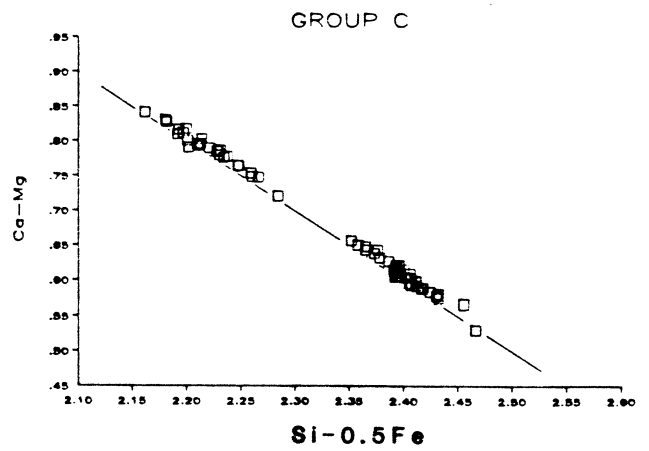
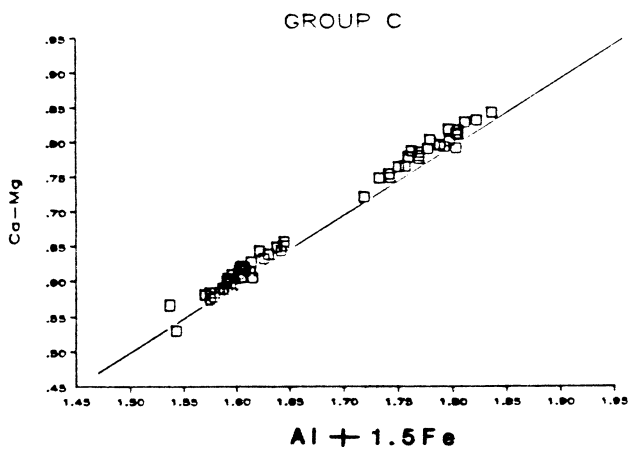
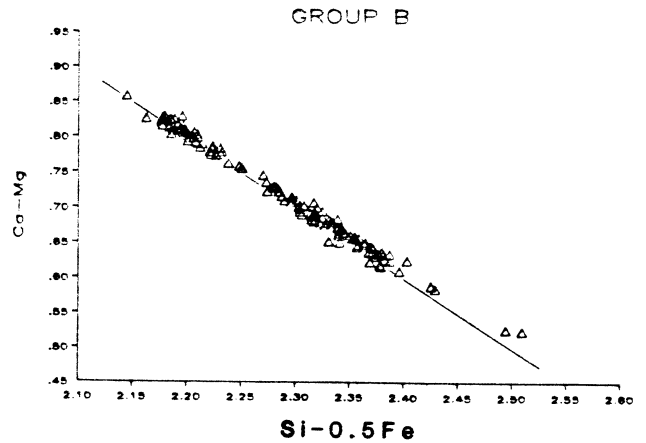
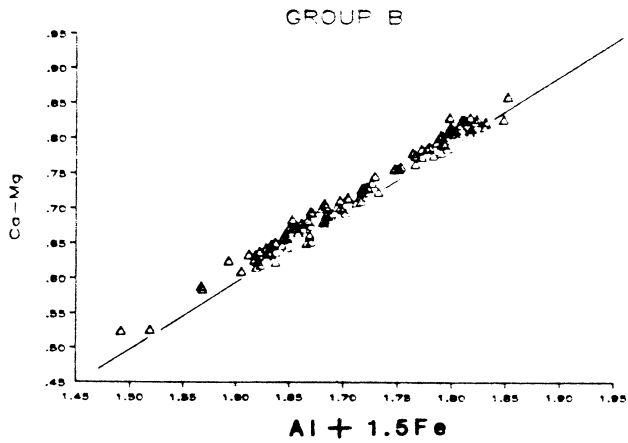
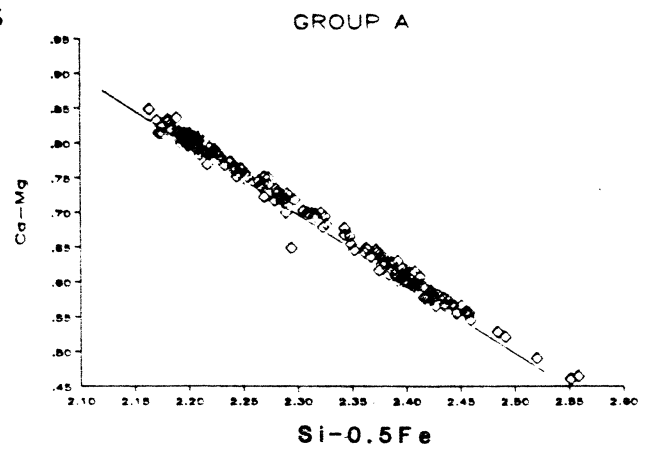
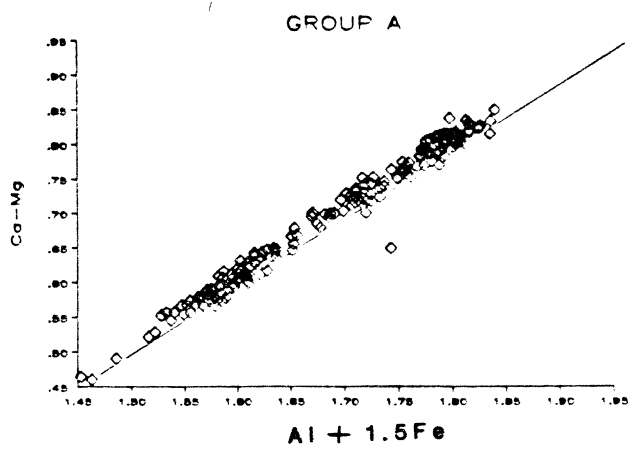


Figure 20 (h) Ca - Mg vs Al + $1\frac{1}{2}$ Fe

(i) Ca - Mg vs Si - $\frac{1}{2}$ Fe

Expected trends are those for some trivalent Fe substituting directly for Al, in a substitution coupled with the Ca - Mg substitution. This assumes half Fe to be trivalent and half divalent.

coordinated Ca site, the Fe/Mg ratio of plagioclases from Group A,B & C basalts are unlikely to directly reflect the Fe/Mg ratio of their source magma(s).

Spinel

Microphenocrysts of Cr-spinel occur in four samples from the highly olivine phyric (Group C) basalts; AG32-3-30, AG32-3-39, AG32-3-40 and AG32-3-46. Mean spinel analyses are presented in Table 10.

Cr-spinels are characterised by Al_2O_3 contents of 25 - 32 wt% and Cr_2O_3 contents of 33 - 42 wt%, and considerable variation in composition is found, even between the spinels found in a single section. Figure 21 compares Cr-spinels from the Group C basalts with those from normal MORB and cumulate rocks (Siggurdson and Shilling, 1976; Basaltic Volcanism Study Project, 1981) and illustrates the Cr rich, Mg deficient nature of these spinels in comparison with those from normal MORB. Slight zoning in a Cr-spinel microphenocryst in AG32-3-40 is evident with increasing Al_2O_3 (27.11 - 28.10 wt%) and decreasing Cr_2O_3 (39.54 - 38.18 wt%) from core to rim.

Chrome-spinel is a liquidus or co-liquidus phase in oceanic basalts and experimental studies have identified a compositional dependence on oxygen fugacity (Hill and Roeder, 1974; Fisk and Bence, 1979) which can potentially provide additional constraints on the genesis of possible parental magma(s).

	AG32-3-30		AG32-3-39	AG32-3-40		AG32-3-46	
SiO ₂	ND	ND	ND	ND	ND	ND	ND
TiO ₂	0.44	0.42	0.46	0.46	0.42	0.50	0.49
Al ₂ O ₃	31.80	25.16	26.74	26.83	32.12	27.84	27.43
Cr ₂ O ₃	33.69	42.05	39.14	39.78	33.51	37.45	37.36
FeO	15.67	14.65	14.69	14.40	16.03	15.81	17.64
MnO	0.17	0.21	0.15	0.15	0.17	0.15	0.27
MgO	16.80	16.16	17.38	17.18	16.70	16.72	15.11
CaO	ND	ND	ND	ND	0.03	ND	ND
TOTAL	98.57	98.65	98.56	98.80	98.98	98.47	98.30

ATOMIC PROPORTIONS CALCULATED ON THE BASIS OF 4 OXYGENS

Si	-	-	-	-	-	-	-
Ti	0.009	0.009	0.011	0.010	0.001	0.011	0.011
Al	1.109	0.902	0.950	0.950	1.116	0.989	0.987
Cr	0.785	1.012	0.933	0.945	0.781	0.893	0.901
Fe	0.388	0.373	0.370	0.362	0.395	0.399	0.450
Mn	0.004	0.006	0.004	0.004	0.004	0.004	0.007
Mg	0.742	0.728	0.781	0.770	0.734	0.752	0.687
Cu	-	-	-	-	0.001	-	-
Sum	3.037	3.030	3.049	3.041	3.040	3.048	3.043
FeO	11.24	11.15	9.61	9.99	11.53	10.80	13.00
Fe ₂ O ₃	4.92	3.90	5.65	4.90	5.00	5.57	5.16
REVISED TOTAL	99.07	99.04	99.12	99.26	99.48	99.04	98.83

Table 10: Cr-Spinel analyses illustrating compositional variability.

However this is not applicable to the Group C lavas, as their cumulate nature indicates that they do not represent liquid compositions. The Group C Cr-spinels plot with spinels from other cumulate rocks (see Figure 21). An olivine - spinel equilibration temperature calculated from a Cr-spinel adjacent to an olivine from AG32-3-40, using the Roeder (1979) modification of the Jackson (1969) olivine-spinel geothermometer, is excessively high (1918°C). This indicates that the Group C spinels are not in equilibrium with the olivine compositions, and thus supports the hypotheses that the Group C basalts are enriched in cumulate olivine.

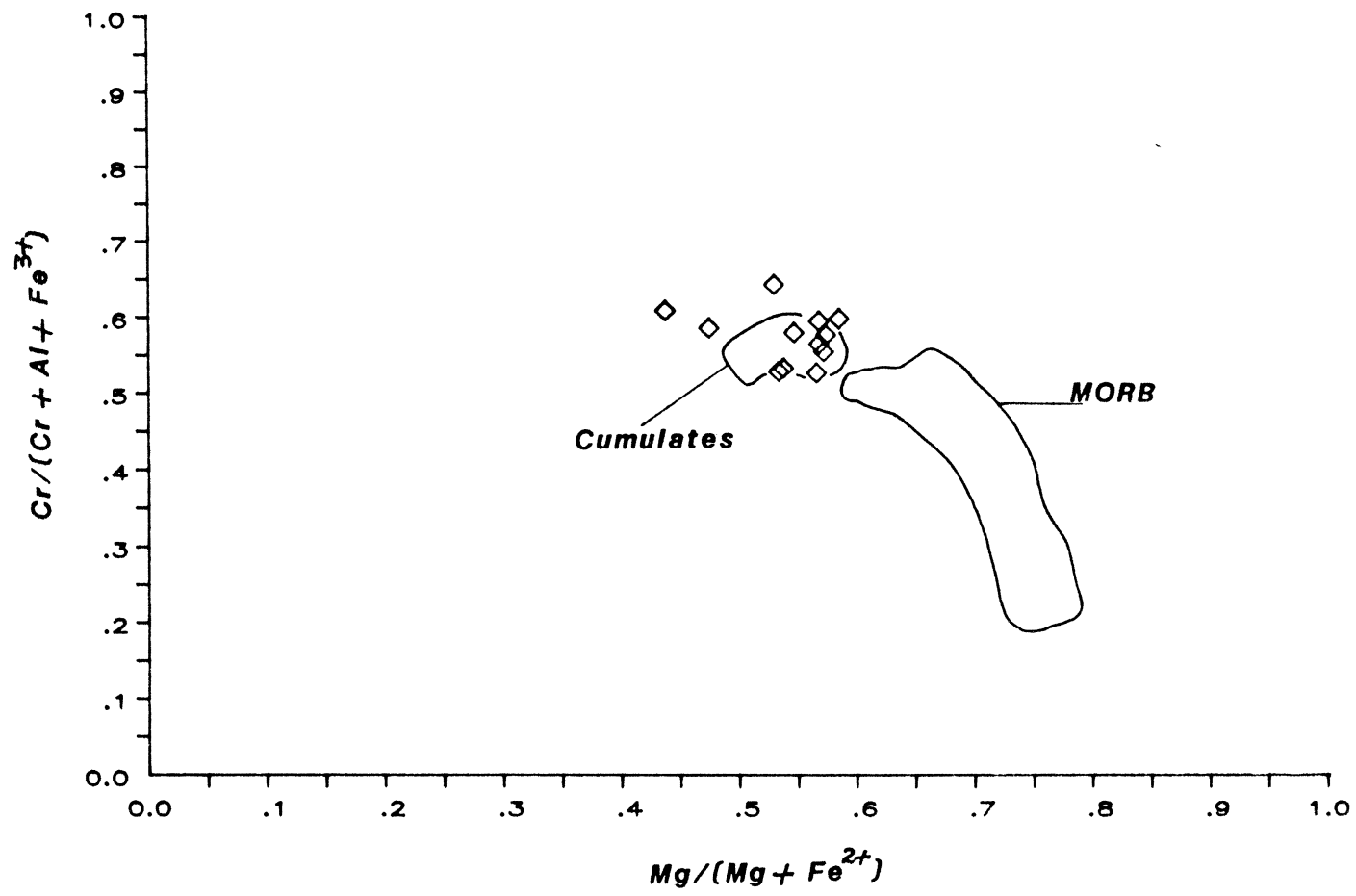


Figure 21. $Cr/(Cr + Al + Fe^{3+})$ vs $Mg/(Mg + Fe^{2+})$ in Cr-Spinels from Group C basalts. MORB and cumulate fields from Basaltic Volcanism Study Project (1981) and from Sigurdsson and Schilling (1976).

Chapter Four

W H O L E R O C K A N D G L A S S G E O C H E M I S T R Y

A combination of petrography and whole rock geochemistry can be used to classify the basalts of AG32, dredge 3 into 4 different groups These are;

Group A: Evolved sparsely plagioclase phyric basalts

Group B: Sparsely plagioclase phyric basalts

Group C: Olivine phyric basalts

Group D: Plagioclase phyric basalt

and their most important characteristics are summarised in Table 11. Sparsely plagioclase phyric groups (A & B) are petrographically similar, but are readily distinguished on the basis of whole rock composition.

Individual whole rock and glass analyses, Mg# (atomic $\text{Mg}/\text{Mg}+\text{Fe}^{2+}$ calculated using $\text{Fe}_2\text{O}_3/\text{FeO} = 0.15$ for whole rock analyses), MI ($\text{FeO}+\text{Fe}_2\text{O}_3/\text{FeO}+\text{Fe}_2\text{O}_3+\text{MgO}$) and selected trace element ratios are presented in Table 12. CIPW norms for all samples, assuming $\text{Fe}_2\text{O}_3/\text{FeO} = 0.15$, are presented in Table 13.

GROUP	PETROGRAPHY	CHEMISTRY
A	Sparsely to highly Plagioclase Phyric Pillow Basalt	- Evolved ($Mg\# = 52.9 - 58.4$) - Slightly elevated TiO_2 (1.89 - 2.15 wt%)
B	Phyric to Sparsely Plagioclase Phyric Pillow Basalt	- Less evolved than Group A ($Mg\# = 59.2 - 66.7$) - Elevated Al (15.85 - 16.76 wt%) - Elevated Ca (11.43 - 12.27 wt%)
C	Olivine Phyric to highly Olivine Phyric Pillow Basalt	- High $Mg\#$ (58.5 - 73.0) - High MgO (8.39 - 15.54 wt%) - Elevated TiO_2 (1.75 - 2.12 wt%) - Very high Cr (502 - 1079 ppm) - High to very high Ni (173 - 653 ppm)
D	Highly Plagioclase Phyric Basalt	- Very High Al_2O_3 (19.35 wt%) - High $Mg\#$ (70.0) - Low TiO_2 (0.68%)

Table 11: Major characteristics of the different basalt groups from AG32-3.

G R O U P A

Sample No.	AG32-3-2	AG32-3-18	AG32-3-32	AG32-3-50	AG32-3-72	AG32-3-74	AG32-3-75	AG32-3-83	AG32-3-86	AG32-3-97	AG32-3-100	AG32-3-109	AG32-3-112
SiO ₂	49.52	49.31	48.88	50.11	48.99	49.44	49.44	48.75	49.85	49.40	48.99	48.83	48.98
TiO ₂	1.88	2.07	1.98	2.01	1.96	1.93	1.93	1.93	1.99	2.09	2.12	2.04	2.04
Al ₂ O ₃	14.75	15.58	14.93	14.58	14.75	14.47	14.62	14.93	14.43	14.53	14.35	14.47	14.50
Fe ₂ O ₃	1.46	1.47	1.57	1.50	1.53	1.50	1.49	1.49	1.47	1.55	1.57	1.54	1.55
FeO	9.73	9.95	10.43	10.01	10.19	10.01	9.92	9.91	9.80	10.33	10.45	10.27	10.36
MnO	0.20	0.17	0.19	0.20	0.20	0.19	0.18	0.19	0.20	0.21	0.20	0.22	0.20
MgO	8.00	6.39	7.41	7.37	7.24	7.00	7.01	6.80	6.89	6.49	6.97	6.82	6.93
CaO	10.73	10.96	10.87	10.84	10.93	10.84	10.86	11.09	10.88	10.86	10.72	10.83	10.85
Na ₂ O	2.77	3.01	2.56	2.65	2.62	2.83	2.84	2.71	2.82	2.86	2.73	2.71	2.67
K ₂ O	0.26	0.40	0.19	0.27	0.25	0.30	0.28	0.34	0.39	0.48	0.30	0.30	0.25
P ₂ O ₅	0.24	0.20	0.21	0.21	0.24	0.22	0.20	0.23	0.27	0.26	0.24	0.25	0.27
LOI	0.78	1.06	0.75	0.93	0.91	1.00	1.02	1.33	1.16	1.31	0.99	1.59	0.93
H ₂ O-	0.15	0.24	0.16	0.21	0.19	0.20	0.18	0.28	0.22	0.23	0.19	0.18	0.18
TOTAL	100.48	100.84	100.11	100.90	100.00	99.92	99.96	99.98	100.35	100.59	99.82	100.04	99.75
Zr	132	132	130	127	130	122	123	126	125	139	143	138	138
Nb	5.1	4.8	6.0	6.2	6.2	5.7	6.0	5.1	6.0	6.1	5.9	6.2	6.1
Y	46	47	49	48	48	44	44	48	46	51	53	53	52
Rb	5.3	8.6	5.0	5.4	5.7	6.2	5.3	7.7	8.8	9.8	6.1	5.9	6.0
Ba	43		41	45	41	37	36	40	47	51	45	42	45
Sr	114	119	98	112	98	118	116	102	114	103	102	104	102
Co	48		47	49	49	49	49	48	46	50	50	50	50
Cr	174		212	241	214	213	212	212	240	183	201	198	205
Ni	87	92	108	104	105	96	95	113	99	97	103	104	102
V	350		327	357	339	334	331	330	346	342	356	347	350
Zn	103	102	106	105	104	100	100	100	99	101	111	109	110
Cu	49	48	48	49	50	50	52	51	46	45	48	50	52
Sc	41		40	42	40	40	41	38	39	36	41	41	41
Zr/Nb	26	28	22	20	21	22	21	25	21	23	24	22	23
Zr/Y	2.9	2.8	2.7	2.9	2.7	2.8	2.8	2.7	2.7	2.7	2.7	2.6	2.6
Y/Nb	9.0	9.7	8.1	7.7	7.7	7.8	7.4	9.4	7.8	8.4	8.9	8.5	8.6
Ti/Zr	85	96	91	95	90	95	95	92	95	90	89	89	89
Ti/P	10.8	14.2	13.0	13.2	11.2	12.1	13.3	11.5	10.1	11.0	12.1	11.2	10.4
Mg#	59.4	53.4	55.9	56.8	55.9	55.5	55.7	55.0	55.6	52.8	54.3	54.2	54.5

Table 12: Major and trace element X.R.F. analyses of AG32-3 whole rocks. Fe₂O₃ / FeO = 0.15.

Sample No.	GROUP B							GROUP C				GROUP D	
	AG32-3-63	AG32-3-94	AG32-3-99	AG32-3-105	AG32-3-108	AG32-3-111	AG32-3-113	AG32-3-30	AG32-3-36	AG32-3-39	AG32-3-40	AG32-3-46	AG32-3-5
SiO ₂	48.73	48.30	48.68	48.67	48.86	49.13	48.59	48.95	47.65	48.48	47.10	49.47	48.51
TiO ₂	1.51	1.06	1.61	1.43	1.27	1.51	1.23	2.07	1.85	2.02	1.74	2.11	0.68
Al ₂ O ₃	15.79	16.52	15.67	15.92	16.21	15.91	16.32	14.00	12.50	13.74	11.71	14.41	19.30
Fe ₂ O ₃	1.36	1.18	1.41	1.33	1.26	1.36	1.26	1.56	1.55	1.55	1.53	1.54	0.96
FeO	9.05	7.85	9.39	8.84	8.43	9.03	8.39	10.40	10.33	10.32	10.22	10.25	6.40
MnO	0.17	0.16	0.17	0.19	0.17	0.18	0.18	0.20	0.19	0.18	0.18	0.21	0.13
MgO	7.84	8.81	7.63	7.74	8.07	7.44	8.70	9.16	13.62	10.10	15.48	8.11	8.38
CaO	11.45	11.97	11.30	11.59	12.19	11.60	11.80	9.98	8.89	9.77	8.36	10.35	13.13
Na ₂ O	2.60	2.51	2.64	2.49	2.52	2.68	2.48	2.63	2.69	2.57	2.85	2.68	2.09
K ₂ O	0.21	0.13	0.22	0.23	0.21	0.31	0.12	0.22	0.24	0.21	0.25	0.28	0.08
P ₂ O ₅	0.18	0.10	0.17	0.19	0.13	0.15	0.13	0.29	0.20	0.19	0.20	0.25	0.07
LOI	1.21	1.33	1.15	1.28	0.74	0.89	0.55	0.76	0.86	0.78	0.96	0.95	0.43
H ₂ O-	0.39	0.41	0.30	0.32	0.17	0.16	0.07	0.14	0.19	0.14	0.21	0.24	0.15
TOTAL	100.48	100.33	100.31	100.22	100.23	100.34	99.79	100.35	100.72	100.04	100.78	100.83	100.31
Zr	93	62	102	87	76	92	73	134	118	130	112	135	35
Nb	2.7	1.6	3.9	3.0	2.6	3.4	2.7	7.4	5.6	7.7	6.5	6.8	1.8
Y	37	26	39	34	30	35	30	46	41	45	40	47	17.5
Rb	3.6	2.8	4.2	4.0	3.7	6.2	2.9	5.1	5.8	4.8	5.7	6.1	1.3
Ba	25	14.6	29	22	21	27	21	51	47	50	44	52	8.9
Sr	122	132	119	124	130	121	125	113	102	111	97	116	95
Co	48	52	50	50	50	49	51	54	68	59	69	51	44
Cr	267	361	270	279	312	263	324	644	994	642	1075	500	359
Ni	127	185	133	129	148	124	150	240	524	310	651	172	136
V	272	212	281	269	242	270	239	345	318	340	249	346	156
Sn	84	68	88	83	76	83	79	105	99	104	98	104	53
Cu	60	76	58	61	67	61	62	52	46	51	45	53	67
Sc	42	40	40	41	40	40	40	40	37	40	32	40	32
Zr/Nb	35	38	27	29	29	27	27	18.0	21	16.7	17.2	18.0	19.6
Zr/Y	2.5	2.4	2.6	2.6	2.6	2.6	2.5	2.9	2.9	2.9	2.8	2.9	2.0
Y/Nb	13.8	15.7	6.4	11.1	11.7	10.3	10.9	6.2	7.3	5.8	6.2	6.9	9.8
Ti/Zr	98	102	95	98	100	98	101	95	94	93	93	94	117
Ti/P	11.5	14.6	13.0	10.3	13.4	13.8	13.0	9.8	12.7	14.8	12.1	11.6	13.3
Mg#	60.7	66.7	59.2	60.9	63.0	59.5	64.9	61.1	70.1	63.6	73.0	58.5	70.0

Table 12 (Cont.) Major and trace element X.R.F. analyses of AG32-3 whole rocks. Fe₂O₃ / FeO = 0.15

Sample									
No.	AG32-3-2	AG32-3-5	AG32-3-18	AG32-3-32	AG32-3-36	AG32-3-39	AG32-3-50	AG32-3-63	AG32-3-105
SiO ₂	49.64	49.67	49.84	49.48	50.12	49.90	50.64	49.03	49.68
TiO ₂	2.07	.77	2.13	1.84	2.13	2.11	1.85	1.41	1.44
Al ₂ O ₃	14.37	17.61	14.48	14.84	14.80	14.69	14.76	15.77	16.10
FeO	12.23	8.49	12.31	11.49	11.66	11.61	11.19	10.40	10.49
MnO	.17	.08	.17	.17	.14	.16	.17	.16	.14
MgO	7.06	9.71	6.75	7.39	6.95	6.86	7.13	7.73	8.00
CaO	10.38	11.33	10.71	10.69	10.30	10.32	10.64	11.02	11.43
Na ₂ O	3.13	2.56	3.08	2.89	2.96	2.61	3.09	2.96	2.82
K ₂ O	.08	.04	.13	.17	.19	.22	.17	.07	.10
TOTAL	99.12	100.30	99.60	98.96	99.25	98.46	99.64	98.56	100.20
Mg#	50.7	67.1	49.4	53.4	51.5	51.3	53.2	57.0	57.6
(All iron as FeO)									
Mg#	53.9	69.8	52.6	56.5	54.7	54.4	56.3	60.1	60.7
(Fe ₂ O ₃ /FeO=0.15)									

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Table 12: (Cont.) Electron microprobe analyses of quench glass from the AG32-3 samples. All iron as FeO.

G R O U P A

		AG32-3-2		AG32-3-18		AG32-3-32		AG32-3-30		73	74	75	83	AG32-3		100	109	112
		WR	G	WR	G	WR	G	WR	G	WR	WR	WR	WR	WR	WR	WR	WR	WR
QZ		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Or		1.54	.47	2.36	.77	1.12	1.00	1.60	.53	1.48	1.77	1.65	2.01	2.30	2.84	1.77	1.77	1.48
Pl	Ab	23.44	26.49	25.47	26.06	21.66	24.45	22.42	26.15	22.17	23.95	24.03	22.93	23.86	24.20	23.10	22.93	22.59
	An	27.04	24.92	27.82	25.30	28.69	27.02	27.09	26.14	27.75	25.89	26.32	27.57	25.56	25.39	26.01	26.43	26.84
Di	Wo	10.39	10.93	10.54	11.60	9.97	10.68	10.57	11.04	10.40	11.04	10.96	10.83	11.13	11.18	10.69	10.72	10.53
	En	5.86	5.96	5.46	6.22	5.33	6.02	5.75	6.22	5.56	5.87	5.86	5.73	5.95	5.72	5.61	5.59	5.52
	Fs	4.10	4.58	4.79	5.00	4.31	4.21	4.45	4.36	4.50	4.82	4.75	4.78	4.81	5.18	4.77	4.82	4.70
Hy	En	7.72	7.47	4.28	7.03	8.72	7.91	10.59	9.40	8.32	7.41	7.30	6.31	8.25	6.16	7.86	7.62	8.50
	Fs	5.39	5.74	3.75	5.65	7.06	5.53	8.19	6.59	6.73	6.08	5.92	5.26	6.67	5.58	6.69	6.56	7.24
Ol	Fo	4.45	2.91	4.32	2.49	3.09	3.13	1.41	1.49	2.90	2.90	3.01	3.43	2.07	3.00	2.72	2.64	2.31
	Fa	3.43	2.46	4.18	2.21	2.76	2.42	1.20	1.15	2.59	2.63	2.69	3.16	1.84	2.99	2.55	2.51	2.17
Mt		2.12	2.10	2.13	2.12	2.28	1.99	2.17	1.93	2.22	2.17	2.13	2.16	2.13	2.25	2.28	2.23	2.25
Il		3.57	3.93	3.93	4.05	3.76	3.49	3.82	3.51	3.72	3.67	3.67	3.67	3.78	3.97	4.03	3.87	3.87
Ap		.47	-	.47	-	.50	-	.50	-	.50	.52	.47	.57	.64	.62	.57	.59	.64
TOTAL		99.51	97.97	99.52	98.50	99.23	97.86	99.76	98.53	98.91	98.74	98.76	98.38	99.01	99.08	98.65	98.30	98.60

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Table 13: CIPW Norms calculated assuming $\text{Fe}_2\text{O}_3 / \text{FeO} = 0.15$

	AG32-3-63		AG32-3		AG32-3-105		AG32-3		AG32-3		AG32-3-36		AG32-3-39		AG32-3		AG32-5-5		
	WR	G	WR	WR	WR	G	WR	WR	WR	WR	WR	G	WR	G	WR	WR	WR	G	
Qz	-	-	-	-	-	-	-	-	-	-	-	-	-	0.91	-	-	-	-	
Or		1.24	.41	.77	1.30	1.36	1.59	1.24	1.83	.71	1.30	1.42	1.12	1.24	1.30	1.48	1.65	.47	0.24
Pl	Ab	22.00	25.05	17.77	22.34	21.07	23.86	21.32	22.68	20.99	22.25	22.76	25.05	21.75	22.09	23.29	22.68	17.68	21.66
	An	30.79	29.54	35.27	30.26	31.58	30.98	32.30	30.47	33.04	25.75	21.32	26.54	25.33	27.72	18.42	26.46	43.04	36.44
Di	Wo	10.37	10.45	9.80	10.31	10.30	10.68	11.41	10.90	10.29	9.13	8.97	10.26	9.14	9.76	9.08	9.71	9.03	8.19
	En	5.90	6.11	5.98	5.75	5.86	6.31	6.67	6.09	6.16	5.29	5.80	5.71	5.47	5.41	6.06	5.43	5.71	5.33
	Fs	4.02	3.84	3.27	4.16	3.99	3.84	4.20	4.37	3.60	3.42	2.56	4.14	3.19	3.98	2.35	3.89	2.76	2.29
Hy	En	5.68	4.89	8.16	5.61	6.41	5.72	3.84	4.68	4.81	9.89	3.39	10.15	8.85	11.68	-	9.76	4.66	7.37
	Fs	3.87	3.07	4.46	4.05	4.37	3.48	2.42	3.36	2.81	6.40	1.50	7.37	5.16	8.59	-	6.99	2.25	3.17
Ol	Fo	5.57	5.78	5.46	5.36	4.90	5.53	6.72	5.43	7.50	5.35	17.33	1.01	7.59	-	22.77	3.51	7.36	8.05
	Fa	4.19	4.00	3.29	4.27	3.68	3.71	4.67	4.30	4.83	3.82	8.44	.81	4.88	-	9.74	2.77	3.92	3.81
Mt		1.97	1.80	1.79	2.04	1.93	1.81	1.83	1.97	1.83	2.26	2.25	2.02	2.25	2.00	2.22	2.23	1.39	1.46
Il		2.87	2.68	2.01	3.06	2.72	2.73	2.41	2.87	2.34	3.93	3.51	4.05	3.84	4.01	3.30	4.01	1.29	1.46
Ap		.43	-	.24	.40	.45	-	.31	.36	.31	.69	.47	-	.45	-	.47	.59	.17	-
TOTAL		98.90	97.62	98.19	98.90	98.63	99.25	99.33	99.31	99.21	99.48	99.72	98.22	99.14	97.43	99.63	99.68	99.74	99.48

Table 13 (Cont.): CIPW Norms calculated assuming $\text{Fe}_2\text{O}_3 / \text{FeO} = 0.15$.

Major elements

All samples studied are olivine normative tholeiites and plot within the field of typical MORB (Dickey et al, 1977) on a total alkali vs silica plot (figure 22). Compositional variation in terms of an AFM diagram is shown in figure 23, together with fields for Atlantic fracture zone and rift valley volcanics (Hekinian and Thompson, 1976). Group A samples plot within the fracture zone field, Group B and some Group C basalts plot at the intersection of the fields for fracture zone and rift valley volcanics. Group D and the two samples with highest Mg# from Group C (AG32-3-36, AG32-3-40) plot in the rift valley volcanic field. whole rock major element characteristics of the four basalt groups can be summarised as follows. Group maxima, minima and mean compositions are presented in Table 14.

- 1: The evolved sparsely plagioclase phyric basalts - Group A. These basalts are characterised by slightly higher silica (49.2 - 50.36 wt%) and higher TiO_2 (1.89 - 2.15 wt%) and FeO (9.77 - 10.59 wt%) and lower MgO (6.42 - 8.04) than the Group B basalts. Mg# ranges from 52.8 - 59.4.
- 2: The sparsely plagioclase phyric basalts - Group B. These samples are clearly separated from the Group A basalts by Mg# which ranges from 59.2 - 66.7 (Group A Mg# = 52.8 - 59.4) and by having slightly lower silica (48.98 - 49.44 wt%) and lower TiO_2 (1.08 - 1.63 wt%) contents, and significantly higher Al_2O_3 (15.85 - 16.76 wt%) and CaO

Major elements

All samples studied are olivine normative tholeiites and plot within the field of typical MORB (Dickey et al, 1977) on a total alkali vs silica plot (figure 22). Compositional variation in terms of an AFM diagram is shown in figure 23, together with fields for Atlantic fracture zone and rift valley volcanics (Hekinian and Thompson, 1976). Group A samples plot within the fracture zone field, Group B and some Group C basalts plot at the intersection of the fields for fracture zone and rift valley volcanics. Group D and the two samples with highest Mg# from Group C (AG32-3-36, AG32-3-40) plot in the rift valley volcanic field. whole rock major element characteristics of the four basalt groups can be summarised as follows. Group maxima, minima and mean compositions are presented in Table 14.

- 1: The evolved sparsely plagioclase phyric basalts - Group A. These basalts are characterised by slightly higher silica (49.2 - 50.36 wt%) and higher TiO_2 (1.89 - 2.15 wt%) and FeO (9.77 - 10.59 wt%) and lower MgO (6.42 - 8.04) than the Group B basalts. Mg# ranges from 52.8 - 59.4.
- 2: The sparsely plagioclase phyric basalts - Group B. These samples are clearly separated from the Group A basalts by Mg# which ranges from 59.2 - 66.7 (Group A Mg# = 52.8 - 59.4) and by having slightly lower silica (48.98 - 49.44 wt%) and lower TiO_2 (1.08 - 1.63 wt%) contents, and significantly higher Al_2O_3 (15.85 - 16.76 wt%) and CaO

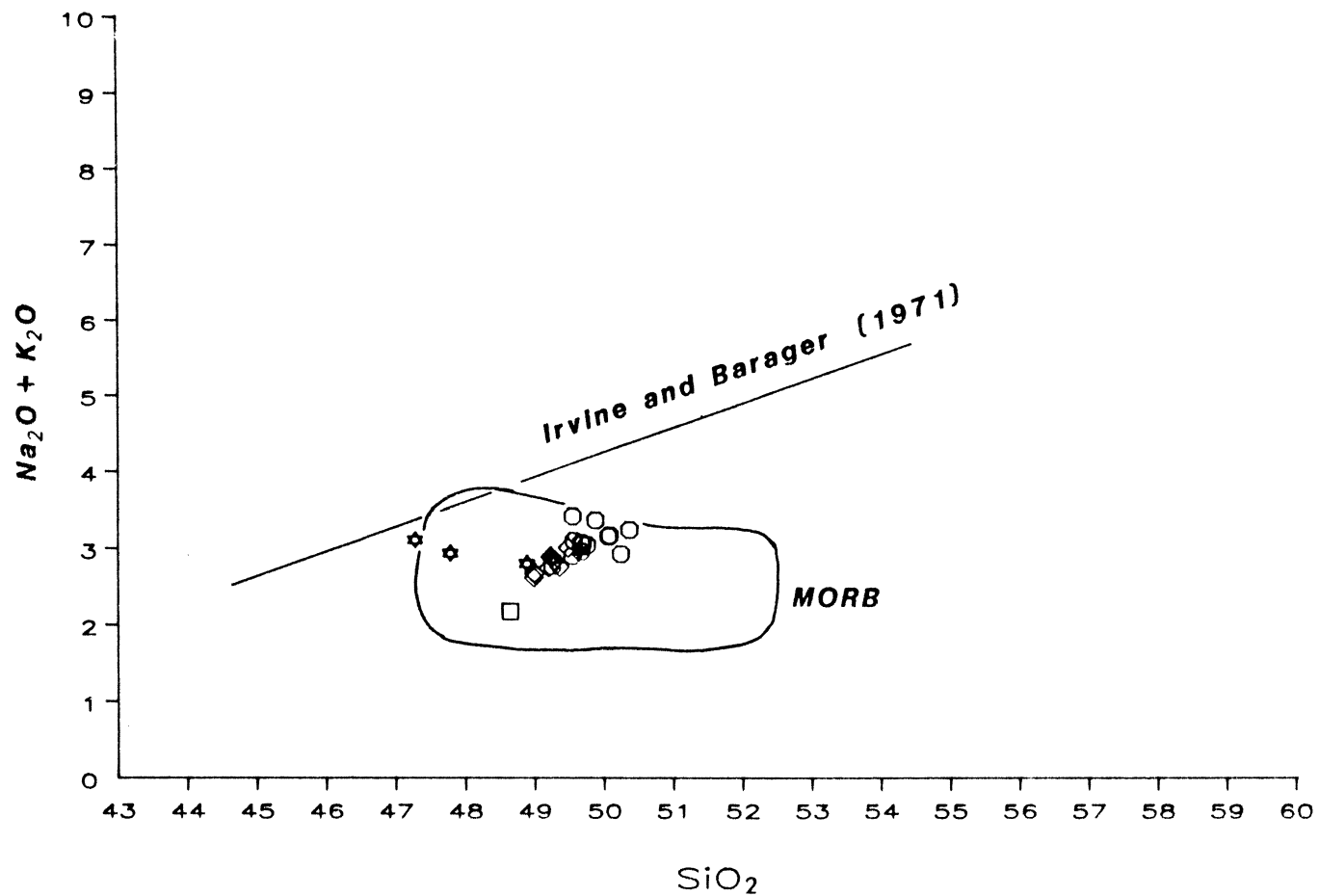


Figure 22. $\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs SiO_2 . Symbols as per Figure 10. MORB field from Dickey et al (1977).

Alkaline - Sub alkaline divide from Irvine and Barager (1971).

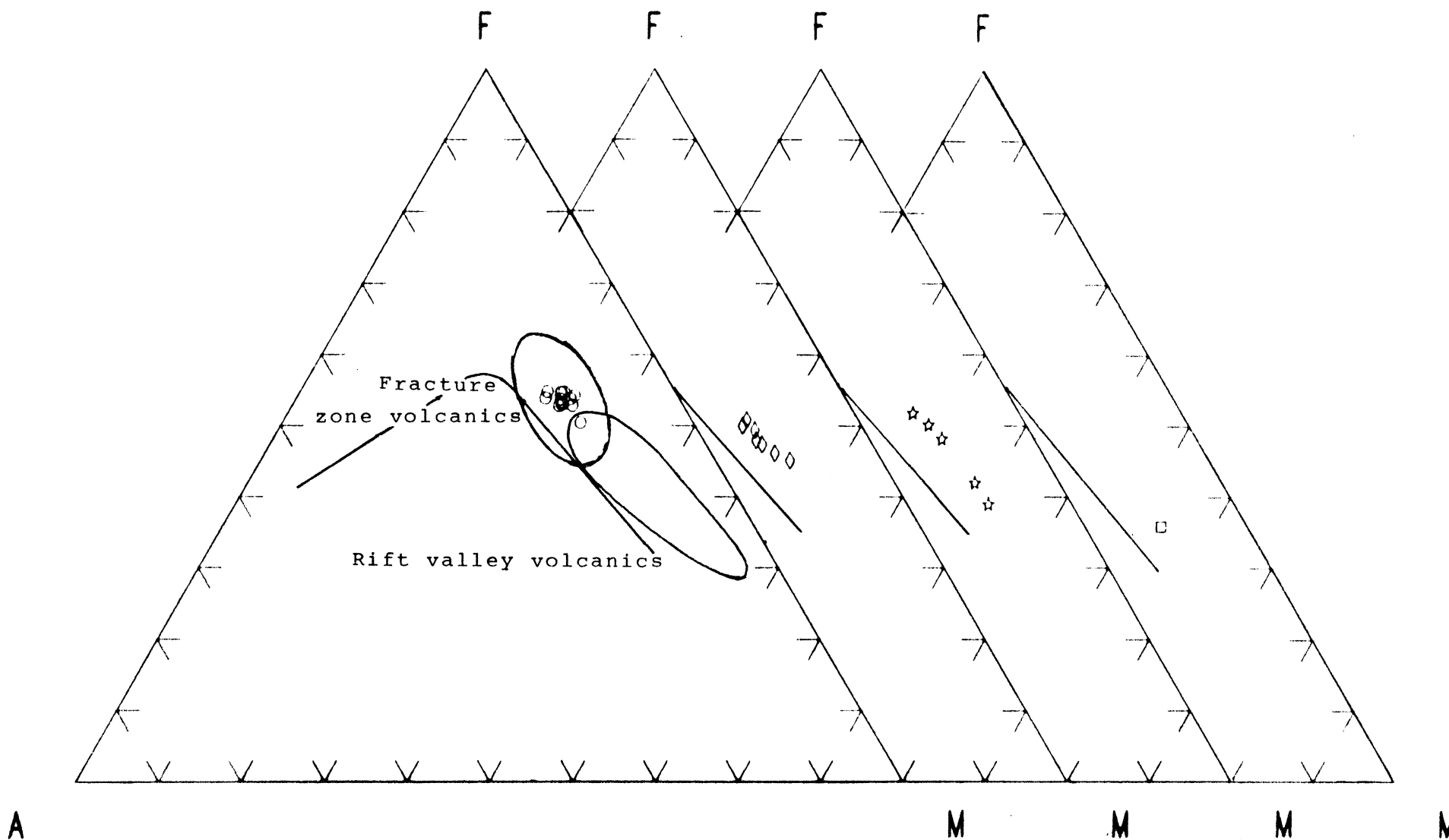


Figure 23. AFM diagram for AG32-3 basalts with all Fe as FeO. Symbols as per Figure 22. Fracture zone and Rift valley volcanic fields from Hekinian and Thompson (1976). All samples plot on the Tholeiitic side of Irvine and Bararger's (1971) tholeiitic - calc alkaline divide.

	GROUP A					GROUP B					GROUP C					GROUP D				
	MIN	MAX	MEAN	S.D.	N	MIN	MAX	MEAN	S.D.	N	MIN	MAX	MEAN	S.D.	N	MIN	MAX	MEAN	S.D.	N
SiO ₂	49.25	50.26	49.49	.31	13	48.98	49.48	49.21	.18	7	47.28	49.64	48.56	1.00	5	48.64	48.64	48.64	0.00	1
TiO ₂	1.89	2.15	2.02	.07	13	1.08	1.63	1.39	.20	7	1.75	2.12	1.97	.16	5	.68	.68	.68	0.00	1
Al ₂ O ₃	14.55	15.65	14.84	.31	13	15.85	16.76	16.22	.32	7	11.75	14.46	13.34	1.14	5	19.35	19.35	19.35	0.00	1
Fe ₂ O ₃	1.47	1.59	1.53	.04	13	1.20	1.43	1.32	.08	7	1.54	1.57	1.55	.01	5	.96	.96	.96	0.00	1
FeO	9.77	10.59	10.21	.27	13	7.96	9.50	8.80	.52	7	10.26	10.46	10.35	.08	5	6.42	6.42	6.42	0.00	1
MnO	.17	.22	.20	.01	13	.16	.19	.18	.01	7	.18	.21	.19	.01	5	.13	.13	.13	0.00	1
MgO	6.42	8.04	7.10	.41	13	7.49	8.94	8.12	.54	7	8.14	15.54	11.35	3.13	5	8.40	8.40	8.40	0.00	1
CaO	10.78	11.27	10.98	.12	13	11.43	12.27	11.82	.30	7	8.39	10.39	9.52	.83	5	13.17	13.17	13.17	0.00	1
Na ₂ O	2.58	3.02	2.78	.12	13	2.50	2.70	2.59	.08	7	2.59	2.86	2.70	.10	5	2.10	2.10	2.10	0.00	1
K ₂ O	.19	.49	.31	.08	13	.12	.31	.21	.06	7	.21	.28	.24	.03	5	.08	.08	.08	0.00	1
P ₂ O ₅	.20	.27	.24	.02	13	.10	.19	.15	.03	7	.20	.25	.23	.02	5	.07	.07	.07	0.00	1
Zr	124	145	132	7	13	63	103	84	14	7	112	135	126	10	5	35	35	35	0	1
Nb	4.8	6.3	5.9	.5	13	1.6	3.9	2.9	.7	7	5.6	7.8	6.8	.8	5	1.8	1.8	1.8	0.0	1
Y	45	54	49	3	13	26	39	33	5	7	40	47	44	3	5	18	18	18	0	1
Rb	5.0	86	13	22	13	2.8	6.2	4.0	1.1	7	4.8	6.1	5.5	.5	5	.01	.01	.01	0.00	1
Ba	36	51	43	4	12	15	29	23	5	7	44	52	49	3	5	8.9	8.9	8.9	0.0	1
Sr	99	120	109	8	13	120	134	126	5	7	97	116	108	8	5	95	95	95	0	1
Co	46	51	49	1	12	49	53	51	1	7	51	69	60	8	5	44	44	44	0	1
Cr	175	242	211	19	12	265	366	300	38	7	502	1079	775	249	5	360	360	360	0	1
Ni	87	115	101	7	13	125	188	144	22	7	173	653	381	202	5	136	136	136	0	1
V	331	361	346	10	12	215	284	258	25	7	250	347	321	42	5	156	156	156	0	1
Zn	100	113	105	4	13	69	89	81	7	7	98	106	102	3	5	53	53	53	0	1
Cu	45	53	50	2	13	59	77	64	6	7	45	53	50	4	5	67	67	67	0	1
Sc	36	42	40	2	12	40	42	40	1	7	32	40	38	3	5	32	32	32	0	1
Zr/Nb	20.0	28.0	22.9	2.3	13	27.0	38.0	30.3	4.4	7	16.7	21.0	18.2	1.7	5	19.6	19.6	19.6	0.0	1
Zr/Y	2.60	2.90	2.74	.10	13	2.40	2.60	2.54	.08	7	2.80	2.90	2.88	.04	5	2.00	2.00	2.00	0.00	1
Y/Nb	7.40	9.70	8.38	.72	13	6.40	15.7	11.4	2.9	7	5.80	7.30	6.48	.61	5	9.80	9.80	9.80	0.00	1
Ti/Zr	85.0	96.0	91.6	3.4	13	95.0	102	98.9	2.3	7	93.0	95.0	93.8	.8	5	117	117	117	0	1
Ti/P	10.1	14.2	11.9	1.3	13	10.3	14.6	12.8	1.5	7	9.80	14.8	12.2	1.8	5	13.3	13.3	13.3	0.0	1
Mg#	52.8	59.4	55.3	1.6	13	59.2	66.7	62.1	2.8	7	58.5	73.0	65.2	6.2	5	70.0	70.0	70.0	0.0	1

Table 14: Basic statistics of AG32-3 basalt groups. Maxima, Minima and Means of whole rock analyses assuming Fe₂O₃ / FeO of 0.15

(11.43 - 12.27 wt%) contents. Significant differences between groups A & B for all major elements at 99% confidence levels are revealed by statistical (t-test) analysis.

- 3: The olivine phyric basalts - Group C. A wide range in Mg# (58.5 - 73.0) is displayed by these basalts, coupled with high to very high MgO (8.39 - 15.54 wt%) contents which reflects the varied proportions of olivine phenocrysts contributing to these rocks. Furthermore, Group C basalts are characterised by high FeO (10.26 - 10.46 wt%), high TiO₂ (1.75 - 2.12 wt%), and by low Al₂O₃ (11.75 - 14.46 wt%) and CaO (8.39 - 10.39 wt%) contents. There is significant compositional variation within Group C and the major element covariations define clear trends showing strong olivine and Cr-spinel control (see Figure 24).

- 4: The plagioclase phyric basalt - Group D. This sample is characterised by extremely high Al₂O₃ (19.35 wt%) and CaO (13.17 wt%) contents, consistent with the highly plagioclase phyric nature of the rock. However, the high Mg# (70.0) indicates that this rock is extremely primitive. The quench glass also has a high Al₂O₃ content (17.61 wt%) and it is suggested that AG32-3-5 may represent a High Alumina Basalt (Kuno, 1960). The low TiO₂ (0.68 wt%), K₂O (0.08 wt%) and P₂O₅ (0.07 wt%) of this rock are consistent with its primitive nature. These concentrations may however, have been reduced by the diluting effect of plagioclase accumulation.

Quench glass rinds are present on 9 of the 26 samples selected for detailed analysis (see Table 15) and microprobe major element analyses of these glasses are presented in Table 12. All the glasses are olivine tholeiitic, save for that from AG32-3-39 which is quartz normative (quartz = 0.91%). CIPW norms were calculated assuming $\text{Fe}_2\text{O}_3/\text{FeO} = 0.15$ and are presented in Table 13.

Glasses from Group A samples (AG32-3-2; AG32-3-18; AG32-3-32; AG32-3-50) are evolved ($\text{Mg\#} = 52.6 - 56.5$) and are characterised by high incompatible element concentrations ($\text{TiO}_2 = 1.84 - 2.13$ wt%; $\text{Na}_2\text{O} = 2.89 - 3.13$ wt%; $\text{K}_2\text{O} = 0.08 - 0.17$ wt%) and by high FeO (12.23 wt%) and low MgO (7.06 wt%) contents. Glasses from Group C samples (AG32-3-36; AG32-3-39) are similar to those from Group A in that they are also evolved ($\text{Mg\#} = 54.4 - 54.7$), are characterised by high incompatible element contents ($\text{TiO}_2 = 2.11 - 2.13$ wt%; $\text{Na}_2\text{O} = 2.61 - 2.96$ wt%; $\text{K}_2\text{O} = 0.19 - 0.22$ wt%) and have high FeO (11.61 - 11.66 wt%) and low MgO (6.86 - 6.95 wt%) contents. The range in Al_2O_3 concentration in the Group A samples (14.37 - 14.86 wt%) encompasses the Group C Al_2O_3 variation (14.80 - 14.36 wt%). Quench glasses from the Group B samples (AG32-3-63; AG32-3-105) have higher $\text{Mg\#}$ than Group A or Group C samples ($\text{Mg\#} = 60.1 - 60.7$), and are characterised by low TiO_2 (1.41 - 1.44 wt%), FeO (10.40 - 10.49) and K_2O (0.07 - 0.10 wt%) contents, and by slightly elevated Al_2O_3 (15.77 - 16.1 wt%) and CaO (10.02 - 11.43 wt%) relative to Group A and Group C samples. The Group D quench glass, like the corresponding whole rock composition, has the highest $\text{Mg\#}$ (69.8) and is notable for the very high Al_2O_3 (17.61 wt%) and high CaO (11.33 wt%) contents. It is further characterised by extremely

GROUP A	AG32-3-2
	AG32-3-18
	AG32-3-32
	AG32-3-50
GROUP B	AG32-3-63
	AG32-3-105
GROUP C	AG32-3-36
	AG32-3-39
GROUP D	AG32-3-5

Table 15 Samples with analysed glass.

low levels of incompatible elements ($\text{TiO}_2 = 0.77 \text{ wt\%}$; $\text{K}_2\text{O} = 0.04 \text{ wt\%}$) and by having the lowest FeO ($8.49 \text{ wt\%}$) and highest MgO ($9.71 \text{ wt\%}$) contents of all the analysed glasses.

The overall chemical diversity of the AG32, dredge 3 basalts is illustrated on the four variation diagrams TiO_2 vs $\text{Mg\#}$, Al_2O_3 vs $\text{Mg\#}$, CaO vs $\text{Mg\#}$ and P_2O_5 vs $\text{Mg\#}$ (Figure 24), in which the four basalt groups are readily distinguished and the tight clustering of samples defining these groups is evident.

On the Harker diagrams of figure 24, Groups B and C can be seen to define linear arrays. The Group A basalts always form a tight cluster near the intersection of the Group B and Group C trends and the Group D sample always plots near the end of the Group B trend furthest from Group A. The within and between group variations in these suites are discussed in detail, together with possible causes for these variations in the chapter on the petrogenesis of these rocks.

From figure 24 the excess CaO and Al_2O_3 in the Group B basalts over the more "Normal MORB" of Group A is evident and this may represent a plagioclase accumulation effect, while as a first approximation, the wide range in MgO content of the Group C basalts may be ascribed to olivine and Cr-spinel crystallization or accumulation. The highly evolved nature of the Group C glasses ($\text{Mg\#} \sim 54$), together with the high incompatible element contents of the whole rocks ($\text{TiO}_2 = 1.75 - 2.12 \text{ wt\%}$; $\text{P}_2\text{O}_5 = 0.21 - 0.28 \text{ wt\%}$) and glasses ($\text{TiO}_2 = 2.11 - 2.13 \text{ wt\%}$; $\text{K}_2\text{O} = 0.19 - 0.22$), and the elevated incompatible trace element values in the Group C whole rocks (e.g. Zr 112 - 135 ppm; see following

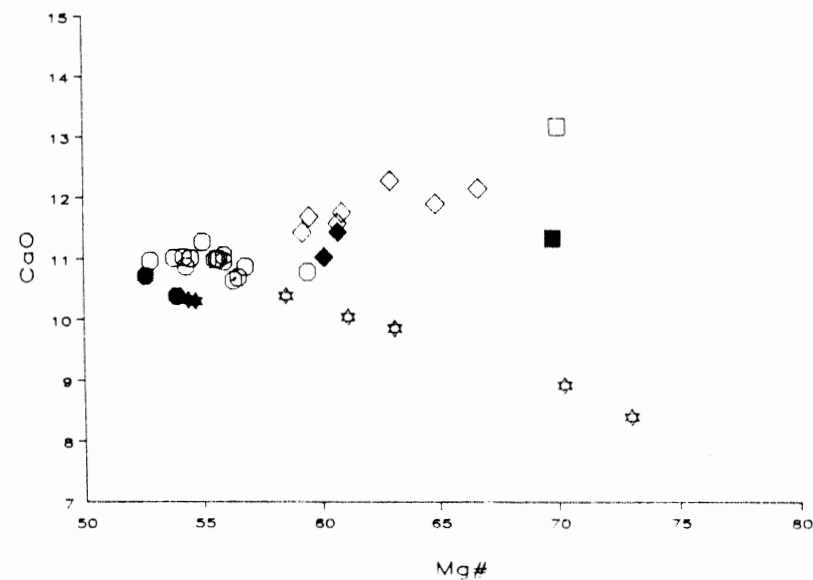
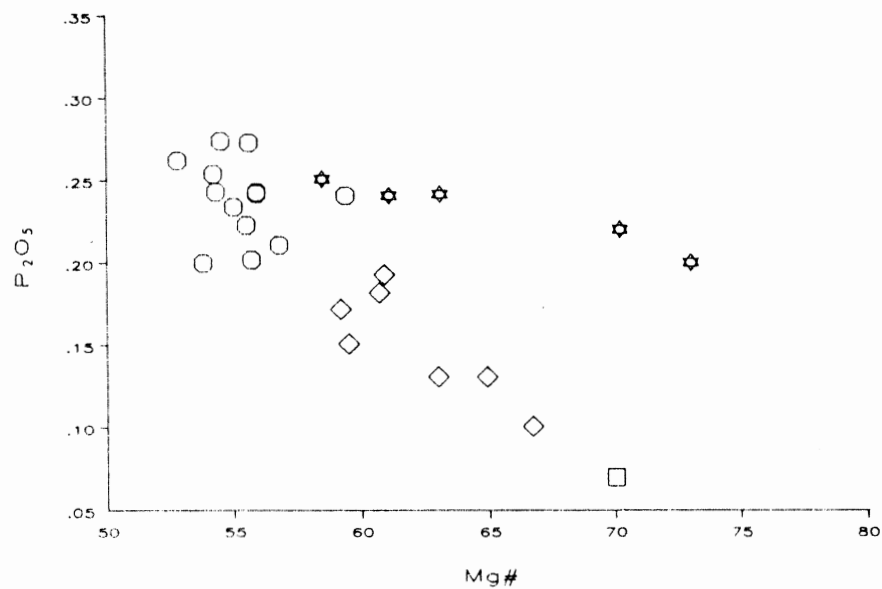
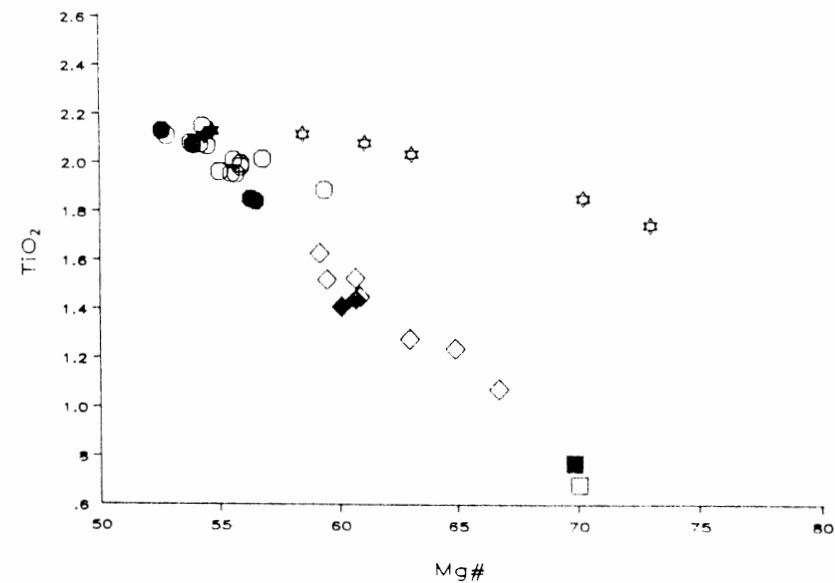
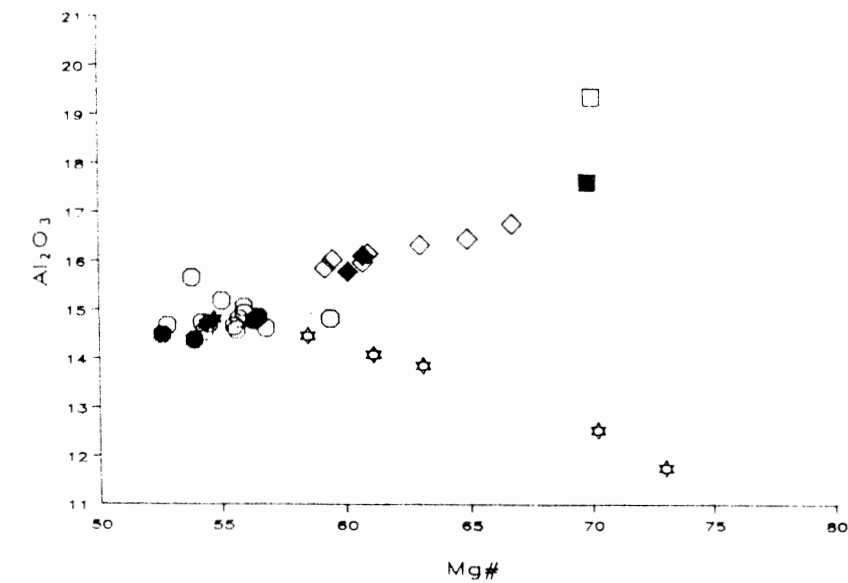


Figure 24. Major element variation diagrams; CaO , Al_2O_3 , P_2O_5 and TiO_2 vs $\text{Mg\#}$. Open symbols are whole rocks, filled symbols are glasses otherwise symbols as per Figure 10.

section) suggests that these rocks are evolved MORB that has been enriched in MgO by olivine accumulation, thus artificially raising the Mg# of these samples. Further evidence for the accumulative origin of these basalts is presented in the trace element discussion and in the Mineralogy and Petrogenesis chapters (III & V).

The Group D sample which has high Al_2O_3 and CaO contents, both in the whole rock and in the glass yet has a high Mg# (70.0 in the whole rock) and low levels of incompatible elements is an example of a fairly primitive oceanic basalt and may either represent a High Alumina Basalt, similar to those from the American-Antarctic Ridge (e.g. V27-19 in which the whole rock contains 16.6 wt% Al_2O_3 and the glass, 16.8 wt% Al_2O_3 (le Roex et al, 1985)), or may represent a plagioclase cumulate, or some combination of these two processes. These two possible modes of origin for the Group D sample are discussed in chapter V.

A comparison of the compositions of the whole rocks, which may or may not represent liquids, with the composition of the coexisting quench glass, which unequivocally represents a liquid, can indicate the processes that have been active during the evolution of a single sample. Coexisting whole rock and glass compositions are plotted on a normative Diopside - Olivine - Plagioclase ternary plot (Figure 25). This illustrates that minor amounts of in situ olivine and plagioclase fractionation have affected some of the Group A samples (notably AG32-3-18), and also illustrates that the Group B glass compositions are enriched in normative plagioclase, relative to the whole rock compositions. The differences

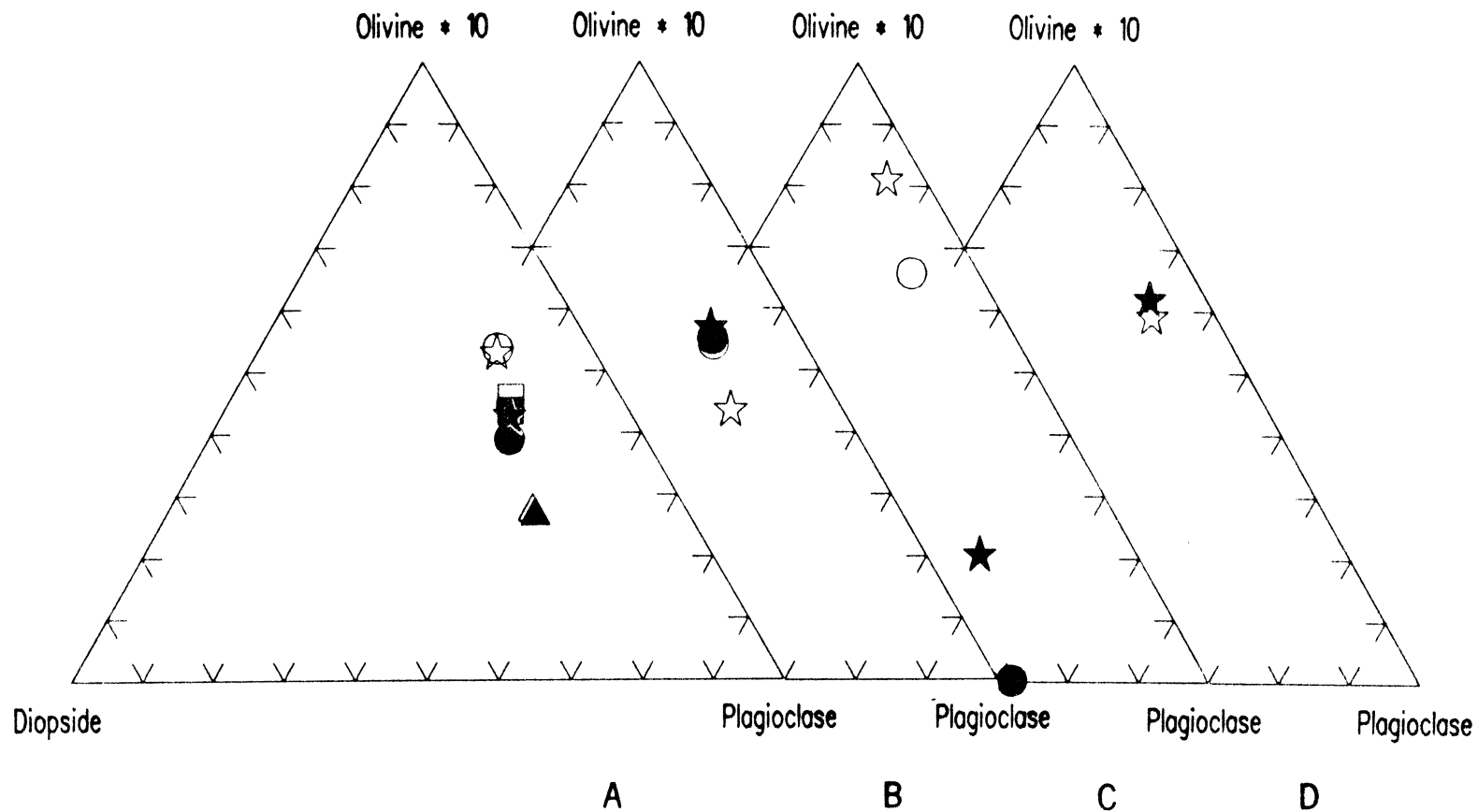


Figure 25. Normative Diopside - Olivine - Plagioclase ternary diagram showing the differences between whole rock and glass compositions. Open symbols = whole rocks, filled symbols = glasses. This illustrates the Olivine enriched nature of the whole rocks from Group C, and the insignificant whole rock - glass compositional differences in Group B and Group D basalts.

between whole rock and glass compositions for the samples from Group C are readily explained by the accumulation of olivine into a liquid of glass composition, to form the whole rock. The lack of any significant difference between the normative whole rock and glass compositions of the Group D basalt is surprising, and could either result from the sample being derived from a high Al liquid, or a result of plagioclase resorption by the glass, thereby increasing the Al and Ca contents.

The processes operating on these rocks, as identified by the differences between whole rock and glass normative chemistry, are consistent with the basalt phase diagram in the ternary plots of figure 26 (Stolper, 1980; Walker et al, 1979). From the Olivine - Diopside - Silica plot it can be seen that all samples plot within the low pressure olivine stability field. The Olivine Plagioclase - Silica and Olivine - Plagioclase - Diopside ternary diagrams further illustrate the data, and indicate that Groups A and B plot on, or very near to, the olivine - plagioclase cotectic, while the Group D basalt plots well within the plagioclase stability field.

The FeO and TiO_2 rich nature of Group A and C basalts is shown on a plot of TiO_2 vs FeO^*/MgO (Figure 27), along with fields for FAMOUS glasses, 90°East Ridge basalts and plagioclase - pyroxene basalts from the Islas Orcadas fracture zone (IOFZ). On this plot, Group D plots within the field of FAMOUS glass. Group B plots within the intersection of the FAMOUS glass and 90°E Ridge fields and Group A samples plot within the field for the 90°E Ridge basalts. No samples are as enriched in TiO_2

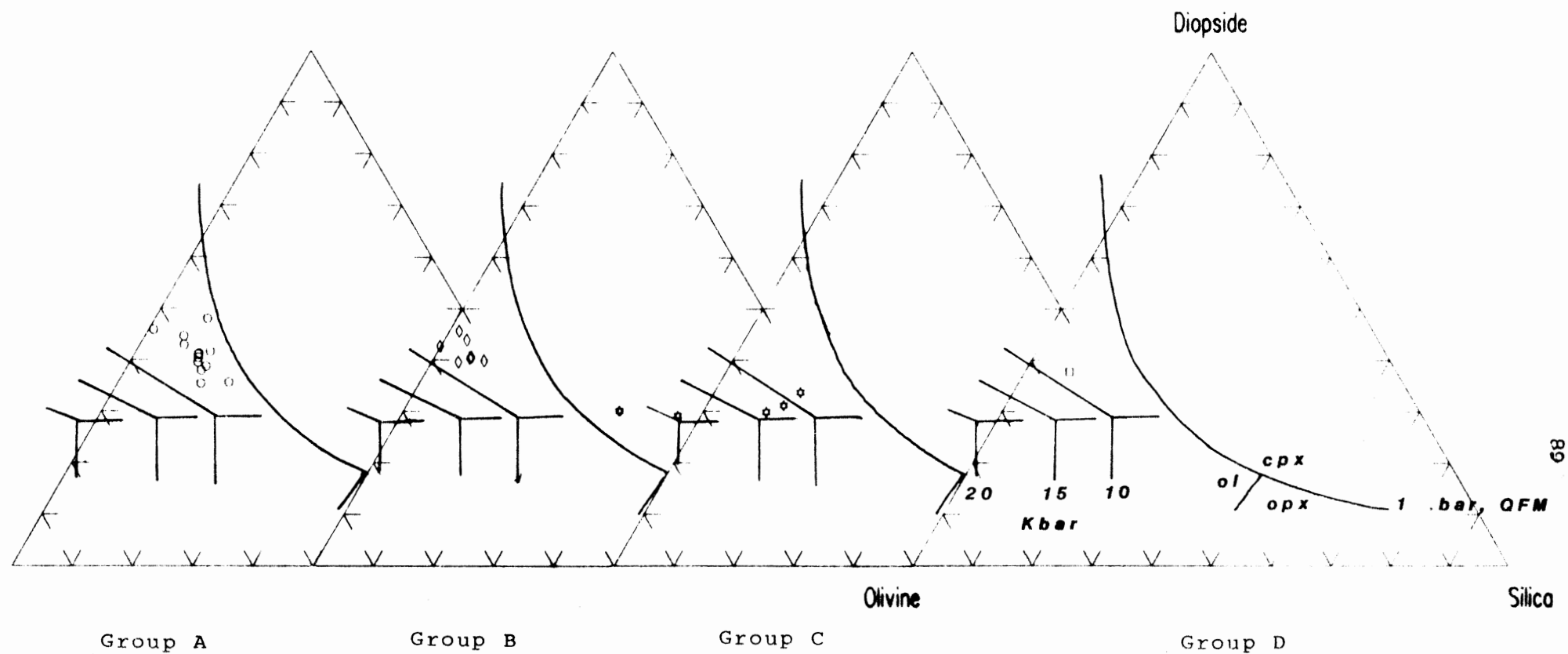


Figure 26. The Basalt phase diagram. Symbols as per Figure 10. Projection calculated from Walker et al (1979). 1 bar cotectic phase boundaries from Walker et al (1979). 10, 15 and 20 Kbar cotectic phase boundaries from Stolper (1980). Projected from plagioclase.

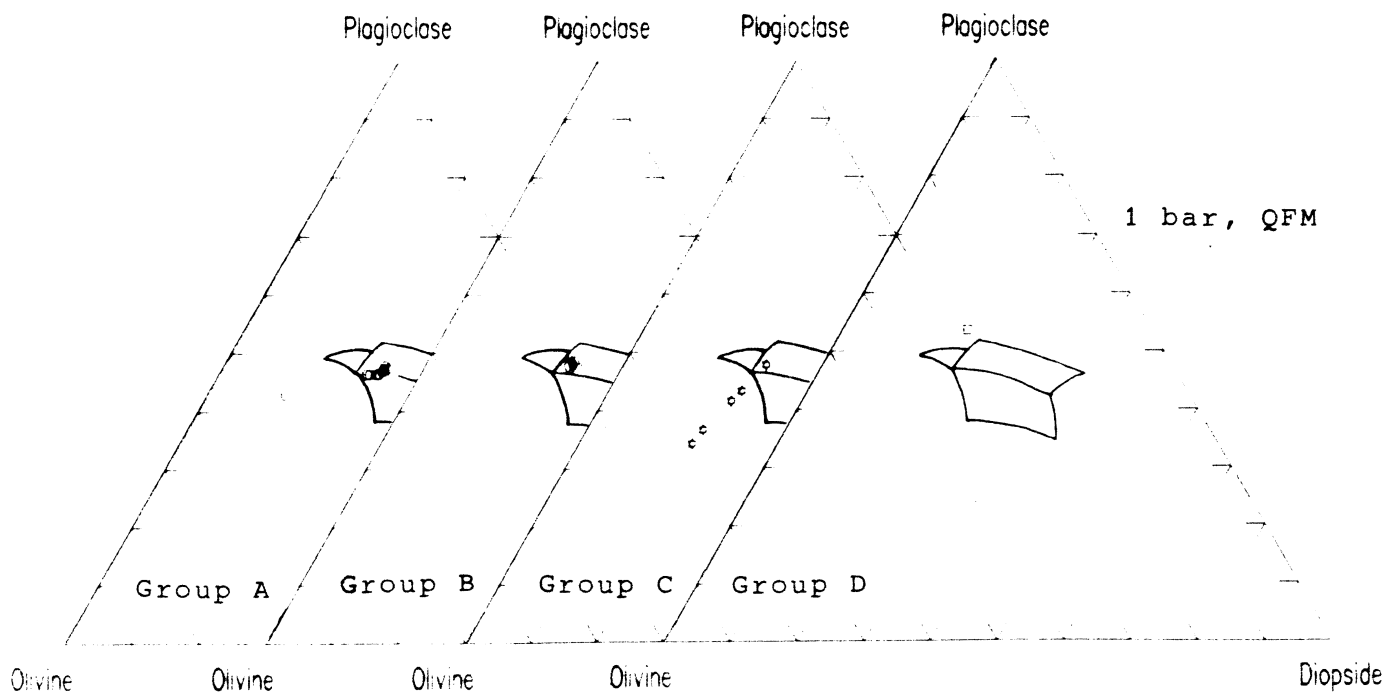
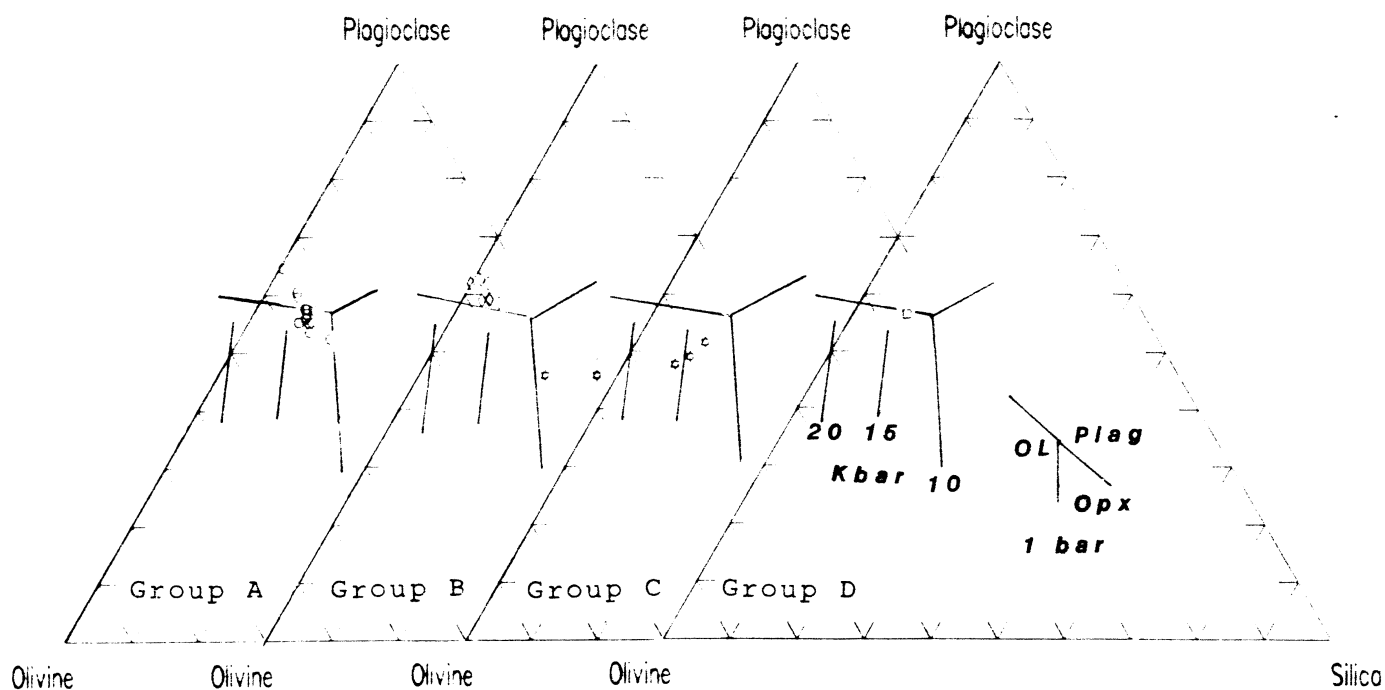


Figure 26 continued The Basalt phase diagram. Symbols as per Figure 10. Phase boundaries on Olivine - Plagioclase - Silica from Stolper (1980) and Walker et al (1979). Phase boundaries on Olivine - Plagioclase - Diopside from Walker et al (1979).

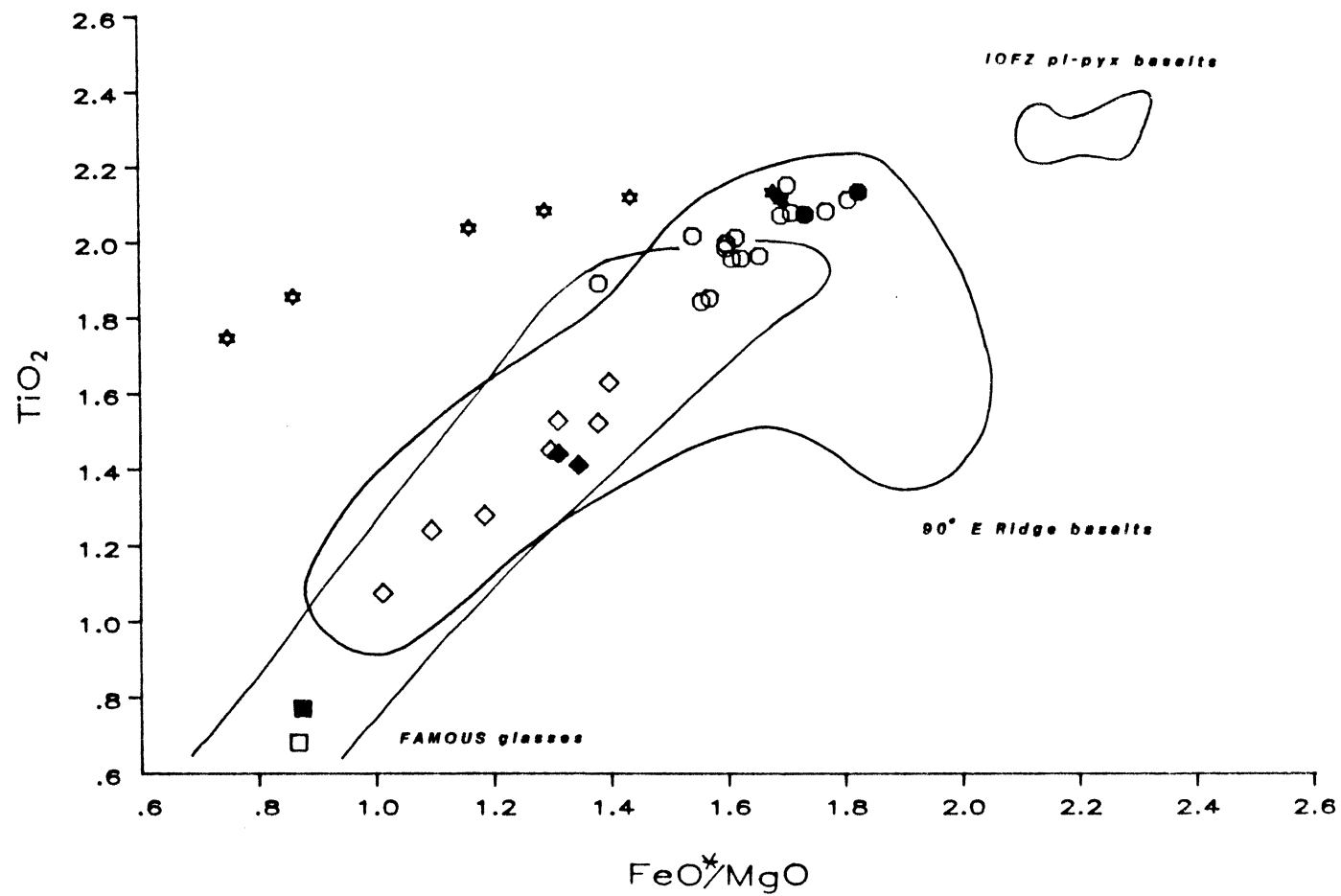


Figure 27. TiO_2 vs FeO^*/MgO . Symbols as per Figure 10. Demarcated fields from le Roex (1980).

and FeO* as the IOFZ plagioclase - pyroxene basalts (le Roex, 1980). The Group C samples show low FeO*/MgO for their TiO₂ contents which reflects the diluting effect of olivine accumulation on FeO*/MgO.

Since fractionation processes are unlikely to significantly alter the ratios of mutually incompatible elements (Gast, 1968; Erlank & Kable, 1976) the covariation of such elements is useful in investigating the genetic relationships within a suite of samples. In the AG32, dredge 3 samples, TiO₂ is such an incompatible element and is plotted against Zr (another incompatible) in figure 28, where an excellent correlation is defined. The average Ti/Zr ratio of Group A basalts is 91.6, of Group B basalts is 98.9, of Group C basalts is 93.8 and the Ti/Zr ratio of the Group D basalt is 117. The Group A and C values are slightly lower than typical MORB ((Ti/Zr = 100 - 110) Pearce & Cann, 1973; Sun et al, 1979), but are higher than values from the Islas Orcadas or Conrad fracture zones (82.6 and 87.5 respectively, le Roex, 1980). The Group B basalts are very similar to normal MORB, while the Group D basalt has Ti/Zr slightly higher than the normal MORB value.

In summary, all samples from AG32, dredge 3 are olivine normative tholeiites and they can be divided into four distinct groups on the basis of chemistry and petrography. Group A is an example of evolved MORB. The Group B samples are typical MORB and the Group D basalt is either an example of a primitive high alumina basalt, or is a simple plagioclase cumulate. The Group C basalts are olivine phyric with

elevated MgO contents and may represent evolved MORB which has experienced MgO enrichment due to olivine accumulation.

The large range in absolute element concentrations and the evolved compositions of many of the basalts suggest that fractional crystallization has played a significant role in the formation of these basalts and this will be discussed and further evaluated in a later section (Ch. V).

Trace elements

XRF analyses of 13 trace elements, Zr, Nb, Y, Rb, Ba, Sr, Co, Cr, Ni, V, Zn, Cu and Sc and selected trace element ratios are presented in Table 12. Analytical techniques and detection limits are presented in appendix 1. In the following section, trace element characteristics of each group are highlighted before discussing between group variations.

1. Group A

The immobile incompatible elements Zr, Nb and Y occur in Group A at abundance levels typical of MORB (Zr = 124 - 125 ppm, Nb = 4.8 - 6.3 ppm, Y = 45 - 54 ppm) and they are positively correlated (Figure 29). They define tight clusters with no distinct evolutionary trends when plotted against Mg# as an indicator of the degree of evolution (see Figure 30). The ferromagnesian trace elements show a fairly limited range in abundance (Ni = 87 - 115 ppm, Cr = 175 - 242 ppm, V = 331 - 361 ppm, Zn = 100 - 113 ppm, Co = 46 - 51 ppm) and all occur at

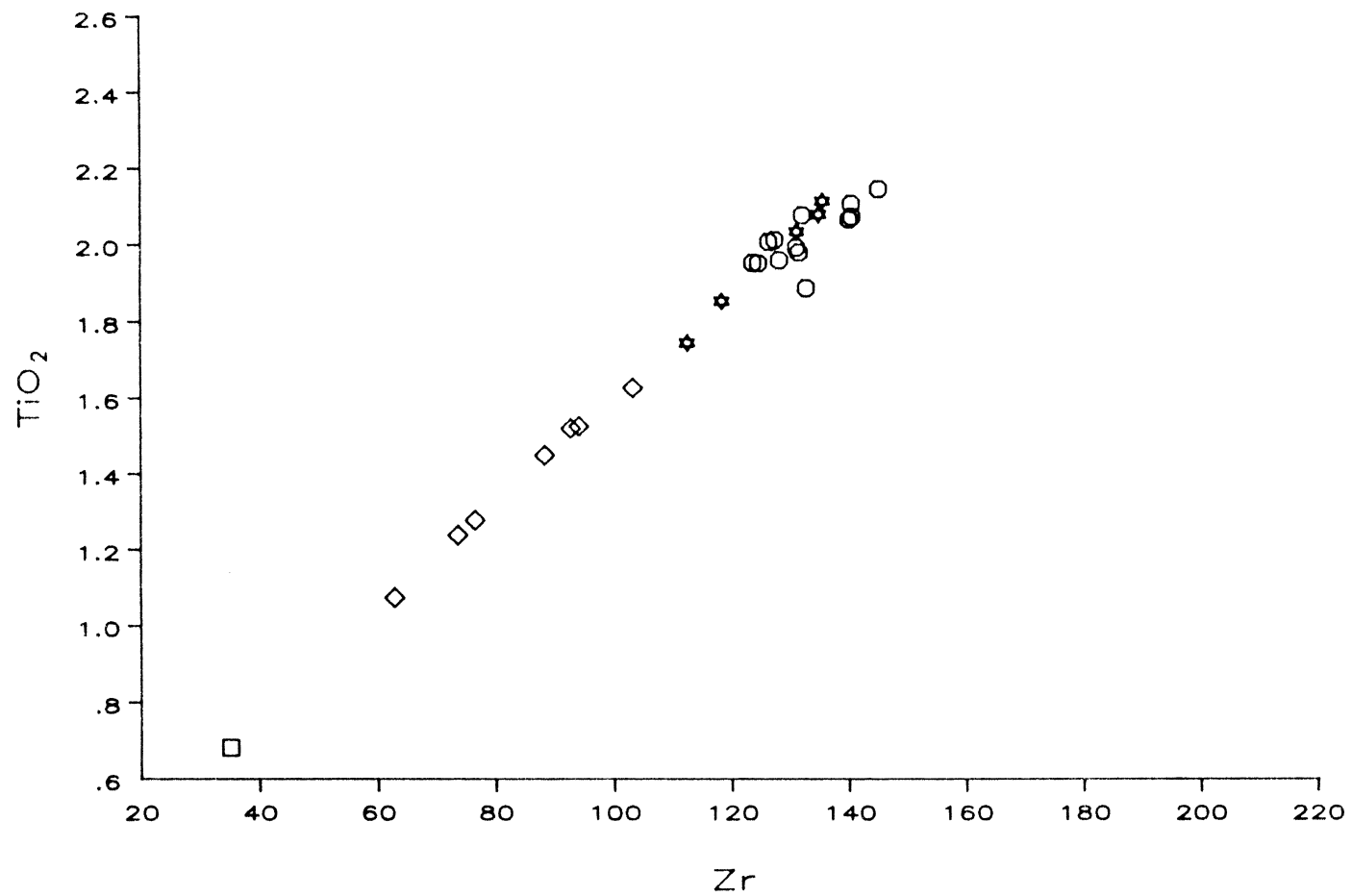


Figure 28. TiO_2 vs Zr showing strong positive correlation. Symbols as per Figure 10.

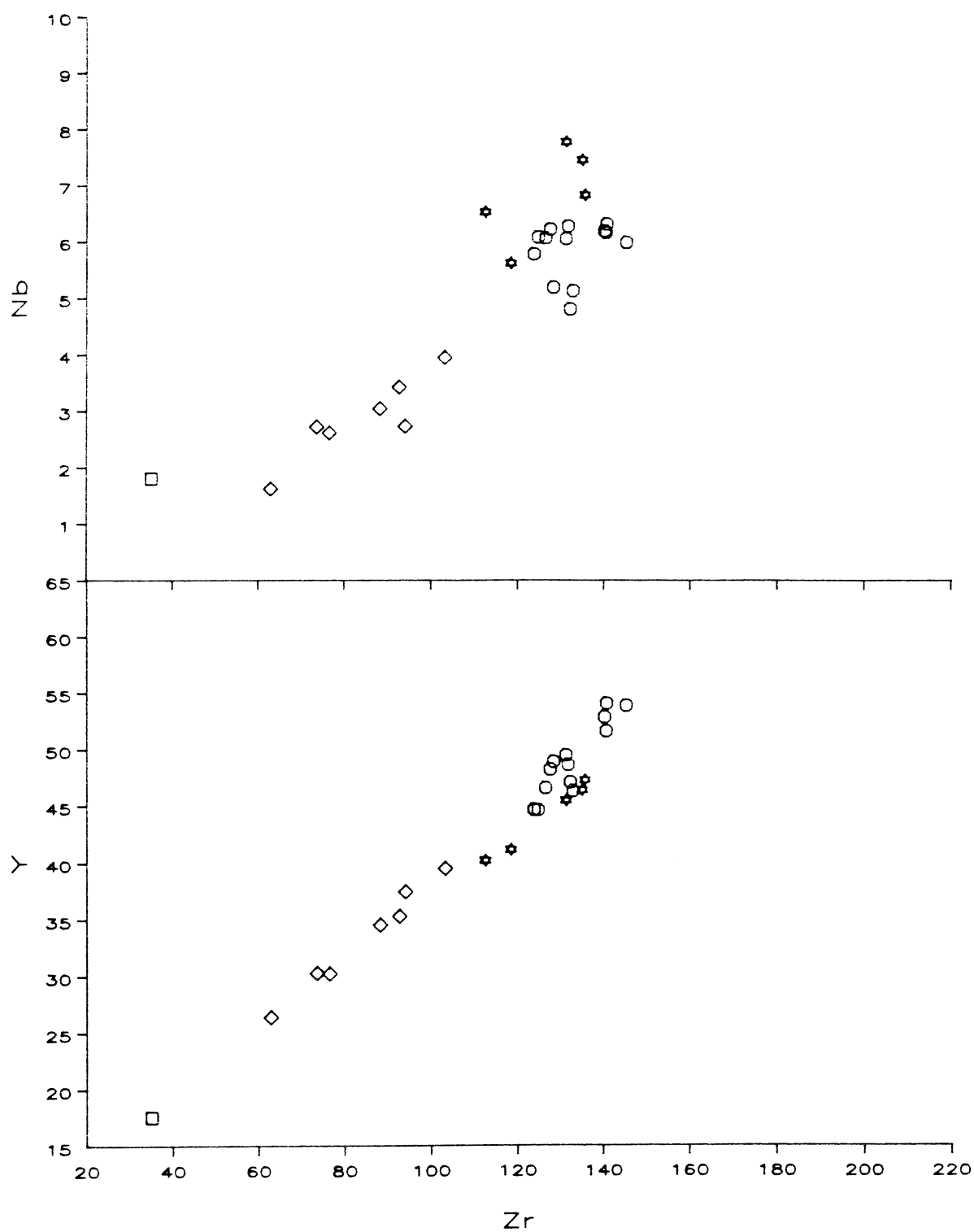


Figure 29. Nb and Y vs Zr showing strong positive correlations. Symbols as per Figure 10.

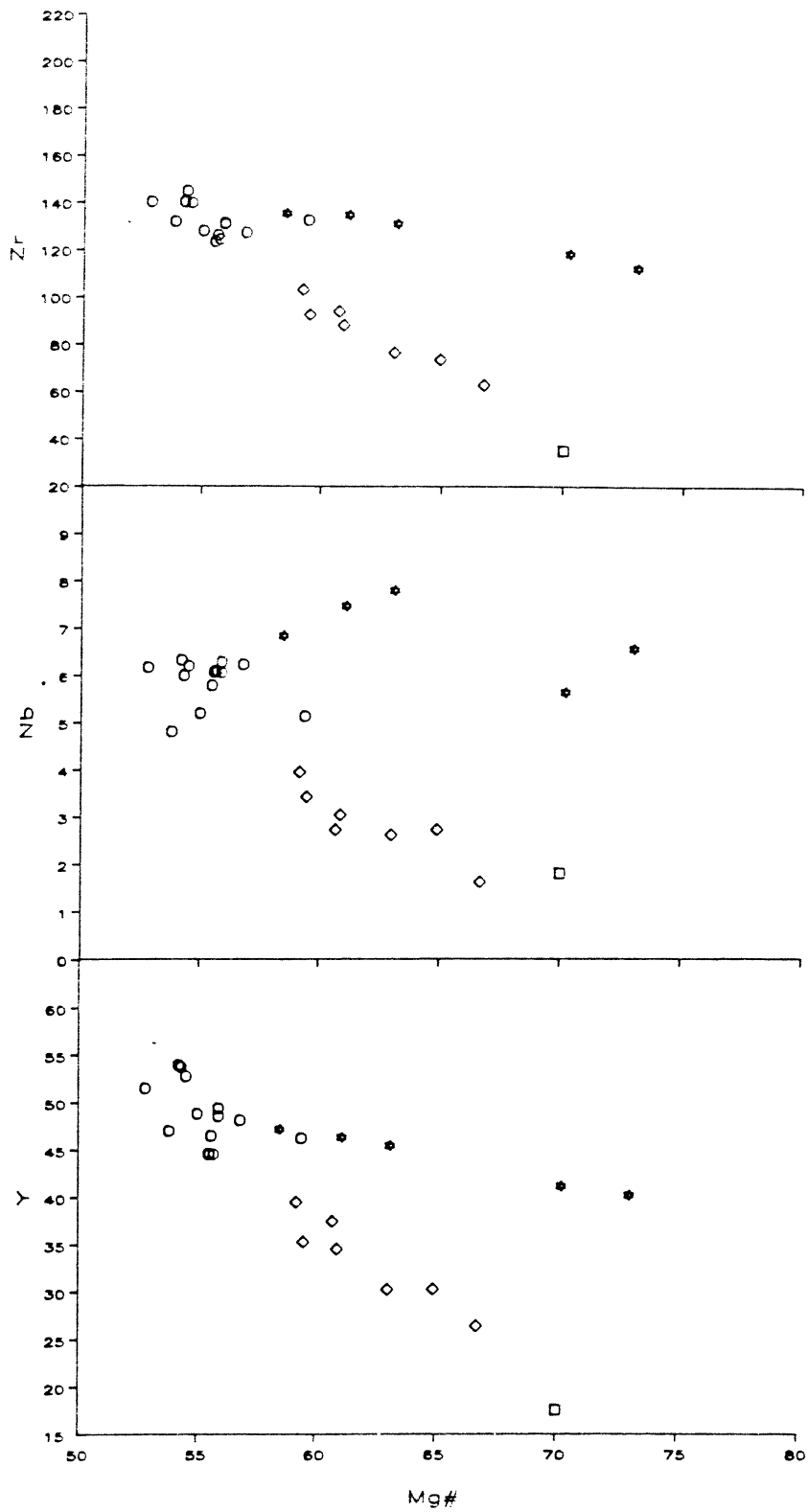


Figure 30. Zr, Nb and Y vs Mg showing distinct group trends. Symbols as per Figure 10.

levels characteristic of MORB (Erlank and Kable, 1976; le Roex, 1983). These basalts are evolved and none of the 13 samples that comprise this group have Cr or Ni at the levels expected for a primary magma (i.e. >200 ppm Ni; Hart and Davis, 1978; Sato, 1977). Of all the basalts of this study, the Group A samples have the lowest abundances of compatible trace elements (e.g. Ni, Cr, Sr) and the highest abundances of incompatible trace elements (e.g. Zr, Nb, Y), consistent with this being the most evolved basalt group.

Rb ranges from 5 - 10 ppm in Group A, (save for a single anomalous sample, AG32-3-18, with 86 ppm Rb) and Ba from 36 - 51 ppm. Both show poor negative correlations with Mg# (Figure 31). Rb and Ba are variously mobile and may be significantly affected by sea water alteration (Hart et al, 1974; le Roex et al, 1985), resulting in the scatter evident on figure 31. Sr in Group A ranges in concentration from 99 - 120 ppm and shows a poor positive correlation with Mg#. K, Rb, Ba and Sr abundances are greater in the Group A samples than in typical N-MORB, indicative of slight large ion lithophile (LIL) element enrichment of these rocks. This enrichment is a result of a fractionation process. Zr/Nb ratios in the Group A basalts (20 - 28) attest to its geochemically depleted character. Cu (45 - 53 ppm) and Sc (36 - 42 ppm) occur at abundances typical of MORB (le Roex, 1985) and Cu correlates negatively with Mg#, while Sc shows a good positive correlation with Mg#. Scatter in these data may also be a function of alteration (Hart, 1971).

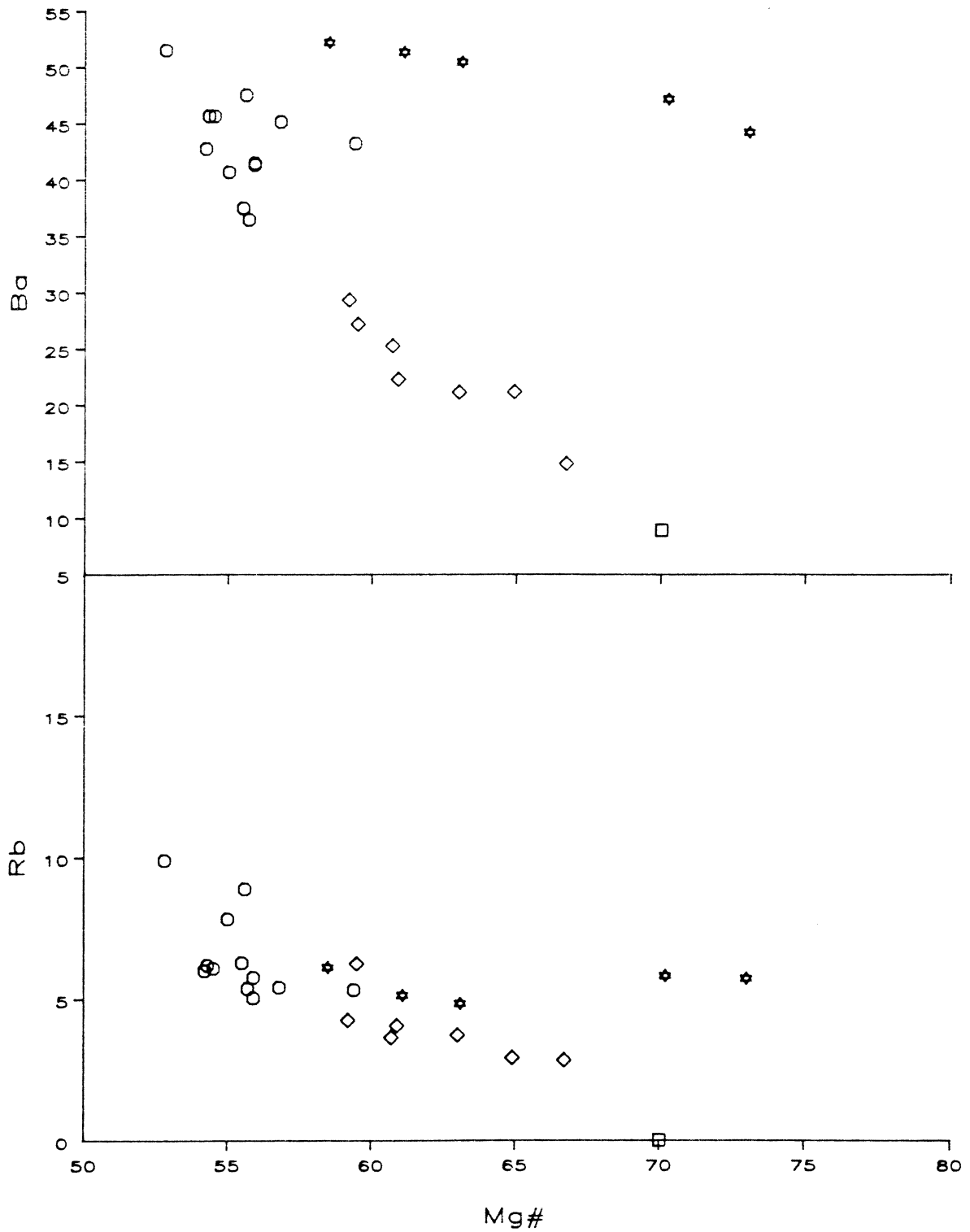


Figure 31. Ba and Rb vs Mg#. Symbols as per Figure 10. AG32-3-18
(Mg = 54, Rb = 86 ppm) not shown on Rb vs Mg.

2. Group B

The immobile incompatible elements Zr, Nb and Y occur in lower concentrations in Group B than in Group A (Zr = 63 - 103 ppm, Nb = 1.6 - 3.9 ppm, Y = 26 - 39 ppm), but still fall within the limits for normal MORB (Erlank & Kable, 1976). Good mutual interelement correlations (See Figure 29) and strong negative correlations between these elements and Mg# (see Figure 30), are consistent with the evolution of Group B by a fractionation process.

Co, Cr, Ni and Cu occur at higher abundances than in Group A basalts (e.g. Co = 49 - 53 ppm, Cr = 265 - 366 ppm, Ni = 125 - 188 ppm & Cu = 59 - 77 ppm). Zn and V show good positive correlations with the incompatible trace elements, and good negative correlations with compatible trace elements (see Figure 32 and Figure 33). As these two elements are compatible in oxide phases in basalts (K_D 's range from 2.6 - 16.9 for Zn in magnetite (Paster et al, 1974; Schock, 1977) and from 0.11 to 67 for V in magnetite (Lindstrom, 1976)), substituting for iron, they can be used as indicators of oxide fractionation. This will result in their rapid depletion with evolution, which clearly does not occur (see Figures 32 and 33). Thus it can be stated that significant amounts of oxide (e.g. titanomagnetite) fractionation did not occur during the evolution of these basalts. Cr, Ni and Cu correlate positively with Mg#, and negatively with Zr, defining an evolutionary trend controlled partially by olivine.

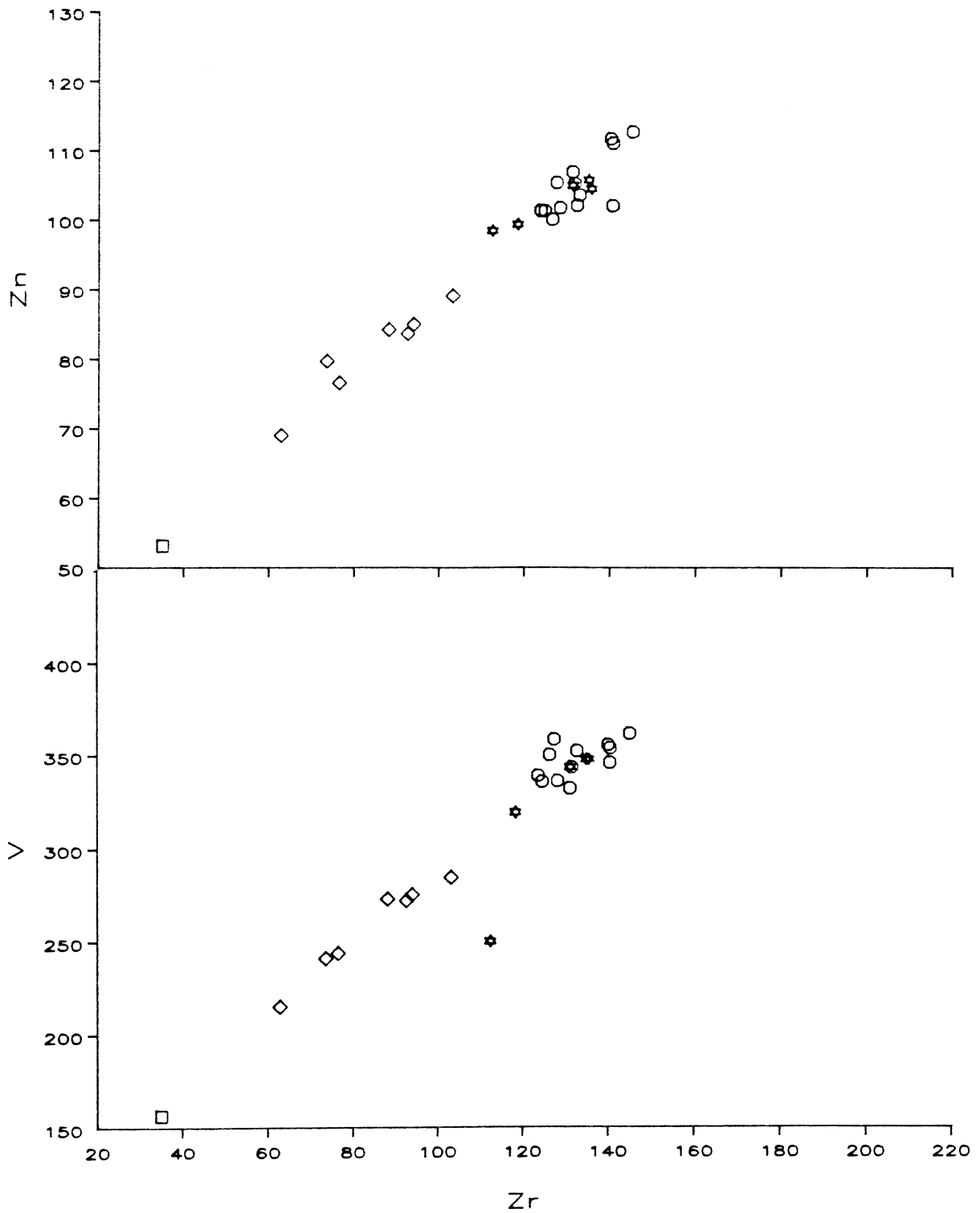


Figure 32. Zn and V vs Zr showing positive correlations. Symbols as per Figure 10.

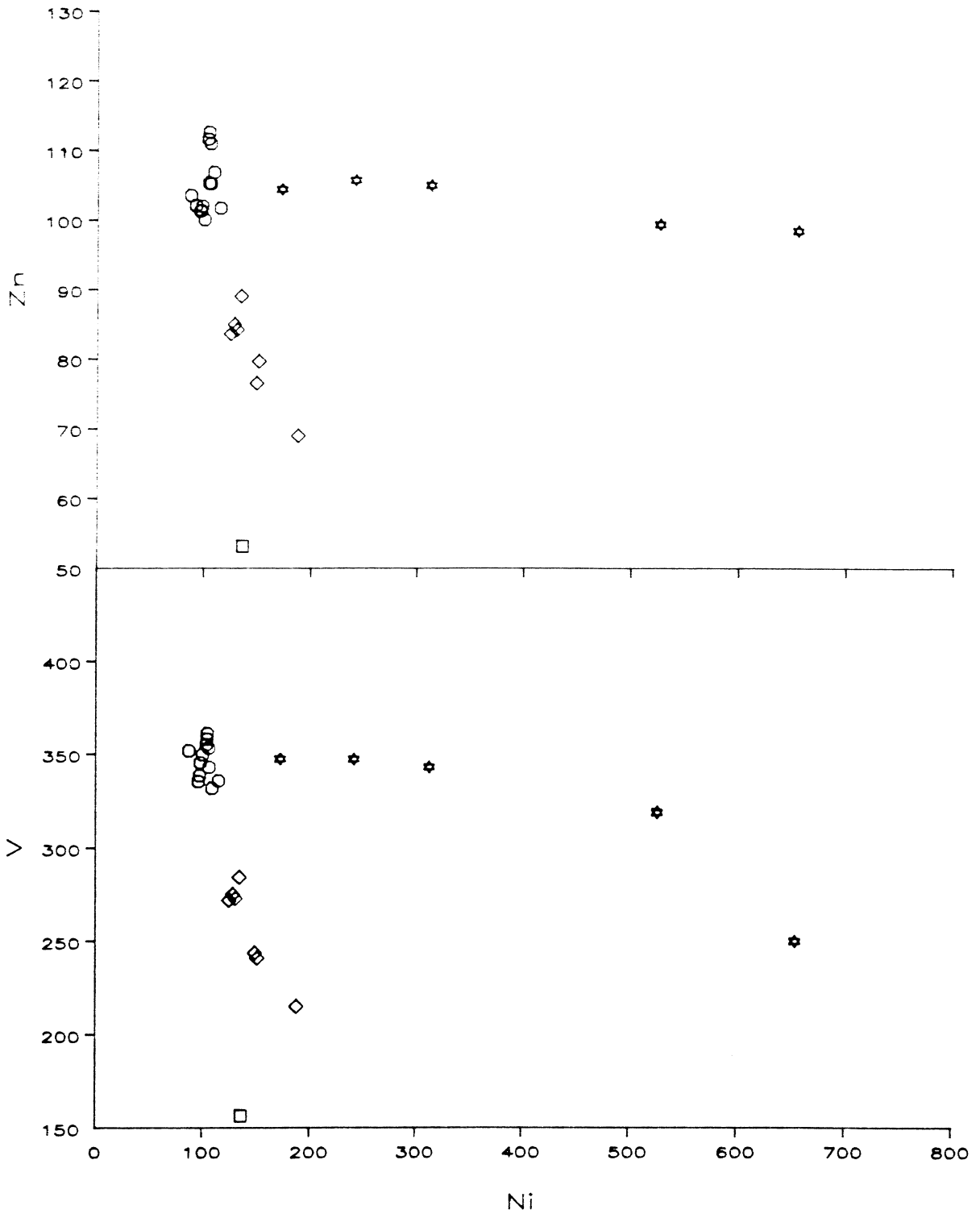


Figure 33. Zn and V vs Ni showing the distinct grouping by basalt type. Symbols as per Figure 10.

Rb (2.8 - 6.2 ppm) and Ba (15 - 29 ppm) occur at slightly higher levels in these basalts than in average N-MORB (Rb = 1.3 ppm, Ba = 9.3 ppm; Erlank and Kable, 1976), and show strong positive correlations with increasing degree of fractionation (i.e. decreasing Mg#; see Figure 31). The lack of scatter in the Ba and Rb data and consistency of Ba/Nb ratios (see Table 12) indicate that sea water alteration has not significantly affected the Group B basalts. Sr ranges in concentration from 120 - 134 ppm and shows strong negative correlations with increasing degree of evolution (Figure 34) indicating that plagioclase is one of the phases controlling the Group B fractionation trend (see Figures 12 & 22). Sc shows little variation in the Group B basalts (40-42ppm) (see Figure 35).

3. Group C

Zr, Nb and Y occur in these basalts at levels characteristic of N-type MORB (Zr = 112 - 135 ppm, Nb = 5.6 - 7.8 ppm, Y = 40 - 47 ppm) and show well defined negative correlations when plotted against Mg# (see Figure 30). The Group C basalts have Rb (4.8 - 6.1 ppm) and Ba (44 - 52 ppm) occurring at slightly higher concentrations than in the other basalt groups (A, B & D) and in typical N-type MORB (Erlank & Kable, 1976). They show good negative correlations with Mg# (see Figure 31). The lack of significant scatter in the Rb and Ba data, and consistency of Ba/Nb ratios indicates that there has not been significant sea water alteration of these rocks. High degrees of evolution are required to explain these high concentrations. LIL element enrichment as a result of the geochemical enrichment of the source regions of these samples is negated by Zr/Nb ratios from

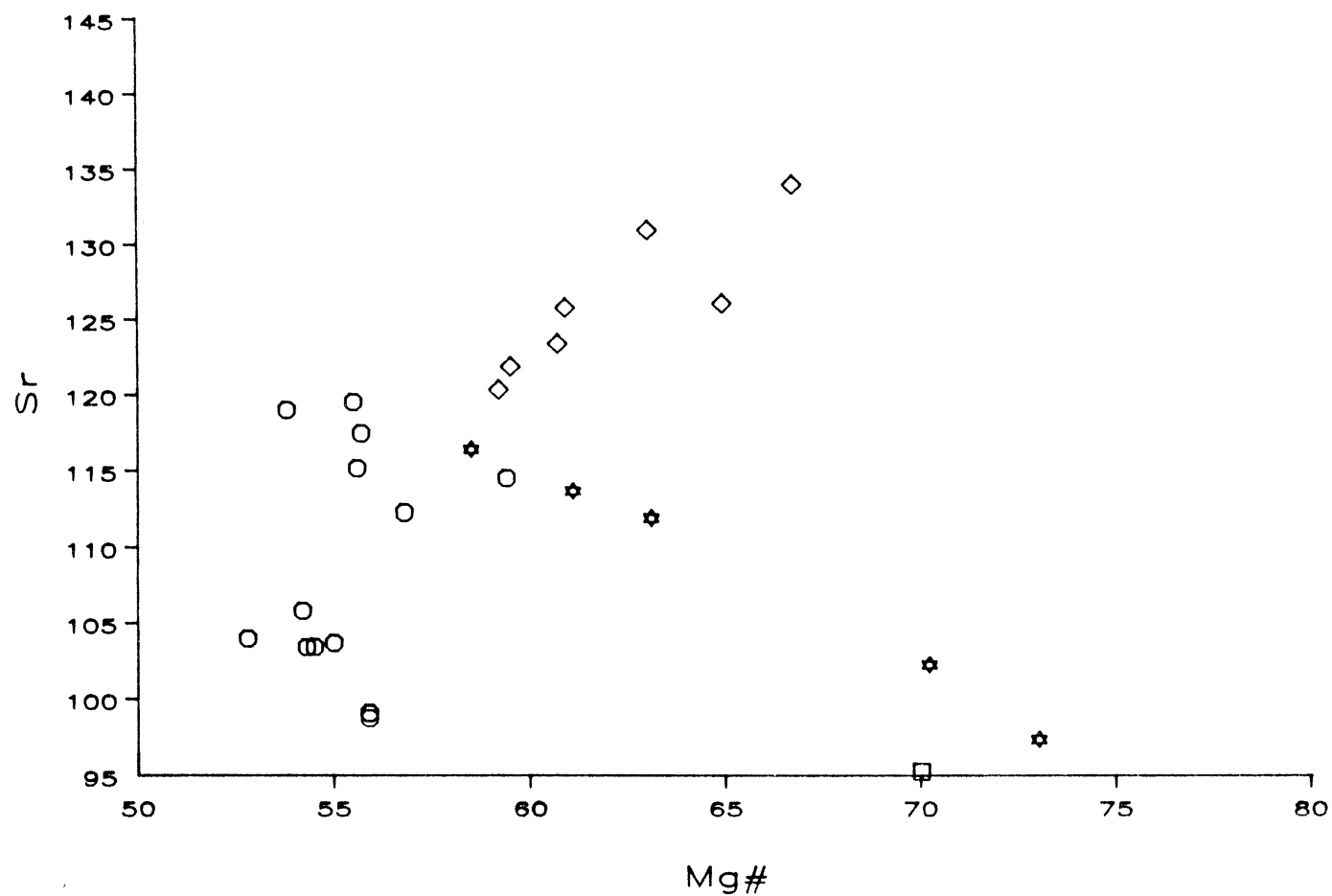


Figure 34. Sr vs Mg#. This particularly illustrates the low Sr content of AG32-3-5 (square). Symbols as per Figure 10.

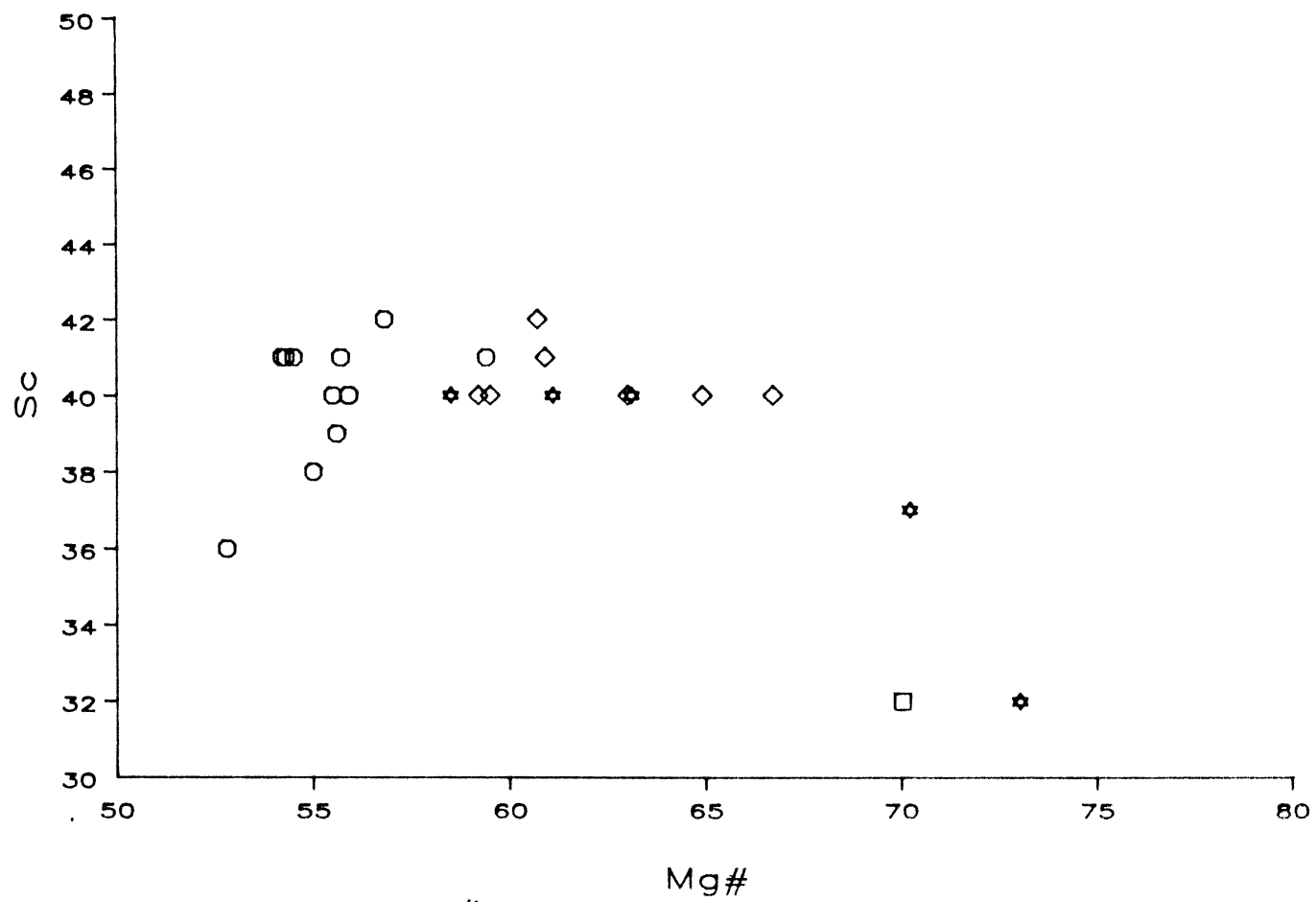


Figure 35. Sc vs Mg#.

16.7 to 21.0; values characteristic of geochemically depleted mantle (le Roex et al, 1981, 1983). Sr concentrations range from 97 - 116 ppm and there is a strong negative correlation with Mg# (see Figure 34), which indicates that plagioclase has played a very minor role in the formation of the Group C basalts.

Co in these basalts correlates positively with Mg# (Figure 36) and with Ni and Cr (Figure 37), suggesting that it substitutes for Ni or Cr in olivine or Cr-spinel, resulting in the high concentrations found (51 - 69 ppm). Zn and V occur at levels typical of N-MORB (Zn = 98 - 106 ppm, V = 250 - 347 ppm).

Of the ferromagnesian trace elements, Cr and Ni occur at high to extremely high levels (Cr = 502 - 1079 ppm, Ni = 173 - 653 ppm) relative to average N-type MORB (le Roex, 1980). In contrast, Cu occurs at levels similar to N-MORB (45 - 53 ppm). All 3 of these trace elements define clear trends when plotted against Mg# (Figures 38, 39 and 40). Cr and Ni are very highly compatible in olivine and Cr-spinel (K_D 's of up to 17 for Ni, and 3.07 for Cr in olivine and 507 for Cr in Cr-spinel have been reported (Leeman and Scheidegger, 1977; Bird, 1977)) and the excessively high levels of these elements (far higher than expected for "primary magmas" (le Roex et al, 1983, 1985; Hart and Davis, 1978; Sato, 1977)) is a result of the high proportion of olivine phenocrysts in these rocks.

The porphyritic nature and apparent olivine control on the composition of the Group C basalts raises an important question: Are these basalts derived by the fractional crystallization of

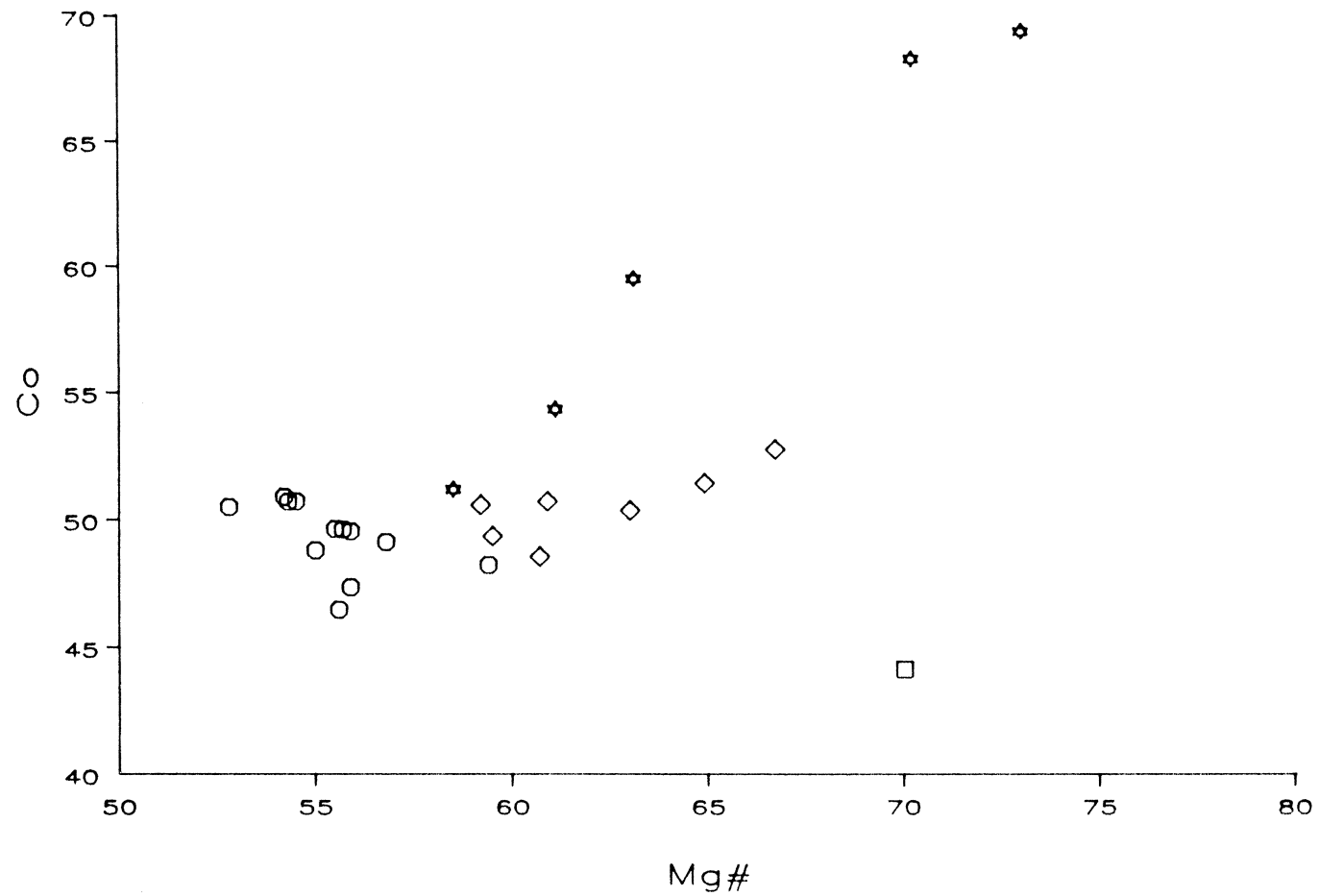


Figure 36. Co vs Mg# showing positive correlations for Group B and Group C basalts. Symbols as per Figure 10.

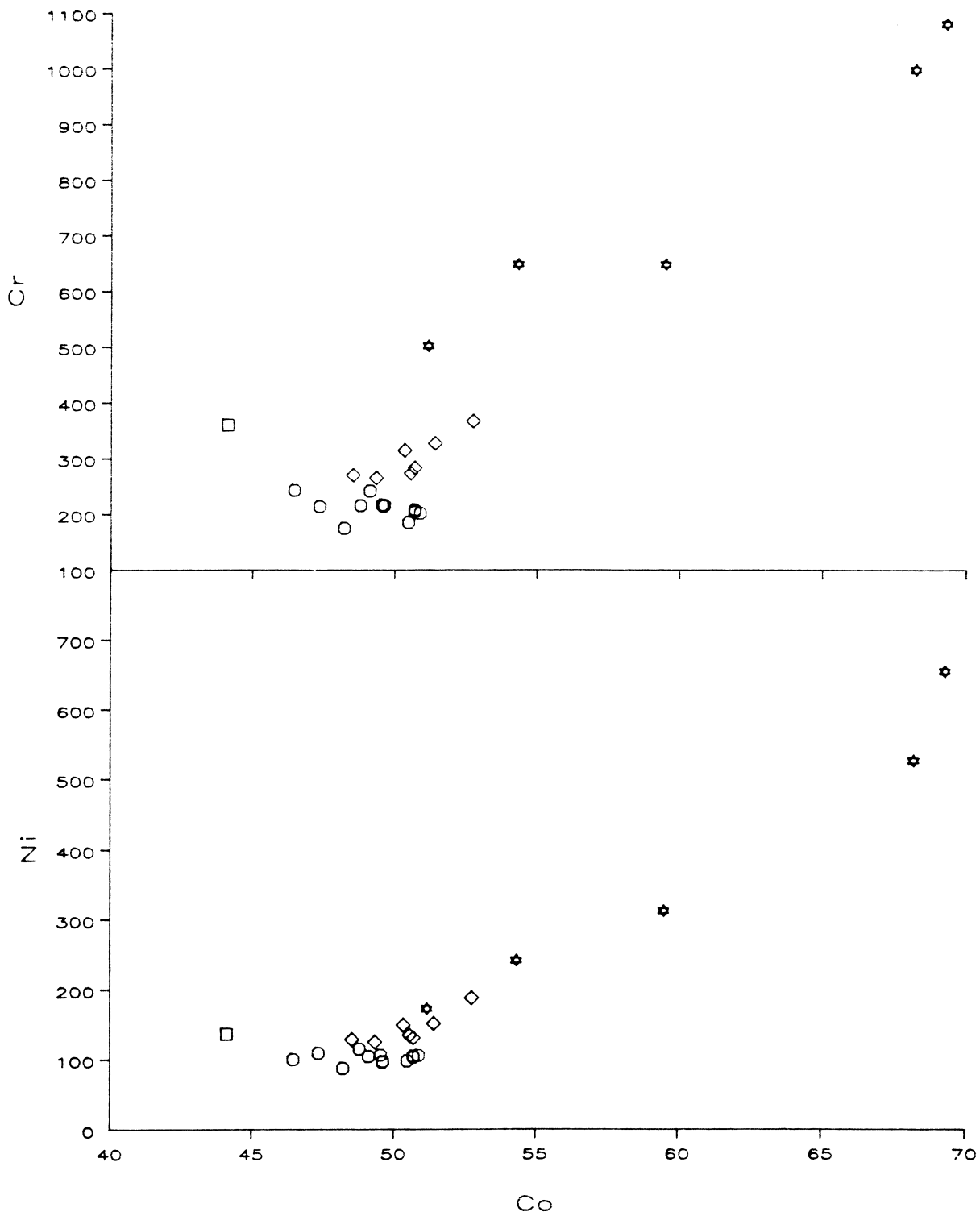


Figure 37. Cr and Ni vs Co showing positive correlations, particularly in the Group C basalts. This illustrates the substitution of Co for Cr and Ni. Symbols as per Figure 10.

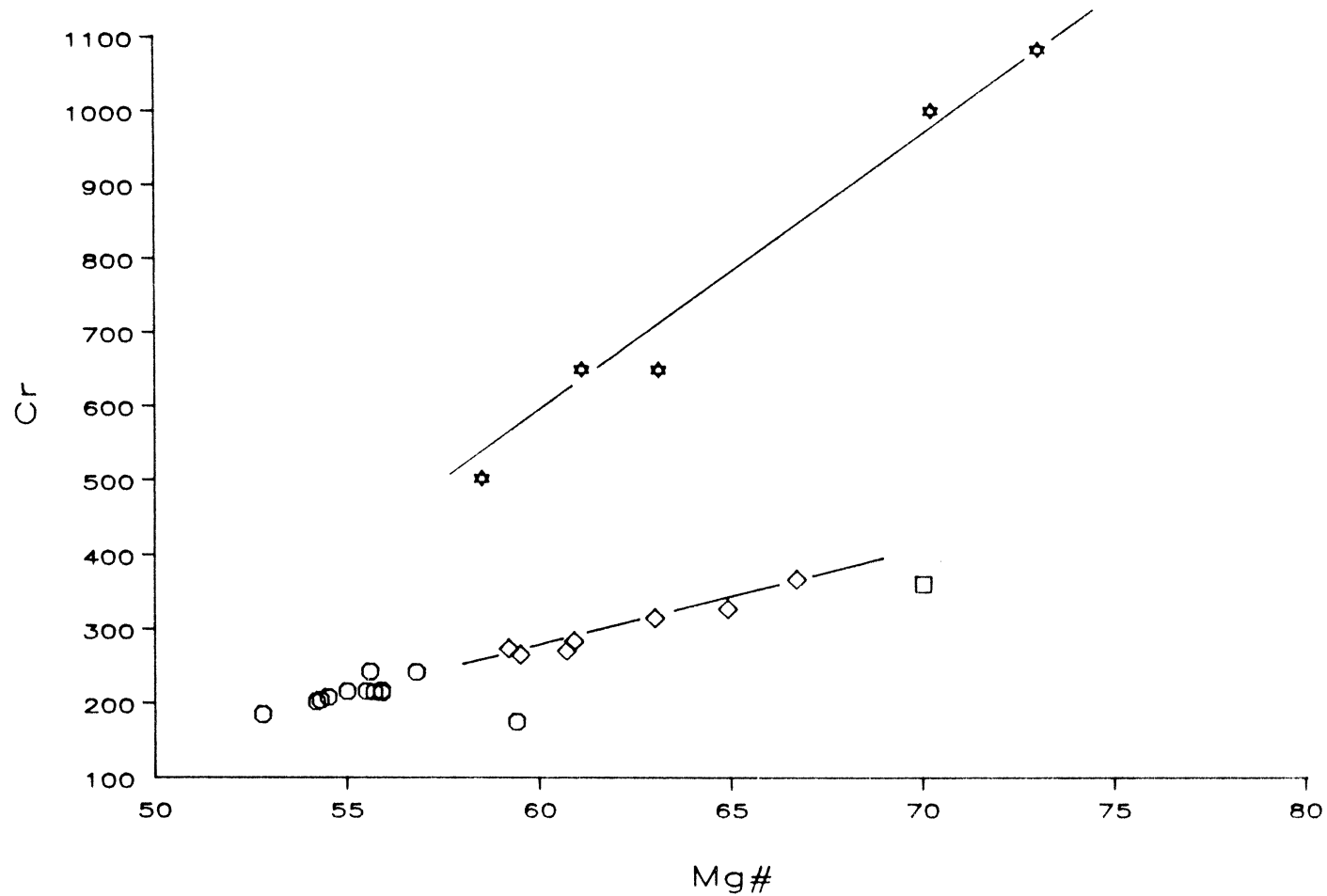


Figure 38. Cr vs Mg#. Distinct groups and clear trends are evident, indicative of Cr-Spinel fractionation. Symbols as per Figure 10.

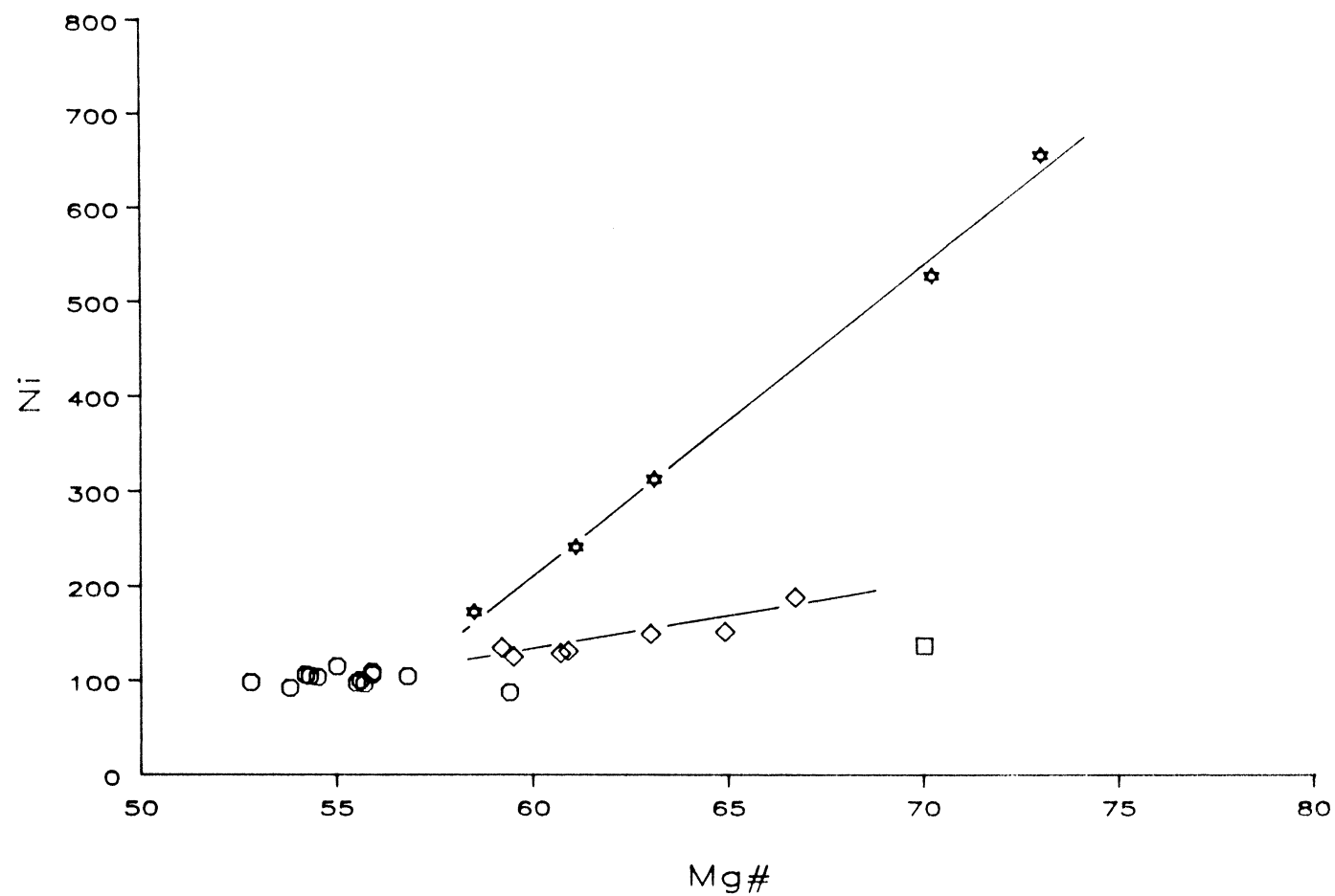


Figure 39. Ni vs Mg# showing clear trends and Ni enrichment in the Group C basalts. Symbols as per Figure 10.

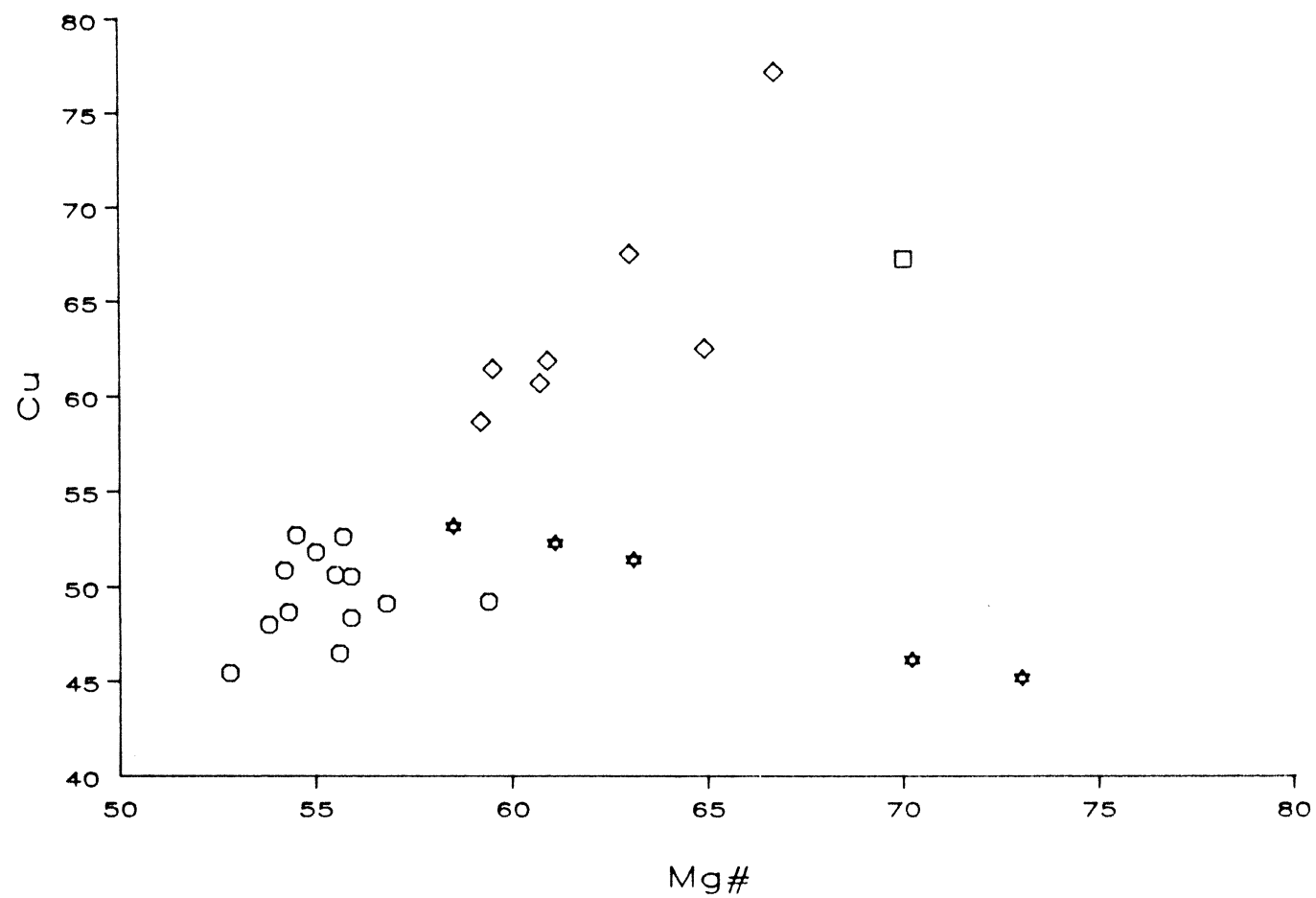


Figure 40. Cu vs Mg#. Symbols as per Figure 10.

olivine, or by the olivine accumulation? This is a problem that the ferromagnesian trace elements, and particularly Ni can help resolve. On Figure 41, theoretical Ni variation with Zr for closed system fractional crystallization is shown. The olivine phyric basalts can be seen to define a distinct linear array ($r = 0.99$) well off the theoretical fractionation trend. Accumulation trends (mixing lines) are straight lines on such plots (Hart and Davis, 1978; Cocherie, 1986) and this clearly illustrates the accumulative nature of the Group C olivine phyric basalts.

4. Group D

The single sample comprising this group is characterised by very low concentrations of the immobile incompatible elements Zr (35 ppm), Nb (1.8 ppm) and Y (18 ppm). Rb does not occur at detectable levels, and 8.9 ppm Ba are found. Sr (95 ppm) concentration is low for a sample that is highly plagioclase phyric. Cr, Ni and Cu occur at levels characteristic of a primitive N-type MORB (Cr = 360 ppm, Ni = 136 ppm, Cu = 67 ppm). Of all four basalt groups A, B, C & D, the lowest concentrations of Co (44 ppm), V (156 ppm), Zn (53 ppm) and Sc (32 ppm) occur in Group D. The Zr/Nb ratio of 19.6 is characteristic of geochemically depleted MORB, however this sample is less depleted than all the Group A and Group B, and many of the Group C samples.

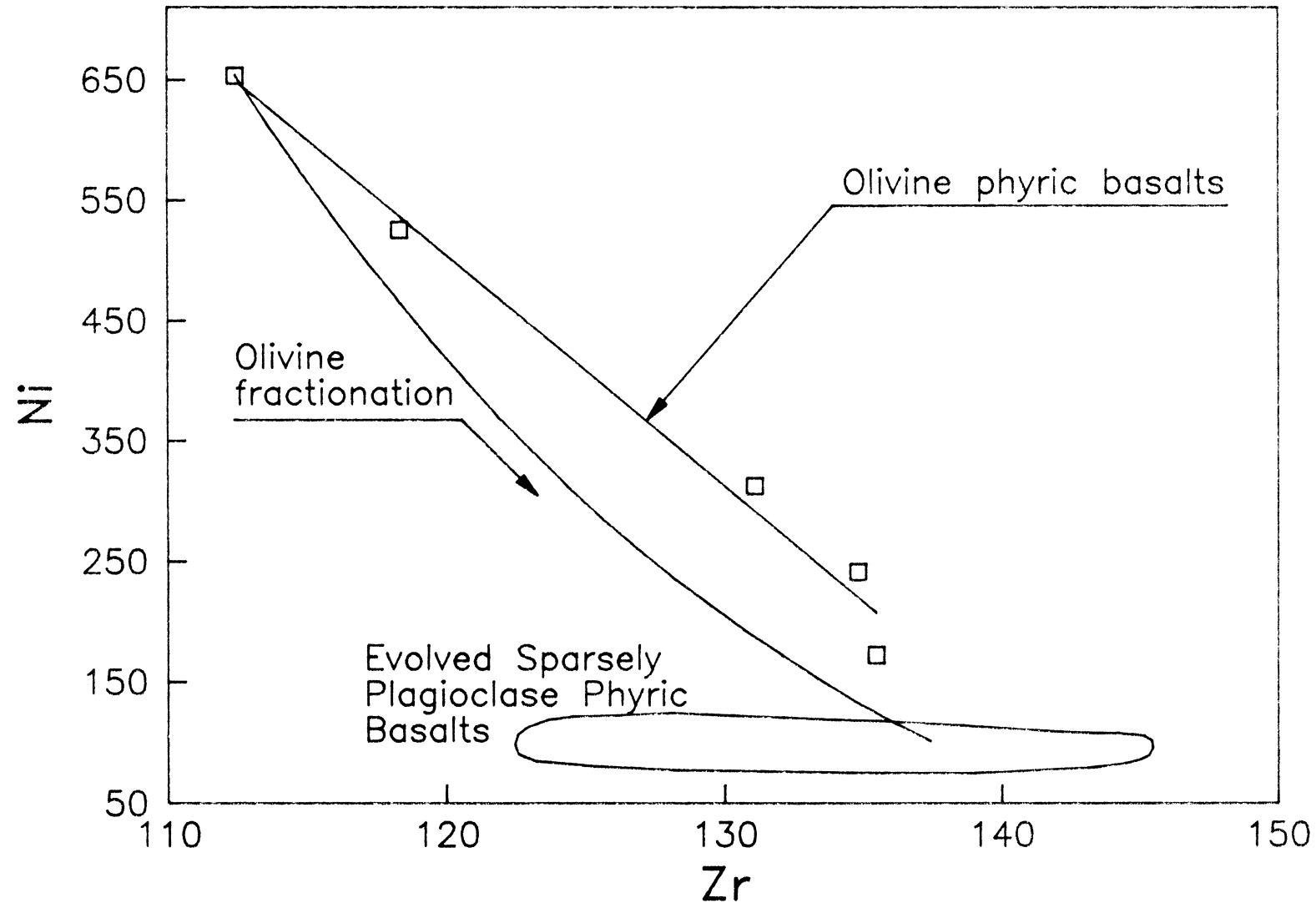


Figure 41. Ni vs Zr for the Group C basalts (squares) with a theoretical olivine fractionation curve calculated in 1% crystallization steps and using $D^{Zr} = 0.01$ and D^{Ni} from Hart and Davis (1978). Also shown is the field enclosing the Group A basalts. The trend through the Group C basalts is a least squares linear regression ($r = 0.99$) and clearly illustrates that these basalts lie on a linear "mixing" line rather than being fractionation products.

Between Group Trace Element Variations

The incompatible elements, Zr, Nb, Y and Ba are not differentially affected by fractional crystallization (Pearce and Norry, 1979). Although absolute elemental concentrations may differ between different samples derived from the same parental magma, the ratios of incompatibles will remain essentially constant reflecting parental magma characteristics. Conversely, the compatible trace elements in basaltic phase assemblages (e.g. Ni, Zn in olivine, Sr in plagioclase, Cr in pyroxene and Cr-spinel, Sc in clinopyroxene and Zn and V in titanomagnetite) are of particular use in understanding within group fractionation processes.

On the basis of incompatible trace element ratios, a minimum of three parental magmas can be identified. The first of these possibly giving rise to the Group A and Group C lavas, with Zr/Nb ranging from 20 to 28 (mean = 22.9) in the Group A and from 16.7 to 21 (mean = 18.2) in Group C. These ratios are the same, within analytical error, as are the ratios Zr/Y (2.6 - 2.9, Group A; 2.8 - 2.9, Group C) and Y/Nb (7.4 - 9.7, Group A; 5.8 - 7.3, Group C). The second parental magma forms the Group B basalts, and is characterised by significantly higher Zr/Nb (27.0 - 38.0, mean = 30.03) and by lower mean Zr/Y (2.54) and higher Y/Nb (11.4) ratios. On the basis of major element chemistry and Ba/Nb ratios, a third parental magma is proposed for the Group D basalts. This is characterised by Zr/Nb of 19.6, Zr/Y of 2.0 and Y/Nb of 9.8; values similar to those identified for the magma forming Groups A and C. It is,

however distinguished by a lower Ba/Nb ratio (4.94) than the other two parental magmas (Group A and Group C, mean Ba/Nb = 7.27; Group B, mean Ba/Nb = 8.14).

It can be seen from the covariation of Ba and Zr (Figure 42) that the Group A and Group C parental magmas cannot be related. These two basalt groups define distinct, parallel linear trends, with the Group C basalts having higher Ba concentrations than the Group A basalts at the same Zr concentration. Thus similar, but different parental magmas are required for these two groups.

All postulated parental magmas are characteristic of normal (N-type) MORB (le Roex et al, 1981), although those giving rise to basalts from Groups A, C and D are slightly less depleted in incompatible trace element abundances.

Trace element profiles (see Figure 43) illustrates evidence for 3 distinct parental magmas. In these, trace element concentrations are normalised to depleted MORB from the AAR (le Roex et al, 1985). Groups A and C have very similar profiles, differing significantly only in Cr and Ni (both compatible elements in olivine and Cr-spinel), consistent with these two basalt groups being derived from similar parental magmas, but having experienced different differentiation histories. Group B profiles show considerably more scatter than those from Groups A and C, with marked depletion in elements between Y and Nb in Figure 43 compared to depleted MORB from the American - Antarctic Ridge. The profile shapes are also in general different from those for Groups A and C. The Group D profile

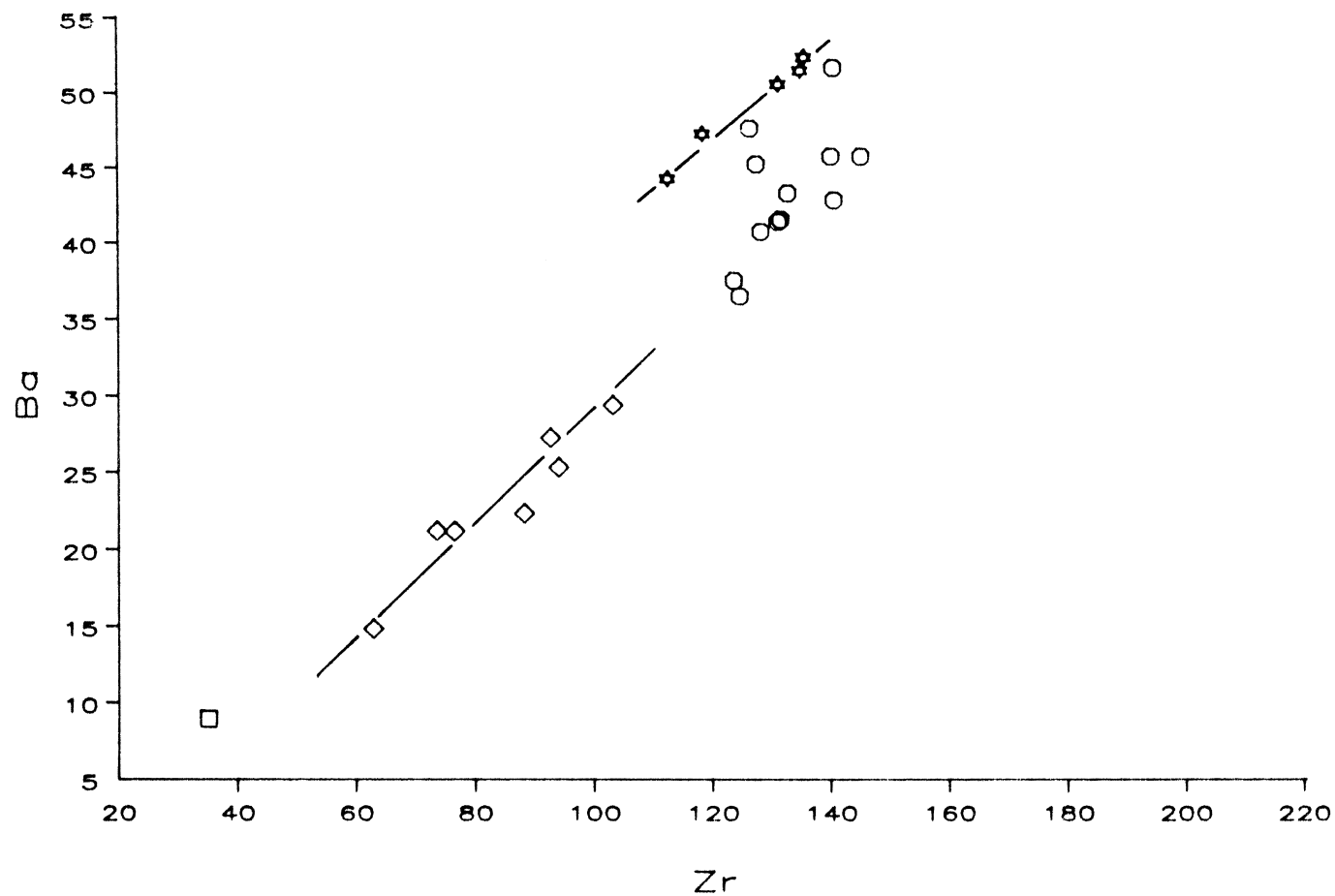


Figure 42. Ba vs Zr. The distinct trends defined by the different basalt Groups preclude any genetic relationships between them. Thus Group A and Group C basalts cannot be products of the same primary magma.

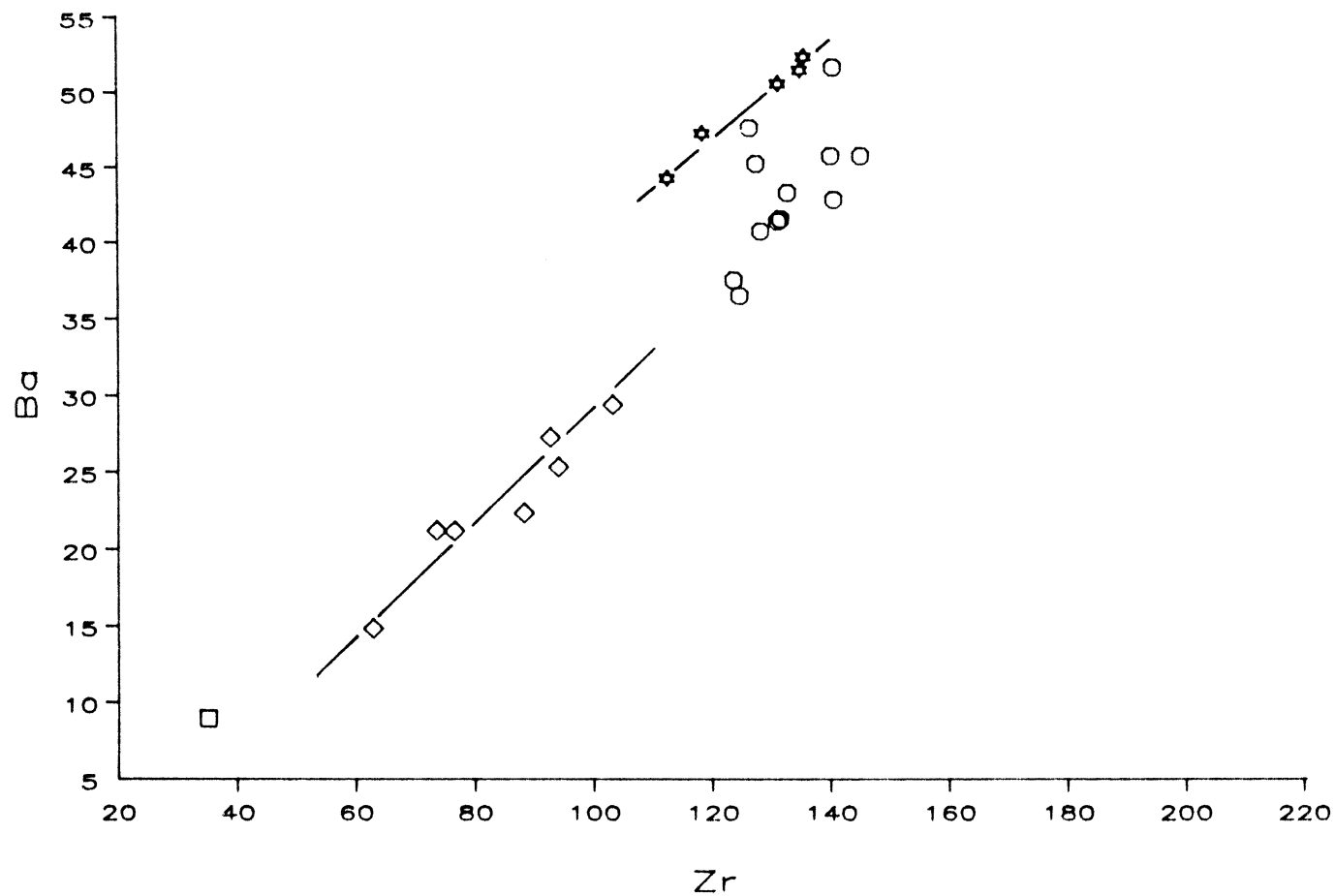


Figure 42. Ba vs Zr. The distinct trends defined by the different basalt Groups preclude any genetic relationships between them. Thus Group A and Group C basalts cannot be products of the same primary magma.

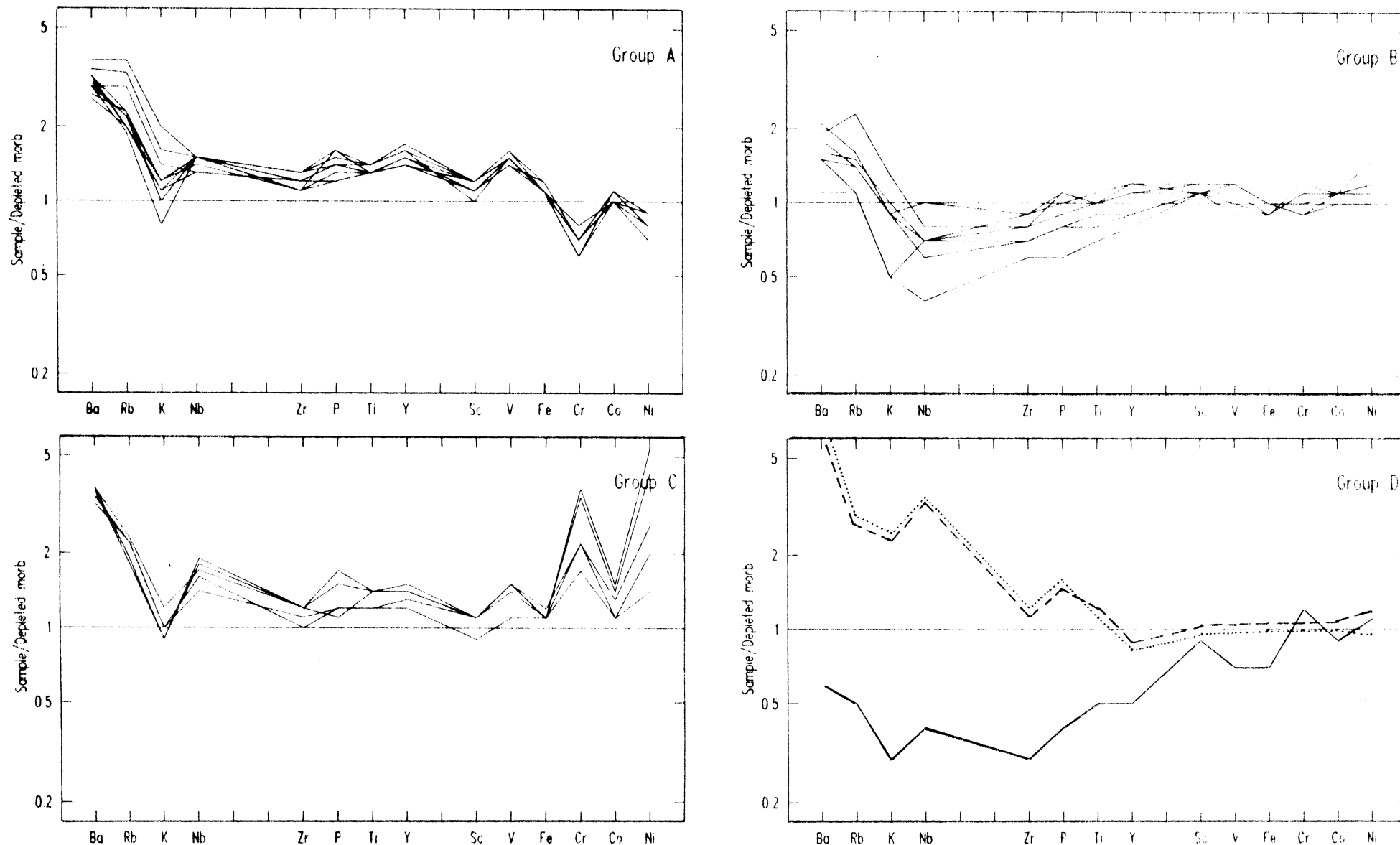


Figure 43. Trace element profiles, normalised to depleted MORB from the SWIR (le Roex et al, 1985). Normalising values are: Ba = 13.9, Rb = 2.64, K = 0.24, Nb = 4.03, Zr = 109, P = 0.17, Ti = 1.49, Y = 32, Sc = 35, V = 228, Fe = 9.02, Cr = 293, Co = 47, Ni = 121. K, P, Ti and Fe in wt% oxides, rest in ppm. The similar profiles for the Group A and Group C basalts indicate a similar source- which is distinct from the Group B and Group D primary magmas. Also shown on the Group D plot are average enriched MORB from the AAR (dotted line) and from the SWIR (dashed line) (le Roex et al, 1985). See chapter V for discussion.

differs not only in shape from the Groups A,B and C profiles, but also in position on figure 43. For comparison, average 'enriched' MORB from the SWIR and AAR (le Roex et al, 1985) are plotted along with the Group D profile. These are further discussed in chapter V.

Behaviour of the compatible trace elements during fractionation will be discussed in the chapter on the petrogenesis of the different basalt suites (Ch. V).

Chapter Five

P E T R O G E N E S I S

A major part of this thesis is an investigation into the possible origins of the four different basalt types discussed in the previous sections, and an analysis of the possible causes of within and between Group variations. Many processes can alter the composition of a magma formed in the oceanic mantle before it erupts as MORB. Some of these are listed in Table 16. Of these crystal fractionation (in its many varieties) and partial melting are the most commonly invoked to explain the ranges in composition in the oceanic environment (e.g. le Roex and Dick, 1981; le Roex et al, 1981, 1983; Baxter et al, 1985; Thompson et al, 1985, Bryan et al, 1981; Bryan and Moore, 1977).

Fractional crystallization may be defined as a process capable of producing substantial changes in magma composition, by the continuous isolation of the crystals in a cooling parental magma. It is expected to occur (Basaltic Volcanism Study Project, 1981). Magmas isolated from their heat sources will cool, producing crystals with densities and transport properties differing from those of the liquid, so that flow or gravitational settling in magma chambers can lead to their separation. Compositional paths for residual liquids are commonly represented on variation diagrams (e.g. Figure 24) or on AFM ternary diagrams (Figure 23). The most common paths for

MELTING	CRYSTALLIZATION	MIXING	FLUID SEPARATION
<u>Equilibrium</u>	<u>Equilibrium</u>	<u>Equilibrium</u>	<u>Equilibrium</u>
Batch	Bulk equilibrium	Two or more magmas	Immiscible liquids
Incremental	<u>Kinetic control</u>	Wall-rock equilibration	Liquid-vapor
Each increment a flow	Surface equilibrium	Fluid addition	Solid vapor
Increments pooled	Rate effects of	<u>Kinetic control</u>	<u>Kinetic control</u>
Inhomogeneous source	Nucleation	Two or more magmas	Immiscible liquids
<u>Kinetic control</u>	Quenching	Wall-rock assimilation	Liquid vapor
Sudden local heating or	Supercooling	or extraction	Solid-vapor
pressure release &	Crystal settling	Xenocrysts and xenoliths	Flow separation of crystals
melt separation	Crystal flotation		during transport
Uniform heating, rapid	Crystal plating		
melt separation	Diffusion		

Table 16: Some processes that can affect magma composition. From Basaltic Volcanism Study Project (1981) p.399

mafic liquids lead towards decreasing Mg#, decreasing Ca/Na and increasing silica (Basaltic Volcanism Study Project, 1981).

In order to test whether the major element variations observed in the lavas may be explained by fractional crystallization, use has been made of a least squares linear approximation program based on that of Bryan et al (1969), and written by A.R. Duncan (Geochemistry Department, U.C.T.). The program solves an overdetermined major element data matrix to test whether postulated parent and daughter compositions can be related by the addition or subtraction of various mineral vectors. The program is described in detail in le Roex (1980).

Input whole rock and glass data were normalised to 100%, volatile free and all iron expressed as FeO. Mean mineral compositions in the samples selected to represent parent and daughter compositions were used, unless there was a wide compositional range. In such cases, theoretical end-member compositions (e.g. pure Fo, pure Fa) were used for the initial mix. The relative proportions of the end members, calculated as best satisfying known compositional variations, were then combined to give the appropriate mineral composition which, if within the observed compositional range, was used in the final mixing calculation. The trace element Zr was used in all linear mixes as a "major element" in order to force F, the proportion of fluid remaining, into being the most realistic value, given trace element constraints on the mixing process.

Output from the program consists of the calculated dependent vector (parent composition as used here), oxide residuals and the sum of squares of oxide residuals, calculated proportions of independent vectors (daughter composition and fractionating mineral phases) and their standard deviations. These data were used to assess the acceptability of the mixing equation. Solutions were assumed to be potentially valid, only if the sum of squares of oxide residuals was less than 0.10 (e.g. Wright, 1974; le Roex, 1980), with no major discrepancies occurring in any of the individual oxide residuals. Additional criteria for acceptable mixes were (1) that proportions of independent vectors should sum close to 100%; (2) that the mixing equation be tightly constrained - i.e. standard deviations on individual vectors and the total standard deviation should be as low as possible; and (3) Most importantly, the calculated proportions of fractionating phases should be realistic in terms of observed modal proportions, and coefficients of mineral vectors should have the same sign. Examples of acceptable and unacceptable mixing approximations are presented in Tables 17 and 18 respectively.

Trace element variations can be used to check the validity of proposed major element fractionation models. Potentially valid models are those that satisfy both major element constraints and trace element variations.

The partitioning of trace elements in crystal - magma systems may be governed by surface equilibrium (i.e. Rayleigh Fractionation) or by total equilibrium (i.e. Nernst

INPUT DATA

	PARENT	PLAG	OLIV	CPX	ILM	TiMG	DAUGHTER
SiO ₂	50.71	53.38	35.28	48.68	0.15	0.23	55.62
TiO ₂	3.54	0.11	0.11	2.03	50.00	23.61	2.00
Al ₂ O ₃	15.56	28.68	0.06	3.90	0.07	0.48	15.90
FeO*	12.11	0.67	36.25	10.69	47.02	74.44	11.21
MnO	0.19	0.01	0.51	0.17	0.53	0.51	0.21
MgO	4.35	0.13	27.29	13.54	2.04	0.22	2.22
CaO	8.82	12.18	0.31	20.51	0.18	0.51	6.12
Na ₂ O	3.40	4.77	0.01	0.47	0.01	0.01	4.42
K ₂ O	1.32	0.08	0.01	0.01	0.01	0.01	2.30

LEAST SQUARES APPROXIMATION

	OBS	<u>PARENT</u> CALC.	DIFF.	VECTOR	<u>MIX</u> Wt. %	S.D.
SiO ₂	50.71	50.70	-0.01	DAUGHTER	56.19	0.91
TiO ₂	3.54	3.54	0.00	PLAG	21.17	0.74
Al ₂ O ₃	15.56	15.55	-0.01	OLIV	4.28	0.45
FeO*	12.11	12.11	0.00	CPX	13.61	0.52
MnO	0.19	0.19	0.00	ILM	3.61	0.30
MgO	4.35	4.36	0.01	TiMG	1.29	0.33
CaO	8.82	8.84	0.02			
Na ₂ O	3.40	3.56	0.16	TOTAL	100.16	1.43
K ₂ O	1.32	1.31	-0.01			

SUM OF SQUARES OF DIFFERENCES = 0.026

Table 17: An example of an acceptable major element least squares approximation relating two arbitrary magmas, 'PARENT' and 'DAUGHTER', via fractional crystallization of Olivine, Plagioclase, Clinopyroxene, Ilmenite and titanomagnetite. From le Roex (1980). Note low sum of squares of residuals and low standard deviations of "vectors".

INPUT DATA

as in Table 17

LEAST SQUARES APPROXIMATION

	<u>PARENT</u>				<u>MIX</u>	
	OBS.	CALC.	DIFF.	VECTOR	Wt. %	S.D.
SiO ₂	50.71	50.68	-0.03	DAUGHTER	56.97	5.57
TiO ₂	3.54	2.56	-0.98	PLAG	20.24	4.48
Al ₂ O ₃	15.56	15.48	-0.08	OLIV	2.31	2.59
FeO*	12.11	12.42	0.31	CPX	15.12	3.12
MnO	0.19	0.18	-0.01	TiMGT	4.63	1.10
MgO	4.35	3.98	-0.37			
CaO	8.82	9.08	0.26	TOTAL	99.27	8.29
Na ₂ O	3.40	3.56	0.16			
K ₂ O	1.32	1.33	0.01			

Sum of squares of differences = 1.29

Table 18: An example of an unacceptable major element least squares approximation from le Roex (1980). Input data as in Table 17, with the exception of ilmenite as a fractionating phase. Note high sum of squares of residuals and high standard deviations on the 'vectors'.

Fractionation). Gast (1968) produced equation (1) below to describe trace element behaviour during surface equilibrium.

$$\frac{C_a^l}{C_a^o} = F^{\overline{D}_a - 1} \quad (1)$$

F = weight fraction of residual liquid. It is calculated here in the major element model as the proportion of daughter required to satisfy least squares approximation.

C_a^l = concentration of trace element a in residual liquid.

C_a^o = concentration of trace element a in initial liquid.

$\overline{D}_a$ = bulk distribution coefficient for element a .

The bulk distribution coefficient is a function of all the phases crystallizing from the initial liquid, such that

$$\bar{D}_a = \sum_i D_i^a \cdot X_i \quad (2)$$

Where D_i^a = distribution coefficient of element a in mineral i.

X_i = weight fraction of mineral i, as calculated from the major element model.

Equations (1) and (2) assume constant D. Greenland (1970) presented equation (1) assuming linear variation in D with crystallization (see equation 3).

$$\ln \frac{C'}{C_0} = (D-1) \ln F + D(F-1) \quad (3)$$

In this work, equation (1) has been used, because application of equation (3) has been limited by lack of knowledge on the variation of D with crystallization and since differences produced by using equation (3) rather than equation (1) are within uncertainties in distribution coefficients.

The above equation allows quantitative modelling of trace element variations during fractional crystallization. To model trace element variations during crystal accumulation, the mass balance equation (4) was used.

$$C_a^p = C_a^o \cdot X_o + \overline{X_i} \overline{C_a} \quad (4)$$

C_a^p = predicted concentration of trace element a in cumulus enriched magma.

C_a^o = concentration of trace element a in initial magma.

X_i = weight fraction of phenocryst assemblage added to the cumulus enriched magma, as calculated in the major element model.

X_o = weight fraction of the original magma, in the cumulus enriched magma, as calculated in the major element model

$\overline{C_a}$ = average concentration of element a in the phenocryst assemblage.

C_a is calculated from equation (5) (le Roex and Reid, 1978)

$$\overline{C_a} / C_a^o = \frac{1-F^D}{1-F} \quad (5)$$

The above equations assume that the volumes of initial and cumulus magmas are the same; that phenocrysts were homogeneously distributed throughout the magma; and that at the instant that the weight fraction of the residual liquid was equal to F, all the phenocrysts migrated to the cumulus enriched magma (le Roex and Reid, 1978).

In addition to simple closed system fractional crystallization and crystal accumulation, a number of more complex models can be proposed to explain trace element variations. These include open system fractional crystallization (O'Hara, 1977), zone refining (Cox et al, 1979) and compensated crystal settling (Krishnamurthy and Cox, 1977), or other petrological processes such as magma mixing (Anderson, 1976; O'Hara and Matthews, 1981).

In this study, most samples are characterised by zoned phenocrysts and thus Rayleigh fractionation has been adopted for modelling trace element variations assuming, as a first approximation, simple closed system fractional crystallization. Applicable mineral distribution coefficients are presented in Table 19. A program (DARTS, written by A.R. Duncan, Geochemistry Dept. U.C.T.) was used for the quantitative modelling of trace element behaviour during closed system fractional crystallization.

	OLIVINE	SOURCE	PLAGIOCLASE	SOURCE	CLINOPYROXENE	SOURCE
Zr	0.01	(1)	0.01	(1)	0.03 - 0.22	(2)
Nb	0.01	(1)	0.01	(1)	0.01 - 0.04	(2)
Y	0.01	(1)	0.013 - 0.031	(3)	0.5	(1)
Rb	0.008	(5)	0.05 - 0.12	(4)	0.004 - 0.08	(4)
Ba	0.005 $\pm$ 0.003	(6)	0.18 - 0.77	(7)	0.013 - 0.046	(5)
Sr	0.0012 - 0.072	(8)	1.27 - 2.84	(5)	0.11	(3)
Co	2.75 - 4.45	(9)	0.03	(10)	0.68 - 1.22	(9)
Cr	0.6 $\pm$ 0.2	(11)	0.05	(3)	4.72 - 11.43	(9)
Ni	(124/MgO) - 0.9	(12)	0.06	(13)	1.42 - 2.11	(9)
V	0.04 $\pm$ 0.02	(11)	0.04	(3)	0.74	(13)
Zn	0.86 or	(13)	0.13	(10)	0.5	(13)
Zn	2.13	(3)				
Cu	0.11	(13)	0.44	(13)	1.5-2.4	(14)
Sc	0.13 - 0.22	(9)	0.01	(10)	2.38 - 3.23	(9)

	Cr - SPINEL	SOURCE
Sc	0.048 $\pm$ 0.001	(15)
Ni	5.9 $\pm$ 0.02	(16)
Co	1.9 - 2.8	(16)
Cr	77	(17)
V	38	(17)

Table 19: Basaltic liquid - mineral distribution coefficients used in petrogenetic modelling. Sources as follows: (1) = Pearce & Norry (1979); (2) = McCallum and Charette (1978); (3) = le Roex (1980); (4) = Goodman (1972); (5) = Philpotts and Schnetzler (1970); (6) = McKay and Weill (1977); (7) = Dupuy and Coulon (1973); (8) = Griffen and Rama Murthy (1969); (9) = Dale and Henderson (1972); (10) = Paster et al (1974); (11) = Lindstrom (1976); (12) = Hart and Davis (1978); (13) = Bougault and Hekinian (1974); (14) = Seward (1971); (15) = Nagasawa et al, 1976; (16) = Bird (1971); (17) = Ringwood (1970)

Petrogenesis of the Group A Basalts

The relatively evolved Group A basalts defined tight clusters on major and trace element variation diagrams (See Figures 24, 30, 31, 34-36, 38-40). Potential parental magmas for the petrogenetic modelling of this group are represented by AG32-3-75 and AG32-3-50. The least evolved sample in this group (AG32-3-2) is precluded as it is distinct from the majority of the samples in this group on major and trace element variation diagrams (See Figures 24, 30, 31, 34-36, 38-40).

Using the parameters defined earlier for acceptable mixing calculations, it is not possible to model the Group A compositional variation as a simple closed system fractionation process involving olivine and plagioclase, or olivine, plagioclase and clinopyroxene. Given the mineralogical and geochemical similarities of the samples comprising Group A, they must thus be related by a more complex process. This is likely to be a fractional crystallization process as (1) the samples are olivine - plagioclase phyric (albeit sparsely), (2) they are relatively evolved ($Mg\# = 52.8 - 59.4$) and (3) an acceptable major element mix may be calculated for the compositional variation between an evolved sample (AG32-3-97) and a less evolved sample (AG32-3-75) (See Table 20). These two samples span the compositional range found in Group A. It is not possible to model the trace element variation between these two samples by closed system fractional crystallization (Table 21). Thus an open system fractionation model is proposed for the evolution of Group A. Further quantitative modelling was not pursued in view of the lack of a well

INPUT DATA

	OLIV-A	PLAG-A	CPX	AG32-3-97	AG32-3-75
SiO ₂	38.16	54.60	51.65	49.87	50.06
TiO ₂	.01	.01	.59	2.11	1.95
Al ₂ O ₃	.01	28.41	3.12	14.67	14.80
FeO	21.13	.88	8.55	11.84	11.40
MnO	.32	0.01	.15	.21	.18
MgO	39.31	.18	15.60	6.55	7.10
CaO	.30	12.00	19.50	10.96	11.00
Na ₂ O	.01	4.31	.19	2.89	2.88
K ₂ O	.01	.09	.01	.49	.28
ZR	.01	.01	.01	140.00	125.00

PARENT: AG32-3-75

DAUGHTER		DIFF.	% DIFF	VECTOR	COEFF	STD.DEV	
OBS	CALC						
SiO ₂	50.06	50.03	-.03	-.06	OLIV-A	.0218	.0029
TiO ₂	1.95	1.90	-.05	-2.47	PLAG-A	.0583	.0031
Al ₂ O ₃	14.80	14.84	.04	.30	CPX	.0288	.0047
FeO	11.40	11.33	-.07	-.61	AG32-3-97	.8929	.0006
MnO	.18	.20	.02	10.77			
MgO	7.10	7.17	.07	.93			
CaO	11.00	11.05	.05	.48			
Na ₂ O	2.88	2.84	-.04	-1.48			
K ₂ O	.28	.44	.16	58.31			
ZR	125.00	125.01	.01	.00			
TOTALS					1.0018	.0063	

SUM OF SQUARES OF DIFFERENCES = .05

Table 20: An acceptable major element least squares approximation relating two Group A basalts: An evolved "daughter" (AG32-3-97) and a less evolved "parent" (AG32-3-75). Clinopyroxene composition from le Roex (1980).

$$F = 0.8929$$

FRACTIONATING PHASES: AS PER TABLE 20.
DISTRIBUTION COEFFICIENTS

	OLIVINE	PLAGIOCLASE	PYROXENE
Zr	.01	.01	.20
Nb	.01	.01	.03
Y	.01	.013	.50
Rb	.007	.05	.01
Ba	.002	.01	.01
Sr	.001	2.00	.12
Co	4.44	.03	1.22
Cr	.50	.05	5.75
Ni	12.97	.01	1.40
V	.04	.04	.74
Zn	.86	.13	.50
Cu	.11	.44	2.40
Sc	.13	.01	3.23

	STARTING (AG32-3-75)	TARGET (AG32-3-97)	PREDICTED
Zr	125.00	140.00	139.04
Nb	6.10	6.20	6.82
Y	45.00	51.00	49.60
Rb	5.40	9.90	6.03
Ba	36.00	51.00	40.28
Sr	117.00	104.00	115.65
Co	50.00	50.00	48.73
Cr	215.00	185.00	199.79
Ni	96.00	98.00	76.79
V	335.00	345.00	365.73
Zn	101.00	102.00	108.42
Cu	53.00	45.00	53.65
Sc	41.00	36.00	41.53

Table 21: A comparison of observed and predicted trace element contents in the Group A mixing approximation from Table 20. This is unacceptable due to the excessive differences between observed (Target) and predicted values for Ba, Sr, Ni, Cu and Sc.

constrained Group evolutionary trend and the poor constraints on open system processes and the comparatively small range in differentiation shown by the Group A lavas (< 10% enrichment in incompatibles).

Petrogenesis of the Group B Basalts

Unlike the Group A basalts, the samples comprising Group B define distinct trends in compositional space (see Figures 24, 30, 31, 34-36, 38-40), with Mg# ranging from 59.2 to 66.7 and Zr from 63 to 103 ppm. This evolutionary trend is defined by the sequence AG32-3-94 (least evolved) → AG32-3-113 → AG32-3-108 → AG32-3-105 → AG32-3-111 → AG32-3-63 → AG32-3-99. The very sparsely phyrlic to aphyric nature of these samples and the lack of significant sea water alteration affords the opportunity to quantitatively model major and trace element variations between the least evolved sample (the parent) in this group (AG32-3-94) and the most evolved sample (the daughter: AG32-3-99). A least squares approximation relating the most evolved sample to the least evolved is presented in Table 22. This model shows that the composition of AG32-3-99 can be explained by the removal of 6.58% olivine (Fo_{84}) + 22.20% plagioclase (An_{83}) + 9.42 % clinopyroxene (Wo_{41} En_{45} Fs_{14}) from a parent composition of AG32-3-94. The low sum of squares of residuals and the close agreement between the observed trace element contents of AG32-3-99 and those calculated from AG32-3-94, using the above model (Table 23) testify to the acceptability of this model. In trace element modelling a D for Ni in olivine of

INPUT DATA

	OLIVINE (Fo76)	PLAGIOCLASE (An83)	CPX (Wo41En45Fs14)	AG32-3-99	AG32-3-94
SiO ₂	38.89	52.12	51.65	49.23	48.99
TiO ₂	.01	.01	.59	1.63	1.08
Al ₂ O ₃	.03	30.12	3.12	15.85	16.76
FeO	21.66	.72	8.55	10.80	9.04
MnO	.36	.01	.15	.17	.16
MgO	39.36	.31	15.60	7.72	8.94
CaO	.23	13.95	19.50	11.43	12.14
Na ₂ O	.01	3.11	.19	2.67	2.55
K ₂ O	.01	.02	.01	.22	.13
ZR	.01	.01	.01	103.00	63.00

PARENT: AG32-3-94

DAUGHTER

	OBS	CALC	DIFF	% DIFF	VECTOR	COEFF	STD.DEV
SiO ₂	48.99	49.11	.12	.24	Olivine	.0658	.0045
TiO ₂	1.08	1.06	-.02	-2.28	Plagioclase	.2220	.0049
Al ₂ O ₃	16.76	16.68	-.08	-.50	CPX	.0942	.0073
FeO	9.04	9.00	-.04	-.04	AG32-3-99	.6116	.0013
MnO	.16	.14	-.02	-9.99			
MgO	8.94	8.85	-.09	-.99			
CaO	12.14	11.94	-.20	-1.65			
Na ₂ O	2.55	2.34	-.21	-8.16			
K ₂ O	.13	.14	.01	8.15			
ZR	63.00	63.00	-.00	-.00			
					TOTALS	.9936	.0100

SUM OF SQUARES OF DIFFERENCES = .11

Table 22: An acceptable major element least squares approximation relating the least evolved Group B basalt (AG32-3-94) to the most evolved Group B basalt (AG32-3-99) by the fractional crystallization of Olivine, Plagioclase and Clinopyroxene. The Clinopyroxene composition is from le Roex (1980) and Olivine and Plagioclase compositions from Group B basalts.

F = 0.6116

FRACTIONATING PHASES: AS PER TABLE 22.

DISTRIBUTION COEFFICIENTS

	OLIVINE	PLAGIOCLASE	PYROXENE
Zr	.01	.01	.20
Nb	.01	.01	.03
Y	.01	.013	.50
Rb	.007	.05	.01
Ba	.002	.01	.01
Sr	.001	2.00	.12
Co	4.44	.03	1.22
Cr	.50	.05	5.75
Ni	8.00	.01	1.40
V	.04	.04	.74
Zn	1.80	.13	.50
Sc	.13	.01	3.23

	STARTING AG32-3-94	TARGET AG32-3-99	PREDICTED
Zr	63.00	103.00	100.17
Nb	1.60	3.90	2.60
Y	26.00	39.00	39.83
Rb	2.80	4.20	4.51
Ba	15.00	29.00	24.42
Sr	134.00	120.00	121.92
Co	53.00	51.00	50.88
Cr	366.00	273.00	281.68
Ni	188.00	134.00	131.35
V	215.00	284.00	316.65
Zn	69.00	89.00	87.84
Sc	40.00	40.00	43.61

Table 23: A comparison of observed and predicted trace element contents, calculated from the major element model of Table 22.

between 8 and 11 has been selected as reasonable in preference to D's calculated from the equation of Hart and Davis (1976). This is in accordance with measured NiO contents in the Group B olivines (mean = 0.19 wt%) and Ni contents of the Group B basalts (Mean = 144 ppm). Cu has not been modelled, due to its strongly lithophile nature and consequent susceptibility to the effects of sulphide crystallization. The apparently large absolute deviations between observed and predicted V concentrations is larger than is desirable but has been assumed to be moderately acceptable given the sometimes large uncertainties associated with the modelling of lithophile elements.

The remaining Group B samples can be related to the parent magma by a similar fractionation scheme involving up to 6.4% olivine; 19.0% plagioclase and 7.7% clinopyroxene. Linear mixing calculations illustrating this are presented in Table 24. In general, the close agreement between observed the calculated trace element concentrations (summarised in Table 25) testify to the acceptability of these models.

A paradoxical situation arises upon comparison of the observed phenocryst phases with those predicted by the major element model; the lack of clinopyroxene in the phenocryst assemblage is immediately apparent. This clinopyroxene paradox commonly occurs in the quantitative modelling of MORB genesis (i.a. Francis, 1986; Basaltic Volcanism Study Project, 1981) and it is usually explained as being the result of high pressure fractionation of clinopyroxene which sinks to form a pyroxene cumulate. This is supported by experimental studies which show

Target Composition	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	Zr
(AG32-3-94)	48.99	1.08	16.76	9.04	0.16	8.94	12.14	2.55	0.13	63
AG32-3-113	48.59	1.23	16.32	9.60	0.18	8.70	11.80	2.48	0.12	73
+ 2.1% Olivine, 8.7% Plagioclase										
+ 3.8% Clinopyroxene	49.02	1.08	16.75	9.02	0.17	8.92	12.09	2.41	0.10	63
Difference	0.03	-	-0.01	-0.02	0.01	-0.02	-0.05	-0.14	-0.03	-
AG32-3-108	48.86	1.27	16.21	9.63	0.17	8.07	12.19	2.52	0.21	76
+ 4.5% Olivine + 11.0% Plagioclase										
+ 2.6% Clinopyroxene	49.05	1.06	16.74	9.15	0.16	8.85	12.08	2.42	0.17	63
Difference	0.04	-0.02	-0.02	0.11	-	-0.09	-0.06	-0.13	0.04	-
AG32-3-105	48.67	1.43	15.92	10.18	0.19	7.74	11.59	2.49	0.23	87
+ 5.4% Olivine + 16.3% Plagioclase										
+ 6.2% Clinopyroxene	49.13	1.08	16.66	9.11	0.17	8.77	11.91	2.33	0.17	63
Difference	0.14	-	-0.10	-0.07	-0.01	-0.17	-0.24	0.22	0.04	-
AG32-3-111	49.13	1.51	15.91	10.33	0.18	7.44	11.60	2.68	0.31	92
+ 6.4% Olivine + 18.4% plagioclase										
+ 6.9% Clinopyroxene	49.16	1.07	16.61	9.11	0.16	8.73	11.84	2.41	0.21	63
Difference	0.17	-0.01	-0.15	0.07	-	-0.21	-0.30	-0.14	0.08	-
AG32-3-63	48.73	1.51	15.79	10.40	0.17	7.84	11.45	2.60	0.21	93
+ 5.8% Olivine + 19% plagioclase										
+ 7.7% Clinopyroxene	49.12	1.07	16.66	9.00	0.15	8.44	11.92	2.37	0.15	63
Difference	0.13	-0.01	-0.10	-0.04	-0.01	-0.10	-0.22	-0.18	-0.02	-

Table 24: Summary of acceptable major element mixing calculations relating Group B basalts to the least evolved basalt (AG32-3-94) by the fractional crystallization of Olivine, Plagioclase and Clinopyroxene. Mineral analyses used in mix as per Table 22.

PARENT											
AG32-3-94	AG32-3-113		AG32-3-108		AG32-3-105		AG32-3-111		AG32-3-63		
	Obs	Pred	Obs	Pred	Obs	Pred	Obs	Pred	Obs	Pred	
Zr	63	74	73	77	76	88	87	93	91	94	92
Nb	1.60	2.70	1.9	2.6	2.0	3.0	2.2	3.4	2.4	2.7	2.4
Y	26	30	30	30	31	34	35	35	37	37	37
Rb	2.8	2.9	3.3	3.7	3.4	4.1	3.9	6.2	4.1	3.6	4.1
Ba	15	21	18	21	18	22	21	27	22	25	22
Sr	134	126	129	131	128	126	126	122	125	123	124
Co	53	51	53	50	50	51	50	49	49	49	51
Cr	366	327	332	314	368	283	320	265	315	271	302
Ni	188	151	154	149	148	131	124	125	121	128	129
V	215	241	244	244	256	273	281	272	295	275	290
Zn	69	80	77	77	78	84	86	84	89	85	90
Sc	40	40	41	40	44	41	43	40	44	42	43
		F = 0.851		F = 0.818		F = 0.716		F = 0.677		F = 0.670	
		$\frac{Ni}{D_{01}} = 12.94$		$\frac{Ni}{D_{01}} = 8$		$\frac{Ni}{D_{01}} = 10$		$\frac{Ni}{D_{01}} = 9$		$\frac{Ni}{D_{01}} = 9$	

Table 25: Calculated and predicted trace element concentrations for the Group B major element models presented in Table 24. Distribution coefficients as per Table 23, save for $\frac{Ni}{D_{01}}$ which are listed separately. See text for discussion.

that clinopyroxene is the high pressure (> 10 Kbar) liquidus phase in MORB (Bender et al, 1978; Bence et al, 1979), and from phase equilibria studies (Walker et al, 1979; Stolper, 1980; Elthon and Scarfe, 1984) (see Figure 26).

Another explanation of the pyroxene paradox is proposed by Francis (1986) in which the compositional variation is ascribed to picritic primary magmas crystallizing olivine along different fractionation paths (between limits of fractional and equilibrium olivine crystallization). The compositional trend marks the arrival of these different crystallization paths at the plagioclase saturation surface. This model is acceptable when there is a lack of a significant Ca variation in a suite of lavas (as Ca is not significantly affected by olivine fractionation) and thus cannot apply to the Group B lavas of this study (see Figure 24; Ca variation in Group B).

Evidence supporting clinopyroxene fractionation in the Group B basalts is found in the covariation of Ti/Zr ratio and an indicator of the degree of evolution (e.g. Mg#. See Figure 44). It is evident that the Group B basalts define a clear trend, while the Group A and Group C basalts do not. The trend shown by the Group B basalts results from the differing distribution coefficient of Ti (0.37 - 0.74 at basaltic liquidus temperatures; Irving, 1978) and Zr (0.03 - 0.22; McCallum and Charette, 1978) in clinopyroxene. Thus, fractionation of this ratio can occur during clinopyroxene crystallization, but it is not affected by olivine or plagioclase crystallization. The linear trend defined by Group B on Figure 44 is therefore indicative of clinopyroxene fractionation, while basalts in

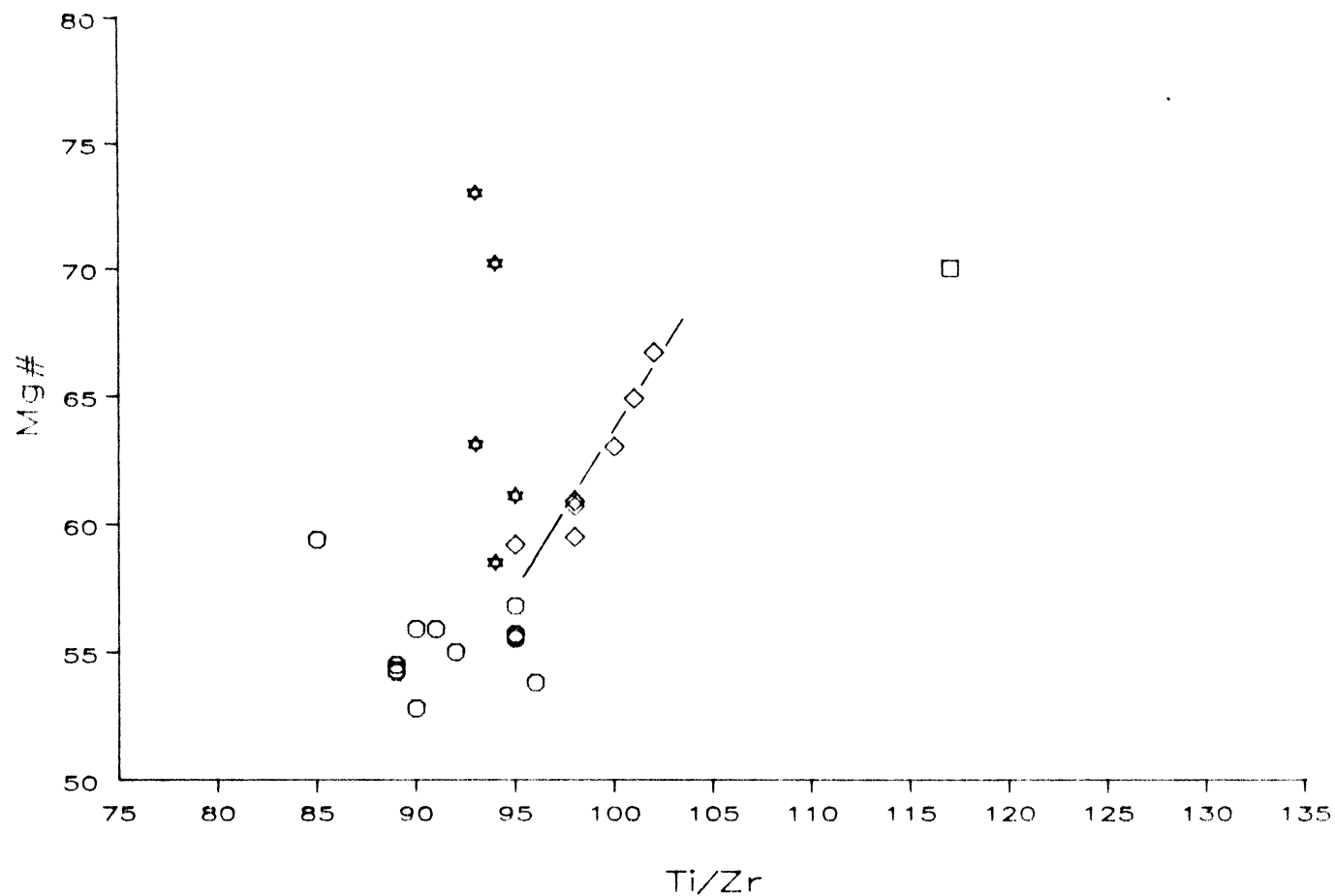


Figure 44. Ti/Zr vs Mg#. Basalt groups in which clinopyroxene is a fractionating phase (e.g. Group B) will show a clear decrease in Ti/Zr ratio with increasing degree of fractionation. Basalts in which clinopyroxene is not a fractionating phase (e.g. Group A) will show no correlation between Ti/Zr ratio and Mg .

which there has been no clinopyroxene fractionation (Group A and Group C) show no correlation between Mg# and Ti/Zr ratio.

A two stage fractionation model is therefore favoured for the evolution of the Group B samples. The first stage involves the high pressure crystallization of clinopyroxene (the Group B samples plot within the clinopyroxene stability field at $P \sim 7$ Kbar, see Figure 26) followed by magma ascent and concomitant low pressure fractionation of olivine and plagioclase during rise to the surface and eruption.

Petrogenesis of the Group C Basalts

The accumulative nature of this Group has been qualitatively shown in the petrography, geochemistry and mineralogy chapters. In this section, an attempt is made to rigorously model the accumulation trends defined by the Group C basalts.

The series is defined by the samples AG32-3-46 $\rightarrow$ AG32-3-30 $\rightarrow$ AG32-3-39 $\rightarrow$ AG32-3-36 $\rightarrow$ AG32-3-40. AG32-3-46 has the lowest Mg# (58.5) and the highest concentrations of incompatible trace elements (e.g. Zr = 135 ppm), while AG32-3-40 has the highest Mg# (73.0) and lowest trace element concentrations (e.g. Zr = 112 ppm). The accumulation of olivine will raise Mg# and decrease the incompatible trace element concentrations. The parent compositions were therefore selected as being the samples with lower Mg# and daughter compositions as the samples with higher Mg#, and assuming the parent compositions were

representative of liquid compositions. This is valid as a first approximation only, given the very sparsely olivine phenocrystic nature of the samples with lowest Mg# (< 3% olivine phenocrysts). Moreover those phenocrysts that are present are clearly not in equilibrium with whole rock compositions, having olivine - whole rock K_D 's of well below 0.27; the lower limit required to indicate equilibrium (Roeder and Emslie, 1970).

Least squares linear approximations for the sample pairs AG32-3-30 -> AG32-3-40; AG32-3-39 -> AG32-3-40 and AG32-3-36 -> AG32-3-40 are presented in Table 26. Sample AG32-3-30 has been selected as the "Group Parent" composition, because sample AG32-3-46 deviates slightly from the Group C evolutionary trend in compositional space (see Figures 30, 31, 34-36, 38-40).

From Table 26 it can be seen that major element compositional trends exhibited by the Group C lavas can be accounted for by up to 17.05% accumulation of olivine + Cr-spinel (olivine 98.21 % to 100 %; Cr-spinel 1.79 % to 0%). Observed trace element contents are compared to expected values, calculated from equation 4 and using the results from Table 26, (See Table 27) to provide a test of the acceptability of the major element model. The close agreement between observed and predicted values confirms the acceptability of the model.

Thus, from quantitative modelling of Group C compositional variation as an accumulation process, together with the porphyritic nature of these samples and mineralogical and geochemical evidence, it is concluded that the Group C MgO-rich

INPUT DATA

	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MnO	MgO	CaO	
Olivine: Fo 88	41.28	-	0.04	0.07	8.24	0.13	50.47	0.23	
: Fo 85	40.90	-	0.03	0.06	10.20	0.14	48.90	0.21	
Cr-Spinel: A	-	0.44	31.80	33.69	15.67	0.17	16.18	-	
: B	-	0.46	26.38	39.78	14.40	0.15	17.18	-	
Target: AG32-3-40	47.28	1.75	11.75	0.16	11.60	0.18	15.54	8.39	
AG32-3-30	48.95	2.08	14.08	0.09	11.87	0.20	9.21	10.04	$\sum r^2 = 0.01$
+16.74% Olivine + 0.31% Cr - Spinel	47.35	1.73	11.76	0.19	11.57	0.17	15.49	8.33	
Difference	.07	-.02	.01	.03	-.03	.01	-.05	-.06	
AG32-3-39	48.88	2.04	14.46	0.09	11.70	0.21	8.14	10.39	
+14.39% Olivine + 0.15% Cr - Spinel	47.41	1.75	11.88	0.13	11.55	0.16	15.44	8.42	
Difference	.13	-	.13	-.03	-.05	.02	-.1	.03	$\sum r^2 = 0.04$
AG32-3-36	47.78	1.86	12.53	0.15	11.80	0.19	13.66	8.91	
5.13% Olivine + 0% Cr - Spinel	47.38	1.77	11.89	0.14	11.56	0.18	15.46	8.45	
Difference	.10	.02	.14	-.02	-.64	-	-.08	.06	$\sum r^2 = 0.04$

Table 26: Acceptable major element least squares mixing calculations relating the Group C by Olivine and Cr - Spinel accumulation.

	Zr	Nb	Y	Rb	Ba	Sr	Co	Cr	Ni	V	Zn	Cu	Sc	
Parent (AG32-3-3)	135	7.4	46	5.1	51	114	54	648	241	347	106	52	40	Total Accumulation
Parent + 16.74% Olivine														17.05%
+ .31% Cr-Spinel	112	6.1	38	4.2	42	95	45	1338	610	288	88	43	33	
Target (AG32-3-40)	112	6.5	40	5.7	44	97	69	1075	653	250	98	45	32	
Parent (AG32-3-39)	131	7.8	45	4.8	50	112	59	647	313	343	105	51	40	Total Accumulation
Parent + 14.39% Olivine														14.54%
+ 0.15% Cr-Spinel	112	6.6	38	4.1	43	96	50	979	630	293	90	44	34	
Target AG32-3-40	112	6.5	40	5.7	44	97	69	1075	653	250	98	45	32	
Parent (AG32-3-36)	118	5.6	41	5.8	47	102	68	997	525	319	99	46	37	Total Accumulation
Parent + 100% Olivine														5.13%
+ 0% Cr-Spinel	112	5.3	39	5.5	45	97	65	997	638	363	94	44	35	
Target (AG32-3-40)	112	6.5	40	5.7	44	97	69	1075	653	250	98	45	32	

Table 27: Group C trace element model. All tr - elements calculated using the mass balance equation (4) (See text, after le Roex & Reid 1978) for accumulation processes.

picrite basalts represent lavas partially enriched in olivine and Cr-spinel and not primary liquid compositions.

Petrogenesis of the Group D Basalt

Group D comprises a single sample, AG32-3-5 and thus no quantitative modelling of the evolution of this group was undertaken.

The high Mg# (70.0 for the whole rock) and similar quench glass and whole rock composition (see Figure 25) together with low incompatible element concentrations suggest that this sample may represent a primitive high alumina basalt (see Chapter III). It is, however, also highly plagioclase phyric and thus may be a product of plagioclase accumulation into a fairly primitive magma. This will dilute the incompatible trace element concentrations without significantly altering Fe/Mg ratios or Mg#.

Evidence supporting the accumulation hypothesis comes from (1) the plagioclase phyric nature of the sample and (2) from a study of the plagioclase mineralogy. In chapter III, it was shown that only in plagioclase crystals from this basalt group could the Fe/Mg ratio be used to infer the Fe/Mg ratio of the magma at the time of plagioclase crystallization. The value of this ratio in the Group D plagioclases (1.87 ; $\sigma = 0.22$, $n = 17$) is inconsistent with them crystallizing in a magma with the Fe/Mg ratio of either the whole rock ($\text{Fe/Mg} = 1.12$, all Fe as

FeO) or the glass ($\text{Fe/Mg} = 1.13$) indicating that these plagioclase crystals are not in equilibrium with this magma and are thus xenocrystic or cumulate crystals.

The high Al and Ca concentrations found in the glass from Group D may be ascribed to the resorption of a quantity of plagioclase by the magma; a process that is petrographically evident. Thus, in light of the paucity of high alumina primary basalts in oceanic rifts and fracture zones (Dana Johnston, 1986), and adopting Occams Razor, a genesis for the Group D basalt by the accumulation of plagioclase into a primitive MORB is preferred.

Discussion

The petrogenetic studies of geographically and tectonically restricted suites of lavas is potentially useful in understanding the nature of the lavas source region, and in helping to understand the processes taking place in these regions. When such lavas can be related to more than one distinct parental magma, the effects of differing degrees of partial melting of the mantle source, and local mantle heterogeneities may be investigated. By thoroughly understanding small scale petrogenetic processes, and by comparison with samples from similar environments, an understanding of global variations and processes may be attained.

The parental magmas of Groups A, B and D were chosen as being the least evolved members of these groups, however none of these lavas can be regarded as representing primary magmas. A comparison of these parental magma compositions with some proposed primary magmas (Table 28) does however illustrate the primitive nature of the Group D sample AG32-3-5, (e.g. high Mg#, high MgO and low TiO_2 contents). As none of the parental magmas represent primary liquids, compositional variation in the basalts of this study can be ascribed primarily to fractional crystallization of the parental magmas which, in turn, evolved separately in space and time from more primitive magmas not yet sampled.

The nearly 40% fractionation experienced by the Group B basalts is consistent with that experienced by basalts at other fracture zones in the vicinity of the Bouvet triple junction (e.g. the Conrad and Bullard fracture zones on the American-Antarctic Ridge (le Roex & Dick, 1981; le Roex et al, 1985, 1987) and the Islas Orcadas and Shaka fracture zones on the Southwest-Indian Ridge (le Roex et al, 1983)). The substantial olivine and Cr-spinel accumulation experienced by the Group C lavas is however unusual for this vicinity. Indeed, highly olivine phyric ocean floor basalts are rare the world over (Francis, 1986; le Roex, Pers. Comm.).

Le Roex et al (1983, 1985, 1987) and le Roex and Dick (1981) have reported on the highly fractionated nature of many ocean floor basalts from the vicinity of the Bouvet triple junction. They ascribe this to either faulting exposing more fractionated

Source	4-107P	FAMOUS		CALCULATED PRIMARY MORB	GROUP A AG32-3-75	GROUP B AG32-3-94	GROUP D AG32-3-5
		P1	P2				
SiO ₂	49.13	48.78	50.19	47.85	49.44	48.30	48.64
TiO ₂	1.02	0.63	0.87	0.59	1.93	1.06	0.68
Al ₂ O ₃	15.80	15.77	15.33	12.08	14.62	16.52	19.35
Fe ₂ O ₃	1.06	-	-	-	1.47	1.18	0.96
FeO	7.41	7.96	8.36	8.98	9.92	7.85	6.42
MnO	0.26	0.15	0.15	0.12	0.18	0.16	0.13
MgO	11.00	12.32	10.72	17.81	7.01	8.81	8.40
CaO	11.98	12.61	11.84	11.22	10.86	11.97	13.17
Na ₂ O	2.09	1.55	1.88	1.31	2.84	2.51	2.10
K ₂ O	0.15	0.09	0.17	0.03	0.28	0.13	0.08
P ₂ O ₅	0.09	0.05	0.09	-	0.20	0.10	0.07
Zr	65	34	49	-	123	62	35
Nb	2.5	3.2	6.1	-	6.0	1.6	1.8
Y	27	17.3	19	-	44	2.6	17.5
Rb	-	1.9	3.4	-	5.3	2.8	ND
Co	58	-	-	-	49	52	44
Ni	340	396	270	-	95	185	136
V	210	202	198	-	331	212	156
Sc	33	42	35	-	41	40	32
Zr/Nb	26	11	8	-	21	38	19.6
Mg #	73	76	76	78	56	67	70

Table 28: Comparison of the parental magmas of Group A (AG32-3-75), Group B (AG32-3-94) and Group D (AG32-3-5) with some proposed primary magmas. 4-107P from Davey (1984), Famous P1 and P2 from le Roex (1980) and calculated primary Morb from Elthon (1979).

basalt than commonly found in moderately slow spreading ridges such as the MAR (le Roex and Dick, 1981), or to "leaky transform faults" in which new oceanic crust is injected into fracture zones in response to spreading normal to the fracture zone (Thompson and Melson, 1972), or to the cooler thermal regime that exists adjacent to fracture zones and the consequent low rate of magma supply (le Roex et al, 1982, 1983). A similar model has been proposed (Christie and Sinton, 1981; Sinton et al, 1983) explaining the presence of both highly fractionated and unfractionated basalts near propagating rift tips by cooling and crystallization in small, ephemeral crustal magma chambers. Such models are necessary when large degrees of fractionation are found at slow spreading ridges like the MAR, the AAR and the SWIR. Thermal models for the formation of oceanic crust (Sleep, 1975; Kuznir and Bott, 1976) suggest that the latent heat of basalts from the top 5 Km of oceanic crust at slow spreading ridges is insufficient to maintain a large, steady state magma chamber necessary for extensive fractionation. Small, ephemeral magma chambers are considered more likely to exist on slow spreading ridges (le Roex et al, 1981, 1983; Robson & Cann, 1982; Nisbett & Fowler, 1978), and are even more likely to exist in the cool thermal environment where a slow spreading ridge abuts against a fracture zone.

Christie and Sinton (1981) have shown that both highly fractionated and relatively unfractionated basalts may be derived from ephemeral magma chambers. The former requiring a delicate balance between rate of magma supply and cooling rate, while the latter will result when cooling rates are too high to permit extensive fractionation and the basalts are

erupted after short residence times in the crust. The presence of a number of ephemeral magma chambers in the vicinity of a fracture zone is conceptually acceptable. It is suggested that the basalts of this study evolved in such magma chambers; the Group A basalts by olivine and plagioclase fractionation in a thermodynamically open system; the Group B basalts by high degrees of olivine and plagioclase fractionation following earlier high pressure clinopyroxene fractionation, and the Group C basalts by small (<20%) amounts of olivine accumulation. The highly plagioclase phyric Group D basalt could also have formed within such a magma chamber. The four distinct parental magmas found in this 54°S fracture zone indicate that the mantle here is heterogeneous on a scale comparable to the volume of magma involved in the generation of individual batches of magma.

Le Roex et al (1983, 1985, 1987) have documented the geochemical enrichment of lavas from fracture zones and ridges around the Bouvet triple junction. For samples from the SWIR and AAR, the geochemically enriched signature (see Figure 43) is ascribed to the lateral influence of the enriched plume (P-type) magmas associated with the Bouvet hot spot, while either the recently proposed Shona hot spot (Hartnady and le Roex, 1985) or contamination of the mantle source with an ancient, isotopically distinct component have been postulated as being responsible for the geochemical enrichment of magmas from the southern MAR (le Roex et al, 1987). This geochemical enrichment is manifested in magmas from this region by high $^{87}\text{Sr}/^{86}\text{Sr}$, $^{207}\text{Pb}/^{204}\text{Pb}$, $^{208}\text{Pb}/^{204}\text{Pb}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios (le Roex et al, 1987), but not to any great extent in whole rock or glass major element geochemistry, save for some

indication of large ion lithophile (LIL) element enrichment (see Figure 43). There is no significant trace element evidence for the geochemical enrichment of the AG32-3 lavas, although on the basis of Zr/Nb ratios the Group A,C and D basalts are slightly less depleted than normal N-type MORB and the Group B basalts. The Group D basalt has a trace element profile (Figure 43) characteristic of depleted MORB, but a Zr/Nb ratio similar to the 'slightly less depleted than N-type MORB' Group A basalts. The 'depleted MORB' profile is thus interpreted as an artifact of the diluting effects of plagioclase accumulation on the absolute trace element concentrations in this basalt.. Le Roex et al (1987) favour a model explaining the geochemical enrichment of basalts from the southern MAR as being a result of source region mixing of low volume partial melts of enriched magma intruding normal depleted sub-oceanic mantle, analagous to models proposed for the SWIR and the AAR (le Roex et al, 1983, 1985) and for the North Atlantic (Wood, 1979). Within the framework of such a model, the AG32-3 basalts would represent the normal, depleted sub-oceanic mantle, uncontaminated by the enriched end-member.

In summary: The Group A lavas were derived from a mantle source slightly less depleted than normal MORB. The within group compositional variation is not explained by closed system fractional crystallization of olivine and plagioclase. An open system process in an ephemeral magma chamber is hypothesized. The Group B lavas are representative of normal depleted sub-oceanic mantle and were formed in a two-stage process involving firstly the high pressure fractionation of clinopyroxene, and secondly low pressure olivine and plagioclase fractionation.

The total degree of fractionation was of the order of 40% and this took place in ephemeral magma chambers in which the delicate balance between cooling rate and rate of magma supply, necessary for large degrees of fractionation (Christie and Sinton, 1981) had been achieved.

The Group C lavas represent olivine and Cr-spinel cumulates into an evolved magma similar to AG32-3-30 or AG32-3-46. The low degrees of accumulation (< 20%) suggest that this could have taken place in an ephemeral magma chamber, similar to those postulated for Groups A and B. The Group C lavas were derived from a similar, but not the same, primary magma as the Group A basalts. The Group D sample is also accumulative and has experienced ~20% (based on petrographic evidence) plagioclase accumulation into a primitive MORB magma.

The presence of four chemically distinct parental magmas in a single dredge haul attests to the heterogeneity of the mantle beneath the southern Mid-Atlantic Ridge on a local scale. The heterogeneities on a regional scale have been further dealt with by le Roex et al (1987).

Chapter Six

C O N C L U S I O N S

From the petrographic investigation of the AG32-3 basalts it can be concluded that most of these are similar to most MORB (Bryan et al, 1983). Unusual however is the presence of a suite of olivine phyric to highly olivine phyric picrite basalts. Such basalts are rare in the oceanic environment and may shed light on the origin of primary magmas.

Textural evidence and the presence of disequilibrium olivine phenocrysts in the Group C basalts however indicates that these basalts represent olivine cumulates, and as such do not represent 'primary magmas'. Like the Group C basalts, the highly plagioclase phyric Group D basalt can be shown to have an accumulative origin. Textural evidence and Fe/Mg ratios in plagioclase phenocrysts indicate that this basalt formed by the accumulation of about 20% plagioclase.

Four different parental magmas are identified on the basis of incompatible trace element ratios and Ba - Zr covariation, with each basalt group being derived from a geochemically distinct magma. Major element compositional variations within the Group A basalts may be explained by the addition of olivine (2.18%), plagioclase (5.83%) and clinopyroxene (2.88%) to a parent composition similar to AG32-3-75 to produce a daughter composition of AG32-3-97, however trace element constraints preclude simple closed system fractional crystallization as the process responsible for the scatter found in the composition of this basalt group. Indeed, the lack of any significant serial compositional variation in these basalts mitigates against any process other than a 'steady state' for the

evolution of the Group A basalts. Successful quantitative modelling of the compositional variation exhibited by the Group B basalts indicates that this group evolved by the polybaric, closed system fractionation of up to 6.58% olivine + up to 22.20% plagioclase + up to 9.42% clinopyroxene. It is hypothesized that this fractionation took place in small, ephemeral magma chambers.

The olivine phyric to highly olivine phyric Group C basalts have been successfully modelled as cumulate lavas formed by the addition of up to 16.74% olivine + up to 0.31% Cr-spinel to an evolved magma, represented by AG32-3-30. No quantitative modelling of the genesis of the highly plagioclase phyric Group D basalt was undertaken. Petrographic and mineralogical evidence is that this basalt formed by about 20% plagioclase accumulation into a primitive magma. The primitive nature of this basalt (whole-rock Mg# = 70.0; >9% MgO in the glass) is significant in that it shows that both highly evolved (e.g. the Group A) and primitive basalts may occur in a geographically restricted area.

Unlike many basalts from the SOLP study area, the source regions of the AG32-3 basalt groups show no evidence of geochemical enrichment and are, in fact, more characteristic of normal depleted mantle. Within the framework proposed by le Roex et al (1987) these basalts were derived from mantle uncontaminated by plume enriched partial melts. This effectively constrains the spheres of influence of the Bouvet hot spot to the south and the Shona hot spot to the northwest.

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Appendix One: Analytical procedure.

1: Whole Rock analysis.

All major elements with the exception of Na were analysed by X-ray fluorescence spectroscopy (XRF) using the lithium tetraborate method of Norrish and Hutton (1969). Na was analysed by using XRF, using pressed powder briquettes (Baird, 1961). Routine operating procedures in the Department of Geochemistry, U.C.T were adopted. They have been discussed in detail by Willis et al (1971, 1972). Estimates of precision and detection limits for the major oxides and the accuracy of the data (derived from the mean absolute error on the standardised calibration curves) are given in Table A-1.

Trace elements were analysed by XRF using pressed powder briquette (Baird, 1961) and the routine procedures in the Department of Geochemistry, U.C.T (Willis et al, 1971, 1972). Estimates of uncertainties based on counting statistics are presented in Table A-2.

2: Electron microprobe analysis.

Mineral and quench glass compositions were determined using a Cameca Camebax Microbeam Electron Microprobe interfaced to a Digital PDP 11/23 microcomputer in the Department of

Geochemistry, U.C.T. Operating conditions and standards used for mineral species and glass analysis are detailed in Tables A-3 to A-6. Standardization was undertaken at the beginning of each run. Reid (1977) discusses the precision of this technique, and this is further illustrated in Table A-7 in which the standard deviation of the mean of 10 replicate analyses of the same basaltic glass standard is compared with the uncertainty due to counting statistics. Data for all elements from basaltic glass standard VG-2, supplied by A.M. Reid.

The possibility that natural mineral standards are not homogeneous on a micron scale makes the absolute accuracy of the technique difficult to establish. However, Reid (1977) notes that good agreement is achieved when analysing one natural mineral standard against another, using abundance data obtained by another methods. When data are presented in this work, all quantities lower than the theoretical detection limits are expressed as "ND" (Not Detected), and as "-" where elements were not determined. Details of mineral standards used in the electron microprobe analysis of mineral phases and basaltic glasses are presented in Table A-8.

Oxide	%	Precision	Accuracy	LLD
SiO ₂	50.0	0.140	0.264	0.036
TiO ₂	1.0	0.008	0.008	0.005
Al ₂ O ₃	15.0	0.080	0.079	0.022
Fe ₂ O ₃	9.0	0.038	0.064	0.014
MnO	0.15	0.008	0.003	0.008
MgO	8.0	0.148	0.086	0.072
CaO	12.0	0.028	0.030	0.008
Na ₂ O	2.5	0.032	0.067	0.080
K ₂ O	0.2	0.002	0.022	0.002
P ₂ O ₅	0.2	0.012	0.018	0.01

Table A-1. Estimated precision and detection limits based on counting statistics, and average absolute error on the calibration curves for the major oxides (accuracy), using XRF techniques. Precision is expressed as an absolute error (2σ) on the given % oxide. From le Roex (1980).

Element	Concentration range (ppm)	2σ	LLD
Zr	40-400	0.94-1.4	1.4
Nb	4-150	1.0-1.2	1.8
Y	20-40	0.8-1.0	1.3
Rb	2.5-90	0.8-1.1	1.5
Ba	10-800	2.1-6.0	6.7
Sr	80-800	1.0-1.7	1.5
Co	18-100	1.9-2.5	4.5
Cr	25-350	1.5-2.1	2.2
Ni	10-208	2.5-3.7	4.0
V	50-350	2.4-3.7	5.5
Zn	8-150	0.9-1.7	1.4
Cu	5-90	1.0-1.2	2.9
Sc	25-40	1.0-1.2	1.2

Table A-2. Counting errors and detection limits for trace elements analysed by xrf. All data are in ppm. Lowest limits of detection (LLD) are given at 99% confidence levels, and 2σ counting errors are at 95% confidence level. From le Roex (1980).

Beam current: 40 nA

Accelerating voltage: 25 KeV

Data reduction: ZAF

Beam size: Fine, 2-5 μm .

Element	Line	Standard	Analysing	Counting	<u>PHA settings</u>			
			crystal	time (s)	v	g	bl	wdw
Si	K α	M-OL	TLAP	10	1420	60	80	340
Ti	K α	K-P	TLAP	10	1420	60	80	340
Cr	ka	CHRO	PET	50	1920	60	10	500
Fe	K α	M-OL	LIF	10	1920	60	30	300
Mn	K α	RHOD	LIF	10	1920	60	30	270
Mg	K α	M-OL	TLAP	10	1485	60	30	500
Ca	K α	DIOP	PET	10	1920	90	40	300
Ni	K α	NISI	LIF	50	1920	60	30	330

Table A-3. Electron microprobe operating conditions for the analysis of olivine. PHA settings; v = voltage, g = gain, bl = baseline, wdw = window.

Beam current: 20 nA

Accelerating voltage: 15KeV

Data reduction: ZAF

Beam size: Defocussed, 20 μm

Element	Line	Standard	Analysing	Counting	<u>PHA Settings</u>			
			crystal	time (s)	v	g	bl	wdw
Si	K α	NUNI	TLAP	10	1420	60	80	340
Al	K α	NUNI	TLAP	10	1420	60	80	340
Fe	K α	K-H	LIF	10	1920	60	30	300
Mg	K α	TLAP	TLAP	10	1485	60	30	500
Ca	K α	LACO	PET	10	1920	90	40	300
Na	K α	NUNI	TLAP	10	1485	60	20	400
K	K α	OR-1	PET	10	1920	90	30	270

Table A-4. Electron microprobe operating conditions for the analysis of plagioclase.

Beam current: 40 nA

Accelerating voltage: 15 KeV

Data reduction: Bence - Albee

Beam size: Fine, 2-5 μm

Element	Line	Standard	Analysing	Counting	<u>PHA settings</u>				
			crystal	time (s)	v	g	bl	wdw	
Si	K α	K-P	TLAP	10	1420	60	80	300	
Ti	K α	RUT	PET	10	1920	60	10	400	
Al	K α	CHRO	TLAP	10	1420	60	80	340	
Cr	K α	CHRO	LIF	10	1920	60	10	500	
Fe	K α	ILMT	LIF	10	1920	60	30	300	
Mn	K α	RHOD	LIF	10	1920	60	30	270	
Mg	K α	CHRO	TLAP	10	1485	60	30	500	
Ca	K α	K-P	PET	10	1920	90	40	300	

Table A-5. Electron microprobe operating conditions for the analysis of Cr-spinel.

Beam current: 20 nA

Accelerating voltage: 15 KeV

Data reduction: ZAF

Beam size: Defocussed, 30 μ m

Element	Line	Standard	Analysing	Counting	<u>PHA settings</u>			
			crystal	time (s)	v	g	bl	wdw
Si	K α	DIOP	TLAP	10	1420	60	80	340
Ti	K α	RUT	PET	10	1920	60	10	400
Al	K α	K-H	TLAP	10	1420	60	80	340
Cr	K α	CHRO	LIF	50	1920	60	10	500
Fe	K α	K-H	LIF	10	1920	60	30	300
Mn	K α	RHOD	LIF	10	1920	60	30	270
Mg	K α	K-H	TLAP	10	1485	60	30	500
Ca	K α	DIOP	PET	10	1920	90	40	300
Na	K α	K-H	TLAP	10	1485	60	20	400
K	K α	K-H	TLAP	10	1920	60	20	400

Table A-6. Electron microprobe operating conditions for the analysis of basaltic glass.

Oxide	Mean	Standard deviation	2σ	LLD
SiO ₂	50.31	0.21	0.34	0.05
TiO ₂	1.99	0.05	0.08	0.06
Al ₂ O ₃	14.33	0.11	0.18	0.04
FeO	12.18	0.17	0.39	0.12
MnO	0.20	0.02	0.07	0.09
MgO	6.73	0.13	0.13	0.03
CaO	10.93	0.08	0.18	0.04
Na ₂ O	2.58	0.07	0.10	0.04
K ₂ O	0.19	0.01	0.03	0.03

Table A-7. Estimates of the precision on major elements analysed by electron microprobe. All quantities in wt% oxides. Detection limits based on 10 second counting times.

STANDARD	NAME	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	NiO
M-OL	Marjalahti Olivine	40.24	-	-	-	11.53	-	48.08	-	-	-	-
K-P	Kakanui Pyrope	41.46	0.47	23.73	-	10.68	0.28	18.51	5.17	-	-	-
CHRO*	Synthetic Chromite	-	0.47	19.41	44.50	22.10	0.21	12.30	-	-	-	0.14
RHOD	Synthetic Rhodonite	46.70	0.03	0.17	-	3.73	40.76	0.56	7.82	0.05	-	-
DIOP	Synthetic Diopside	55.47	-	-	-	-	-	18.62	25.90	-	-	-
NISI	Synthetic Ni Silicate	28.69	-	-	-	-	-	-	-	-	-	71.32
NUNI	Nunivac Plagioclase	66.10	-	19.90	-	0.06	-	0.05	0.98	8.40	3.30	-
K-H**	Kakanui Hornblende	40.37	4.72	14.90	-	7.95	0.09	12.80	10.30	2.60	2.05	-
LACO	Lake County Plagioclase	51.55	-	30.99	-	0.41	-	-	13.24	3.63	0.13	-
OR - 1	Orthoclase, Val Switzerland	64.39	-	18.57	-	0.03	-	-	-	1.15	14.92	-
RUT	Synthetic Rutile	-	100.00	-	-	-	-	-	-	-	-	-
ILMT***	Synthetic Ilmenite	-	45.70	-	-	36.10	4.77	0.31	-	-	-	-

Table A-8: Composition of standards used for microprobe analysis of minerals

* CHRO : Contains 0.25% V₂O₅.

** K-H : Contains 3.30% Fe₂O₃, 7.95% FeO and 0.94% H₂O.

*** ILMT : Contains 11.60% Fe₂O₃, 0.92% Nb₂O₅.