ASPECTS OF THE GEOCHEMISTRY OF ONVERWACHT GROUP LAVAS FROM THE BARBERTON GREENSTONE BELT

by

Hudson Stuart Smith

Department of Geochemistry,
University of Cape Town.

April 1980

VOLUME 1

Text

Acknowledgements

References

Thesis submitted in fulfilment of the requirements for the degree of Doctor of Philosophy at the University of Cape Town.

ABSTRACT

The 3540 million year old komatiitic and tholeiitic lavas in the Onverwacht Group (Barberton greenstone belt) crop out in the rugged terrain of the eastern Transvaal lowveld. The results of an investigation into the geochemistry of the lavas, mainly from the three lower Formations of the Onverwacht Group - the Lower Ultramafic Unit (LUU) - are reported. While the lavas generally show excellent textural preservation, their primary phases have usually been reconstituted to greenschist facies mineral assemblages. Although original phenocryst phases are often pseudomorphed, they can still be identified from occasionally preserved relict grains and from the secondary mineral assemblages. In this way all the major phases that occur in the various lava types could be identified. However, before the geochemical data could be used to examine the effects of partial melting and/or crystal fractionation processes in the development of the magma compositions, it was necessary to investigate which elements had been redistributed by later metamorphic and other alteration events that have occurred in the history of the lavas.

The effects of alteration processes on the chemistry of the LUU volcanic rocks were investigated using a suite of komatiite pillow lava samples. Precise X-ray fluorescence analysis of 25 major and trace elements, demonstrates significant compositional variability for some elements, both within individual pillows and between pillows from the same flow. Observed variations of concentration have been considered in relation to the expected analytical error (±2 standard deviations). The elements: Si, Ti, Al, P, Nb, Zr, Y, Co, V and Sc are considered to have remained immobile in the pillow lavas as they do not show significant variations of concentration. Variations in the concentrations of the elements: Fe, Mn, Mg, Ca, Cr, Ni and Ga are small, but significant and can be accounted for either by the fractionation of phenocryst phases (e.g. Cr in chromite), or by the formation of secondary mineral phases (e.g. Ca, Fe and Ga in epidote). It is considered that these elements can be used for interpreting the igneous processes that have affected the lavas, provided samples selected for analysis show good preservation of their igneous textures, contain some relict igneous minerals and do not contain abundant amounts of secondary minerals such as epidote or chlorite. Large variations in the concentrations of the elements Na, K, S, Sr, Rb, Ba, Cu and Zn

are observed and the changes of concentration of some of the elements (e.g. S and Cu), by analogy with the alteration patterns observed in modern ocean floor pillow lavas can be attributed to sea water alteration processes. However, it is considered that the concentrations of this latter group of elements have been significantly modified by the alteration processes (sensu lato) that have affected the LUU lavas and consequently, they have not been used for interpretation and modelling of the data from komatiite and thoeliite lavas.

The effects of crystal fractionation processes in thin komatiite flows from the Komati Formation have been investigated in order to ascertain the importance of such processes in modifying lava compositions. Samples through sections of two flows have been analysed, the first, a 3.1 m thick spinifex textured ultramafic flow and the second, a 7.3 m thick mafic flow. Detailed modelling shows that the range of compositions occurring in the ultramafic flow (25-32% MgO) can be accounted for by the crystallisation and settling of olivine phenocrysts, before the flow solidified. Variations of composition in the mafic flow (12-19% MgO) can be accounted for by the crystallisation and accumulation of olivine and clinopyroxene. The fact that relatively thin lava flows were significantly affected by crystal fractionation processes before solidification, illustrates the importance of selecting phenocryst free samples for classification purposes, to ensure that geochemical parameters (e.g. CaO/Al_2O_3) have not been accentuated by the accumulation of phenocryst phases such as clinopyroxene.

Numerous classification schemes for komatiites are currently being employed in the literature. Evaluation of these schemes shows that there are many common parameters used for defining the rocks, such as general agreement that komatiites are a different suite of rocks from tholeiites, that textural criteria should also be included in the definition and that not all komatiites (including many samples from the Barberton greenstone belt) have CaO/Al_2O_3 ratios >1. However, there is very little agreement on precisely which geochemical parameters should be used to define komatiites. In this work a three fold division of the komatiites is proposed, which is simpler than the five fold division proposed originally by Viljoen and Viljoen (1969c). An attempt has been made to include the points of agreement listed above into the defining parameters. The three komatiite types are:-

a) Ultramafic Komatiites (>24% MgO) which incorporate the Komati Formation and Sandspruit Formation peridotitic komatiites of Viljoen and Viljoen (1969c,d).

b) High-Mg Mafic Komatiites (19-24% MgO) which are approximately equivalent to the Geluk type of basaltic komatiites of Viljoen and Viljoen (1969c).

c) Low-Mg Mafic Komatiites (8-16% MgO) which incorporate the Barberton and Badplaas types of basaltic komatiites of Viljoen and Viljoen (1969c).

Tholeiitic basalts are readily distinguished from komatiites in the LUU, as komatiites generally contain $\leq 12\%$ Al_2O_3 and ≥ 600 ppm Cr, while the tholeiites contain $\geq 13.5\%$ Al_2O_3 and ≤ 300 ppm Cr.

Two groups of ultramafic komatiites have been identified from the LUU. The group I lavas are rare and characterized by Al/Ti ratios of 20, while the group II lavas are common and characterized by Al/Ti ratios of 10. Certain major and trace element trends (e.g. CaO-MgO) displayed by the group II aphyric lavas (aphyric samples are considered to represent phenocryst free magma compositions) indicate that olivine differentiation, either by low pressure crystallisation or by melting in the mantle, could not have generated the range of ultramafic magma compositions (24-33% MgO). In contrast, the group I aphyric lavas (25-33% MgO) have compositional trends that are compatible with differentiation of olivine alone, either by melting processes in the mantle or by the crystallisation and settling of olivine in the magmas.

The group I and II porphyritic ultramafic komatiites (30-45% MgO) have textures, mineralogies and compositions consistent with their derivation from their respective magmas by olivine accumulation. Calculations of viscosities and phenocryst settling rates suggest that olivine phenocrysts could have settled through stationary ultramafic liquids with velocities in the order of m/hour. These rapid settling rates provide an adequate mechanism for obtaining the porphyritic rocks in the flows and sills by olivine accumulation.

Comparison of key geochemical trends developed in the LUU ultramafic komatiites with similar lavas from other greenstone belts (Belingwe, Western Australia and Munro Township) shows that both group I and II lavas

from the LUU are depleted in Al_2O_3 but have similar CaO contents to the lavas from these other areas. The CaO-MgO trend developed in the LUU group II lavas is similar to that in the Munro Township rocks, but has a steeper slope and lower MgO intercept value than the same trends developed in the Belingwe and Western Australia lavas. On the other hand the LUU group I lavas have a similar CaO-MgO trend to the Belingwe and Western Australia rocks. These data illustrate that aspects of the regional variations of ultramafic komatiite compositions can be duplicated in individual greenstone belts and suggests that the Archaean mantle was heterogeneous on a global and local scale.

The geochemical data of some of the porphyritic high-Mg mafic komatiites and all the aphyric low-Mg mafic komatiites, indicate that they could have been derived from the group II ultramafic magmas by the fractionation of olivine initially, and then by clinopyroxene and olivine, in magmas with $\leq$14% MgO. There are several problems in this model. Firstly, olivine occurs as a minor phenocryst phase in the mafic komatiites and not in the amounts that would be expected from the model. Clinopyroxene occurs as the major phenocryst phase in all the mafic komatiites. Secondly, no lavas with MgO contents in the range $\sim$16 to $\sim$19% MgO have been found and if this represents a real compositional break, it would eliminate the possibility that the low-Mg mafic komatiites were derived from the ultramafic magmas by processes of crystal fractionation. This compositional 'gap' may be due to inadequate sampling, although it has been noted in lavas from other greenstone belts. To account for these problems, a model of uncompensated and partially compensated crystal settling has been developed, and in principal could account for the observed range of mafic komatiite compositions and roughly their correct phenocryst proportions. It is also suggested, from consideration of the viscosities of the magmas, that porphyritic lavas in the compositional range from $\sim$14 to $\sim$24% MgO should be relatively scarce.

The porphyritic low-Mg mafic komatiites have compositions consistent with their derivation from the aphyric magmas by olivine and/or clinopyroxene accumulation. In order to account for the high CaO contents and CaO/Al_2O_3 ratios of some porphyritic rocks, they must have accumulated over 60% clinopyroxene.

It is tentatively suggested that olivine and augite were the main fractionating phases in the aphyric high-Mg mafic komatiites, although insufficient data have been obtained for this group of lavas to rigorously test petrogenetic models. Consideration of inter-element ratios indicates that the aphyric high-Mg mafic komatiites may have been derived from a similar source composition to the group I ultramafic komatiites.

Three distinct types of tholeiitic basalts have been identified in the Onverwacht Group volcanics:-

a) Low-Ti tholeiites that have similar inter-element ratios to the group I ultramafic komatiites.

b) High-Ti tholeiites that have similar inter-element ratios to the group II ultramafic komatiites.

c) High-Mg tholeiites that have similar inter-element ratios and absolute concentrations of elements (e.g. Si, Ti, Al, Mg, and Cr) to the rare boninite lavas from island arc tectonic environments.

The similarity of inter-element ratios in the low-Ti and high-Ti tholeiites with the group I and II ultramafic komatiites respectively, suggests that the tholeiitic lavas could have been derived from similar source compositions as their respective ultramafic counterparts. If this is correct it can be shown that the group I source has some inter-element ratios and Rare Earth element (REE) patterns, similar to chondrites and some lavas from the ocean floor (e.g. the FAMOUS area and the Scotia Sea Rise). The group II lavas have many inter-element ratios similar to some typical Mid-Ocean Ridge basalts, but the Onverwacht group II lavas have light REE enriched patterns. Similar inter-element ratios and REE patterns to the Onverwacht group II rocks are also found in lavas from the Scotia Sea Rise. From this it is inferred that the two source compositions that gave rise to the major proportion of Onverwacht Group lavas are still yielding lavas. The fact that lavas with similar geochemical characteristics to Archaean greenstone belt lavas, occur in close proximity in the Scotia Sea Rise, indicates that greenstone belts may have developed in a similar back-arc basin tectonic environment. However, this cannot be unambiguously proved with the available geochemical data at the present time.

Finally, high pressure processes that may have contributed to the generation of the Onverwacht Group lavas are considered in relation to high pressure phase boundaries projected from different mineral compositions in CMAS space. Some models that have been proposed in the literature for the generation of ultramafic komatiite magmas (such as advanced degrees of melting with olivine being the only phase left to melt in the residue during the formation of the ultramafic magmas) are consistent with the available data for the group I ultramafic komatiites. These models, however, do not account for the observed trends in the group II ultramafic komatiites and models involving mixing initial melt with overlying mantle (as outlined by Cox, 1978) are favoured. Specifically, a two stage melting and mixing process is discussed for the generation of the group II ultramafic komatiites. It is speculated that the group II tholeiites could have been generated during the first melting stage, provided garnet had been depleted in the residue. Neither of the two groups of tholeiites represent unmodified mantle derived melts, but they probably crystallised olivine and/or clinopyroxene before being extruded.

TABLE OF CONTENTS - VOLUME 1. Page No.

VOLUME 1 (CONTINUED) Page No.

VOLUME 1 (CONTINUED) Page No.

VOLUME 1 (CONTINUED) Page No.

VOLUME 1 (CONTINUED) Page No.

TABLE OF CONTENTS - VOLUME 2. Page No.

VOLUME 2 (CONTINUED) Page No.

TABLES

VOLUME 2 (CONTINUED) Page No.

VOLUME 2 (CONTINUED) Page No.

VOLUME 2 (CONTINUED) Page No.

VOLUME 2 (CONTINUED) Page No.

FIGURES

VOLUME 2 (CONTINUED) Page No.

VOLUME 2 (CONTINUED) Page No.

VOLUME 2 (CONTINUED) Page No.

VOLUME 2 (CONTINUED) Page No.

CHAPTER 1.

THE GENERAL GEOLOGY AND GEOCHEMISTRY OF ARCHAEAN GREENSTONE BELTS

1-1 INTRODUCTION

The Barberton Mountain Land occurring in the eastern Transvaal lowveld (South Africa) incorporates most of the geology and outcrop of the Swaziland Sequence commonly known as the Barberton greenstone belt. This area has contributed significantly to our understanding of the geology and geochemistry of Archaean granite-greenstone belts as the volcanic and granitic rocks (sensu lato) are well exposed and relatively well preserved. From the detailed field and geochemical evidence that has accumulated, the Barberton greenstone belt has been used as a model for the development and evolution of the early crust (Anhaeusser et al., 1968, 1969; Viljoen and Viljoen, 1969i; Anhaeusser, 1971a,b). Much of the interest displayed in these belts can be ascribed to their economic potential (Viljoen et al., 1970; Viljoen and Viljoen, 1969h; Anhaeusser, 1972, 1976a; Watson, 1976), as well as for obtaining information on the composition and evolution of the early crust and mantle (e.g.Hart et al., 1970; Glikson, 1971; White et al., 1971; Anhaeusser, 1973 ; Nesbitt and Sun, 1976). Part of the recent interest has been stimulated by the discovery and characterisation of high magnesium mafic and ultramafic volcanic rocks termed komatiites (Viljoen and Viljoen, 1969c,d), from the lower Formations of the Swaziland Sequence. Although non-cumulate sub-alkaline rocks richer in MgO than basalts had previously been recorded (Drever and Johnson, 1957, 1966; Gass, 1958; Drever et al., 1961) the discovery of the high magnesium lavas from the Archaean greenstone belts appears to have had much more impact for stimulating research on ultramafic lavas. High magnesium

mafic and ultramafic lavas have now been recorded from many other greenstone belts (e.g. from South Africa, Rhodesia, Canada, Australia, India and Russia) and rocks corresponding to mafic komatiites in composition have been found in younger tectonic environments, although so far ultramafic komatiite lavas appear to be confined to the Archaean greenstone belts (Brooks and Hart, 1974).

This thesis is predominantly concerned with the petrography and geochemistry of the ultramafic and mafic lavas from the lower Formations of the Swaziland Sequence outcropping in the Barberton Mountain Land. This area was chosen for study as it has recently been mapped in detail by Viljoen and Viljoen (1969c,e), the rocks are well exposed, relatively well preserved, and accessibility is good.

In the following sections of this Chapter the present status of the Swaziland Sequence, along with other well documented Archaean greenstone belts is briefly reviewed. The ages of these greenstone belts and their relationships to the surrounding granites are also discussed. The aims, objectives, the methods employed for this work and the layout of this thesis are discussed in the final section.

1-2 GEOLOGY OF THE BARBERTON GREENSTONE BELT

The Barberton greenstone belt occurs on the Kaapvaal craton (as defined by Anhaeusser et al., 1969) and has been strongly folded and subjected to regional greenschist facies metamorphism. Locally a thin aureole of amphibolite or granulite facies metamorphism is developed within the volcanic rocks along the boundaries with granite intrusions, (Viljoen and Viljoen, 1969a). The volcanic and sedimentary rocks of the Barberton greenstone belt are now referred to as the Swaziland Sequence, which is subdivided into three Groups (Viljoen and Viljoen,

1969a). These are the lower predominantly volcanic Onverwacht Group, the middle predominantly argillaceous unit containing ferruginous chert and tuffacious material, forming the Fig Tree Group, and the upper, mainly arenaceous Moodies Group. Anhaeusser (1976b) has recently compiled a bibliography of publications pertaining to the geology, geochemistry and mineralization of the Barberton greenstone belt and surrounding granitic terrain. Some of the relevant references for this work are discussed in the following sections.

ONVERWACHT GROUP

Descriptions of the volcanics from the Onverwacht Group are readily available in the literature (e.g. Anhaeusser et al., 1968, 1969; Viljoen and Viljoen, 1969a,b,c,d,e, 1970a; Anhaeusser, 1971a,b; Viljoen and Viljoen, 1971; Allsopp et al., 1973; Anhaeusser, 1973, 1976a; Williams and Furnell, 1979), and these references should be consulted for more detailed accounts of the geology of this area than will be given here. The Onverwacht volcanic rocks attain a thickness of over 15 km. in the southern portion of the Barberton greenstone belt, and have been subdivided into six Formations on the basis of distinctive lithologies (Viljoen and Viljoen, 1969a,c,d,e). Starting with the oldest, these Formations are called the Sandspruit, Theespruit, Komati, Hooggenoeg, Kromberg and Swartkoppie. The three lower Formations are collectively termed the Lower Ultramafic Unit (LUU) and the three upper Formations are referred to as the Upper Mafic-Felsic Unit (UMFU). Anhaeusser (1975) has re-named these two units as the Tjakastad and Geluk Sub-Groups respectively, but for the purposes of this thesis the original terms will be retained.

The Lower Ultramafic Unit consists predominantly of ultramafic and mafic komatiite rocks often distinguishable as pillows and lava

flows. Well preserved igneous textures have played an important role in understanding the primary nature of the ultramafic lavas. Tholeiitic basalts, while present, are relatively rare, and acidic volcanic material occurs as felsic tuffs in the Theespruit Formation and very rarely as pillows in the Komati Formation. Sedimentary rocks show only minor development in the Sandspruit Formation, while cherty horizons, often closely associated with the felsic tuffs, are common in the Theespruit Formation. The Komati Formation is capped by a persistant sedimentary horizon termed the Middle Marker. This horizon is believed to represent a time break between volcanism forming the Lower Ultramafic Unit and the onset of volcanism of the Upper Mafic-Felsic Unit (Viljoen and Viljoen, 1969a).

The Sandspruit Formation occurs mainly as xenolithic blocks detached from the Theespruit Formation by enveloping tonalitic granite and an unknown amount of the stratiographic section may have been eliminated from the base of the Sandspruit Formation. The upper two Formations of the lower Ultramafic Unit crop outaround the whole belt, but attain their best development in the southern and south-eastern areas, where the type sections for these Formations have been established (Viljoen and Viljoen, 1969a). The Jamestown schist belt extends out on the north west flank of the Barberton greenstone belt and has been described in detail by Anhaeusser (1972). This belt which consists mainly of mafic and ultramafic rocks has been correlated with the Theespruit and Komati Formations. The rocks have been extensively deformed, metamorphosed and generally are not as well preserved as the lavas from the type areas of the above Formations. As the scope of this project has mainly been restricted to the Lower Ultramafic Unit, sampling was predominantly carried out in the type areas of the three lower Formations.

The Formations of the Upper Mafic-Felsic Unit crop out above the Middle Marker, and are characterised by their different volcanic rock assemblages compared to the Lower Ultramafic Unit. The Hooggenoeg Formation consists predominantly of tholeiitic lavas with lesser amounts of andesite, dacites and rhyodacites. An outstanding feature of this Formation is the cyclic nature of the volcanism (Viljoen and Viljoen, 1969a,e; Anhaeusser, 1971a), grading from tholeiitic lavas at the base through to dacite-rhyodacite lavas at the top. Each cycle is often terminated by a chert horizon. The basaltic component of each cycle generally becomes progressively thinner and the chert component thicker, from the lowest to the uppermost cycle in the Hooggenoeg Formation. The basaltic base of many of the cycles also contain sill-like sheets of ultramafic material, which have differentiated into a lower peridotite zone and an upper pyroxenitic zone. The top of the Hooggenoeg Formation is marked by a thick (~150 m.) felsic volcanic and pyroclastic horizon capped by chert (Viljoen and Viljoen, 1969e).

The Kromberg Formation conformably overlies the Hooggenoeg Formation and only shows significant development in the southern area of the belt. It consists essentially of the same rock types as occur in the Hooggenoeg Formation but with less well developed cyclic nature. The Swartkoppie Formation (also spelt Zwartkoppie) occurs at the top of the Onverwacht Group and consists of basic, intermediate and acidic volcanics, now altered to schists. Cherts, pyroclastics and greywackies also occur along with serpentinized ultramafic bands and lenses. Two of these ultramafic pods are mined for asbestos at the Havelock mine in Swaziland and the Msauli mine in the Transvaal (Viljoen and Viljoen, 1969e,h). Ultramafic rocks and material of basaltic komatiite

composition, while much less abundant than in the three lowest formations, do occur in the Upper Mafic-Felsic Unit. As already mentioned the scope of this work has mainly been limited to the Lower Ultramafic Unit, however some samples of the lavas from the Upper Mafic-Felsic Unit have been analysed and will be discussed in the relevant Chapters.

FIG TREE AND MOODIES GROUPS

The Fig Tree Group conformably overlies the Onverwacht volcanics and has been subdivided into three Formations in the Stolzburg syncline area (Riemer, 1967, in Viljoen and Viljoen, 1969a). The basal Sheba Formation consists of greywackes, shales and narrow bands of chert and ferruginous chert. The greywackes show graded bedding, flute casts and groove casts indicative of turbidite deposition (Kuenen, 1963). At the base of this formation, a hard siliceous chert-like rock is developed known locally as the Consort Bar, along which much of the gold mineralization has been located in the Sheba hills area (Anhaeusser, 1972). The overlying Belvue Road Formation consists of cherts, sandy shales, trachytic tuffs and fine grained shales. The upper most Schoongezicht Formation consists of tuffs, agglomerates and trachytic lavas. Anhaeusser (1976c) has compiled a detailed map of the Fig Tree and Moodies Groups in the mineralized area around the Eureka Syncline.

Danchin (1967) has shown that the shales from the Fig Tree Group are enriched in Cr and Ni and suggests an ultramafic rich source area for these sediments. Condie et al. (1970) have noted a similar enrichment in the greywackes from the Sheba Formation and have on the basis of texture and chemistry suggested a mixed source area of chert, volcanics, granitic and metamorphic rocks.

The Moodies Group is well developed in the Eureka Syncline where it has been divided into the Clutha, Joe's Luck and Baviaanskop Formations (Anhaeusser, 1976c). These Formations consist of repeated cycles of arenacious rocks that alternate with argillaceous sediments. A basaltic lava horizon and several jaspilitic iron-formations also occur. The sedimentary structures together with conglomerate horizons has led Anhaeusser (1976c) to suggest a high energy, shallow water depositional environment for most of the Moodies succession. Eriksson (1978) considers that the Moodies Group sediments accumulated in diverse alluvial and marginal marine depositional environments and that the mineralogy of the fluvial sediments is indicative of an anoxygenic atmosphere at the time of their deposition. The chert horizons developed sporadically through most of the stratigraphy of the Swaziland Sequence have assumed great importance, as they contain traces of complex organic molecules and early life forms (e.g. Barghoorn and Schopf, 1966; Muir and Grant, 1976).

INTRUSIVE ULTRAMAFIC-MAFIC ROCKS

Numerous intrusive differentiated ultramafic complexes have been emplaced into the Onverwacht Group, and are particularly abundant within the Komati Formation (Viljoen and Viljoen, 1969h, 1970b; Anhaeusser, 1975). The magmas forming these bodies are believed to have been intruded penecontemporaneously with the mafic and ultramafic lavas. Viljoen and Viljoen (1969a,h, 1970b) hava classified these bodies into different types on the basis of their differentiation products. Generally they consist of cyclic repetitions of dunite, harzburgite, peridotite, pyroxenite, gabbro, norite and

anorthosite (Anhaeusser, 1975). Monomineralic cumulates of olivine or orthopyroxene invariably form layers of dunite or orthopyroxenite respectively at the base of the bodies. Although the igneous mineralogy is usually extensively altered, localised areas occur within many of the bodies where fresh samples can be obtained. Viljoen and Viljoen (1970b) have computed bulk compositions of the Ship Hill and Koedoe bodies. The presumed parental magma, as represented by these bulk compositions, is similar to the average composition of the ultramafic extrusives from the Komati Formation. These bodies, both in the Lower Ultramafic Unit and the Upper Mafic-Felsic Unit, contain economic deposits of chrysotile asbestos (Viljoen and Viljoen, 1969h; Anhaeusser, 1976d).

1-3 GRANITIC ROCKS AND ISOTOPIC AGES FROM THE BARBERTON AREA

Summaries of the ages obtained (using various isotopic systems) from the rocks of the Barberton Mountain Land can be found in Allsopp et al. (1973), Anhaeusser (1973), Jahn and Shih (1974) and Davies and Allsopp (1976). Viljoen and Viljoen (1969f,g) have proposed a broad four fold classification of the granitic rocks from this area and although dated, provides a useful framework within which to discuss the ages of the granitic rocks. The oldest 'granites' in the immediate vicinity of the Barberton greenstone belt are the Ancient Tonalitic Gneisses represented by five plutons in the south western zone of the belt, including the Theespruit pluton (dated at 3432±135 Ma., Barton, pers. comm.) and the Helshoogte pluton (dated at 3180±75 Ma., Barton, pers. comm.). The contact relations of these plutons with the belt are intrusive, forming a type of lit-par-lit injection into the volcanic rocks. Recent detailed mapping (Anhaeusser and

Robb, 1978) of portions of these plutons has shown that there may be several intrusive cycles within a single pluton.

The Nelspruit migmatites occurring to the north and south west of the belt are the next oldest 'granitic' rocks (2992-3170 Ma., see Anhaeusser, 1973). Robb (1977) has recognized six distinct granite types in this complex terrain and one type, the Nelspruit Porphyritic granite has been dated by the Rb-Sr method at 3205±49 Ma. (Barton, pers. comm. in McCarthy and Robb, 1977).

The third oldest granitic rocks in the Viljoen and Viljoen (1969f) classification is the homogeneous Hood granite cropping out over most of the area to the south east of the belt. It is described as a potassium rich granite and has been dated at 3070-3075 Ma. by both Rb-Sr and U-Pb methods (Allsopp et al., 1962; Oosthuyzen, 1970, in Anhaeusser, 1973). The youngest granitic event in this area is reflected by a number of coarse grained plutons that have clearly defined intrusive contacts with the country rock. Initial $^{87}Sr/^{86}Sr$ ratios of some of these plutons are high (de Gasparis, 1967, in Anhaeusser, 1973; Robb, 1977), which suggests that some of the plutons have their source areas in much older crustal material. It should be noted however that J.M. Barton of the Bernard Price Institute of Geophysical Research has recently obtained many more isochron ages from the different granitic rocks surrounding the Barberton greenstone belt. Detailed interpretations of these ages have yet to be published.

Attempts to date the rocks of the Swaziland Sequence directly, using both sedimentary and igneous rocks, have been made. Allsopp et al. (1968) obtained an Rb-Sr isochron age of 2980±20 Ma. from the Fig Tree shales, which they suggest reflects a metamorphic

overprinting by the intrusive granites. Hurley et al. (1972) have dated the sediments from the Middle Marker at 3350±70 Ma. Ulrych et al. (1967) have obtained a common Pb age of 3460 Ma. from sulphides in mines from the Fig Tree and Moodies Groups while Saager and Koppel (1976) have recently obtained an essentially identical common Pb age of 3450 Ma. also using sulphides from the Barberton area. They found two stage leads which indicates that differentiation of the U-Pb system had already occurred by this time.

The volcanic rocks from the Onverwacht Group have yielded ages of 3360 Ma.by the common Pb method from acid volcanics (van Niekerk and Burger, 1969), 3230 m.y. (common Pb method), 3290 Ma. (U-Pb concordia) from basic volcanics (Sinha, 1972) and 2629±20 Ma. (Rb-Sr) from felsic volcanics from the Hooggenoeg Formation (Allsopp et al., 1973). The Rb-Sr age is clearly anomalous and Allsopp et al. (1973) have attributed this younger age to later hydrothermal alteration. In fact, open system behaviour of Rb-Sr has been shown to be a real problem with respect to using this method for dating the volcanic rocks (Allsopp et al., 1973; Jahn and Shih, 1974).

Jahn and Shih (1974) have obtained a density separate internal isochron age from a basaltic sample from the Komati Formation, of 3500±200 Ma. They interpret this age as the time of the low grade metamorphism and the initial $^{87}Sr/^{86}Sr$ ratio of 0.70048±0.00005 to characterise the upper mantle at this time. Recently, however, a precise age has been established for the Onverwacht Group volcanics of 3540±30 Ma. by Hamilton et al. (1979) using the Sm-Nd isotopic method.

Hunter (1973, 1974) has introduced the term Ancient Gneiss Complex for a suite of rocks that include metasediments and

serpentinites which are associated with tonalitic gneisses and migmatites in Swaziland. Viljoen and Viljoen (1969f) consider the Ancient Gneiss Complex to belong to their group of Ancient Tonalitic Gneisses and therefore to be younger than the Swaziland Sequence. Hunter (1973) on the other hand has noted a set of deformed dykes within the Ancient Gneiss Complex that does not occur in the Swaziland Sequence, and has concluded from this that the Ancient Gneiss Complex predates the Barberton greenstone belt.

Barton (pers. comm.) has obtained an Rb-Sr isotopic age of 3555±100 Ma. from the Ancient Gneiss Complex which is within error of the Sm-Nd age for the Onverwacht volcanics of 3540±30 Ma. (Hamilton et al., 1979). Although the relationships between the granites and greenstone belt in this area remain controversial, evidence for a granitic basement for greenstone belt volcanics from other areas has been found. Bickle et al. (1975) for example have mapped a basalt unconformity between the Belingwe greenstone belt and the underlying granite, although the age of this belt is much younger than the age of the Swaziland sequence rocks. Gneissic rocks which are several hundred million years older than the Onverwacht lavas have been identified in other parts of the world (e.g. Amîtsoq Gneisses, West Greenland, Moorbath et al., 1972, Sand River Gneisses,Limpopo Mobile belt, Barton and Ryan, 1977) while van Niekerk and Burger (1975) have dated zircons and sulphides from a granitic conglomerate boulder from the Moodies Group at close to 4200 Ma. If this age for the granitic boulder is confirmed it would constitute evidence for a pre-existing granitic floor for the Swaziland Sequence rocks.

1-4 GEOCHEMISTRY OF THE ONVERWACHT GROUP LAVAS

As this topic will be discussed in depth in the following Chapters, only a brief summary will be given here. The rocks from the Onverwacht Group have been subject to at least greenschist facies metamorphism and most of the primary mineralogy has been altered, consequently the geochemistry of the lavas has played a major role in their classification. Viljoen and Viljoen (1969c,d,e) considering the unusual major element chemistry of the lavas, proposed a new class of igneous rocks, the Komatiites, with four subclasses. The peridotitic komatiites have the highest MgO content, the Geluk type, Badplaas type and the Barberton type basaltic komatiites have progressively lower MgO contents. The least magnesian Barberton type basaltic komatiite was not rigorously defined due to insufficient data (Viljoen and Viljoen, 1969c). The important diagnostic parameters are the high MgO, low alkali contents and the high CaO/Al_2O_3 ratios. Equally as important is the field evidence, particularly for the peridotitic komatiites, which indicate that the rocks represent magma compositions. Evidence such as pillows, olivine spinifex textured horizons, chilled margins and stratographic thinning was advanced by Viljoen and Viljoen (1969c,d) in support of magmas having existed with the same (or nearly so) composition as the komatiites. These lavas are best developed in the Lower Ultramafic Unit. Tholeiitic lavas also occur, but are more abundant in the Upper Mafic-Felsic Unit.

Relatively little chemical data are available on the composition of the felsic volcanics developed in the Upper Mafic-Felsic Unit. This is mostly due to the poor preservation of these lavas, but what little data there are, show a range in composition from andesite

to rhyodacite. Viljoen and Viljoen (1969e) have proposed a calc-alkaline trend of magma evolution for these rocks.

The important chemical distinction has been made between the predominantly komatiitic lavas below the Middle Marker and the predominantly tholeiitic lavas, with lesser amounts of felsic extrusives above the middle marker (Viljoen and Viljoen, 1969e). The Viljoens have suggested that this fundamental change is probably related to thickening of the sialic crust during the break in volcanism in which the Middle Marker was deposited. The genesis and inter-relationships between the different magma types will be discussed in the following Chapters.

1-5 GEOLOGY AND GEOCHEMISTRY OF OTHER ARCHAEAN GREENSTONE BELTS

Within South Africa detailed data have recently become available on two greenstone belts, the Roodekrans Ultramafic Complex and surrounding Archaean volcanic rocks (Anhaeusser, 1977) and the Murchison Range (Minnett, 1975; Saager and Koppel, 1976). These belts show many geological and geochemical similarities to the Barberton greenstone belt. Numerous other Archaean greenstone belts are documented in the literature; for the purposes of this brief description only those occurring in Rhodesia, Western Australia and Canada will be discussed.

RHODESIA

Recent reviews on the evolution of the Rhodesian Craton have been published by Bliss and Stidolph (1969), Stowe (1971), Wilson (1973), Moorbath (1977) and Wilson et al. (1978).

The greenstone belts have been subdivided into three groups:-

(a) The Sebakwian Group consists of metamorphosed remnants of ultramafic and mafic rocks and sediments. The geological relationship of this group to the granitic rocks and the overlying Bulawayan Group is not always clear. A minimum age has been obtained for the Selukwe schist belt which is geologically older than the Mont d'Or granite dated at 3420±60 m.y. (Moorbath et al., 1976). Greenstone relics occur within some of the oldest gneisses dated at 3500-3600 m.y. (Hawkesworth and Bickle, 1976).

(b) The Bulawayan Group greenstone belts have been divided into two groups (Wilson et al., 1978): the lower greenstones, consisting of an ultramafic/mafic-felsic bimodal volcanism and the upper greenstones have a range of lava compositions from high-magnesium to calc-alkaline in nature, (andesitic rocks tending to be more common higher in the sequence, Wilson, 1973). The Bulawayan Group greenstone belts have been dated directly and give ages of 2600-2700 m.y. (Hawkesworth et al., 1975) 2730-3080 m.y. (Jahn and Condie, 1976) and 2640±140 Ma. (Hamilton et al., 1977).

(c) The overlying Shamvian Group consists predominantly of sedimentary material, but also contains some basaltic, andesitic and felsic volcanics.

Detailed geochemical data are available for some greenstone belts in Rhodesia (Bickle et al., 1975, 1976, 1977; Condie and Harrison, 1976; Hawkesworth and O'Nions, 1977; Nisbet et al., 1977). Well preserved peridotite, high-magnesium and tholeiitic lavas have been analysed. The peridotitic lavas and high-magnesium basalts have been classified as komatiites as they show many of the petrographic and chemical characteristics of the komatiites from the Barberton greenstone belt, although they often lack the high CaO/Al_2O_3 ratio. Data from the Belingwe greenstone belt (2640 Ma., Hamilton et al., 1977) in particular are used for detailed comparison with the Barberton rocks analysed in this work. More detailed discussions will be incorporated in later sections of this thesis.

WESTERN AUSTRALIA

The granite greenstone terrain in the Yilgarn block from Western Australia shows many similar features to the Kaapvaal and Rhodesian cratons, the Coolgardie-Kalgoorlie-Norseman area being particularly well documented (Glikson and Lambert, 1976). At least two volcanic-sedimentary mega-cycles are developed in this area, termed the lower greenstones and upper greenstones respectively. The upper greenstones have been shown to unconformably overlie granitic rocks (Durney, 1972) in places, while the lower greenstones are intruded by granodioritic-tonalitic plutons. No unconformable contacts between the lower greenstone and granites have yet been observed. The oldest granitic gneisses so far recorded from this area are 3100 Ma. old (Arriens, 1971) but the major emplacement of granites and gneisses occurred between 2700 and 2600 Ma. (Arriens, 1971). A younger age limit has been set on the upper greenstones

(2675 Ma. Turek and Compston,1971) by considering the regional metamorphic patterns. Hallberg et al. (1976a) have determined the age of the Marda volcanics at 2635±80 Ma. by the Rb-Sr whole rock method. No lower age limits have yet been placed on the older lower greenstones (Turek and Compston, 1971; Glikson and Lambert, 1976).

Recent geochemical data have been obtained on the ultramafic-mafic rocks from the greenstones (2700 Ma.) within the Yilgarn block by Nesbitt (1971), Hallberg (1972), Hallberg and Williams (1972), Williams (1972), McCall (1973), Williams and Hallberg (1973), Hallberg et al. (1976a,b), Nesbitt and Sun (1976) and Naldrett and Turner (1977). Ultramafic and high magnesium lavas are present as well as tholeiitic and felsic rocks. Detailed comparison of these chemical data from the ultramafic and high-Mg basaltic rocks with the komatiites from the Barberton show many similar features, although the Western Australian lavas generally lack the high CaO/Al_2O_3 ratio (Nesbitt, 1971; Williams, 1972; McCall, 1973).

Many of the Australian geologists have objected to the Viljoen and Viljoen (1969c) classification scheme for komatiites. The basis of their objection is that a very limited number of analyses were used for the original definition; more significance was attached to the chemical rather than textural parameters (Williams, 1972), and that the definition was too specific, as each area has its own characteristics. The definition of komatiites will be examined in greater detail later in this thesis.

CANADA

Vast areas of Precambrian rocks are well exposed on the Canadian shield. Greenstone belts from 160-800 kms. long and 16-160 kms.

wide are present in six of the seven structural provinces of the Canadian shield (Baragar and McGlynn, 1976). Recent summaries and syntheses of the geology have been published by Goodwin (1971), Goodwin (1972) and Baragar and McGlynn (1976).

The greenstone belts of the Superior province are believed to have formed in a 200 m.y. interval, from 2750 to 2950 Ma. (Krogh and Davis, 1972), and the Yellow Knife volcanic rocks have been dated at 2625±160 Ma. (Green et al., 1968). Nunes and Thurston (1978) dating zircons from acid volcanics, have shown that a greenstone belt in the Uchi Lake area (N.W. Ontario) evolved over a 220 m.y. period from 2959±3 to 2738±5 Ma. Zindler et al. (1978) have obtained an age using the Sm-Nd isotopic system of 2765±47 Ma. for Abitibi greenstone belt in the Munro Township area. Recently evidence has become available for the existence of very ancient granitic rocks within the Canadian shield. Barton (1975) has dated the Hebron Gneiss in Labrador at 3620±60 Ma. (see summary by Moorbath, 1977). Gneisses and granites from the nearby Minnesota River Valley have yielded ages of 3630±60 Ma. (Goldich and Hedge, 1975). Baragar and McGlynn (1976) have reviewed the evidence for six widely scattered unconformaties between greenstone belts and underlying granitoid rocks. The evidence suggests that, at least, in places granitic crust preceded the greenstone belt volcanism.

The main components of the Superior province greenstone belts are flows and pyroclastics of basalt, andesite, dacite and rhyolite compositions, often arranged in mafic to felsic cycles. Intercalated with the volcanics, especially in the upper sections of the belts, are greywackes, argillites, tuffs, conglomerates and banded iron formations. Geochemical data are available on the volcanic

rock sequences (Baragar, 1966, 1968; Baragar and Goodwin, 1969) and recently numerous publications have appeared on the composition of ultramafic and high magnesium rocks particularly from the Abitibi greenstone belt (Naldrett and Mason, 1968; Pyke et al., 1973; Pearce and Birkett, 1974; Fleet and MacRae, 1975; Jolly, 1975; Gélinas et al., 1976; Arndt, 1977a; Arndt et al., 1977; Arth et al., 1977).

The ultramafic lava flows show the characteristic spinifex textures developed in the Barberton ultramafic komatiites (Naldrett and Mason, 1968; Pyke et al., 1973; Arndt et al., 1977), and pyroxene spinifex textures are well developed in the mafic komatiites (Arndt et al., 1977). As with the Rhodesian and Australian komatiitic rocks, the Abitibi lavas have many chemical similarities to the Barberton komatiites except for the high CaO/Al_2O_3 ratio. Arndt et al. (1977) have re-defined the komatiite classification scheme. On the basis of field petrographic and chemical criteria, they recognise three komatiite types: the peridotitic, pyroxenitic and basaltic komatiites. One set of their chemical criteria for discriminating between komatiites and tholeiites (a plot of $FeO^*/(FeO^*+MgO)$ vs. Al_2O_3) gives rise to an undesirable paradox whereby the komatiites from the Barberton greenstone belt fall in their tholeiitic field. One important feature noted by Arndt et al. (1977) is the complete gradation of chemical compositions from the most ultramafic (highest MgO) to the least mafic (lowest MgO) komatiite. They suggest the spectrum of compositions can be derived by crystal fractionation and/or accumulation processes.

Although other greenstone belts and Archaean granitic terrains are well documented in the literature (i.e. India, West Greenland, Minnesota and Finland), they will not be discussed in this brief

introduction. However no review, however brief, would be complete without mention of the recent geochronological findings in West Greenland. Detailed work has shown that the ancient Amîtsoq gneisses existed 3700-3800 Ma. ago (see discussion in Moorbath, 1978) and incorporated in these gneisses are remnants of even older sequences of ultrabasic, basic and metasedimentary rocks, the Isua supracrustals, which may be fragments of greenstone belts (McGregor and Mason, 1977). Recent ages established for Isua supracrustals, (basic and acidic metavolcanics) by the Sm-Nd isotopic method, of 3770±42 Ma. (Hamilton et al., 1978) show them to be ∿200 million years older than the Onverwacht lavas. Major and trace element compositions for some ultramafic rocks of the Akilia association (McGregor and Mason, 1977) are very similar to ultramafic lavas from the Onverwacht Group. However, these rocks have not been used for detailed comparison with the Onverwacht lavas in the following Chapters as they have been intensely metamorphosed and deformed.

1-6 OUTLINE OF OBJECTIVES FOR THIS THESIS

The detailed geological and geochemical investigation carried out by Viljoen and Viljoen (1969b,c,d,e) on the Onverwacht Group volcanics convincingly demonstrated the existence of a group of high magnesium ultramafic and mafic lavas as well as basalts of tholeiitic composition. The main aim was to obtain additional major element and precise trace element data from the komatiitic and tholeiitic lavas. The data were used to further develop and place constraints on models for the genesis and evolution of these rocks. The scope of the project has been limited to the lavas occurring in the Lower Ultramafic Unit. Sampling of the lavas

was predominantly restricted to the Komati Formation type area (see Viljoen and Viljoen, 1969c,d) as the rocks within this area appear to be better preserved, texturally and mineralogically, than in either the Sandspruit or Theespruit Formations. However, a number of samples of the different rock types from these latter Formations, as well as from the Upper Mafic-Felsic Unit, were also analysed for comparative purposes. The sampling methods and analyitcal techniques used in this study are described in Appendix II. Brief petrographic descriptions of all the samples analysed are given in Appendix I.

Before detailed interpretation of the geochemical data could be undertaken, several other problems had to be investigated. One of the most serious problems affecting the geochemical data of these samples is that of alteration. Although most of the samples finally selected for analysis showed good textural preservation and minor carbonate and epidote contents, some of the major and trace element contents of the samples have been changed by the processes of alteration that have affected these rocks. The alteration problem was investigated using a suite of basaltic komatiite pillows. The data are discussed in Chapter 2. The second problem investigated is the extent to which crystal fractionation has modified the composition of relatively thin ultramafic and mafic komatiite flows. Textural and geochemical data have been obtained for two flows, one of ultramafic composition and the other of mafic composition. These data and the models developed to account for the observed variations are discussed in Chapter 3.

Another problem that has become apparent recently is that the definition and classification scheme originally proposed by Viljoen

and Viljoen (1969c) for the komatiites is unsatisfactory. This problem is discussed at length in Chapter 4, and using selected data from this project a new classification scheme is proposed. The geochemistry of the ultramafic komatiites, mafic komatiites and tholeiitic basalts is discussed in Chapters 5-7 respectively. The final Chapter is more speculative in nature, dealing with models on the origin of komatiites and mantle compositions.

CHAPTER 2.

ALTERATION EFFECTS ON THE GEOCHEMISTRY OF MAFIC KOMATIITE LAVAS FROM THE KOMATI FORMATION

2-1 INTRODUCTION

Geochemical data and lavas from Archaean greenstone belts have been utilized in a diverse set of fields such as for radiometric age determinations, classification of rock types, models of petrogenesis and crustal evolution and for monitoring possible temporal evolution of the mantle compositions. A general problem that besets these investigations is one of mineralogical and chemical reconstitution of the rocks by processes of sea water alteration, metamorphism and recent subsurface weathering. The effects of chemical modification of the lavas must be considered before the geochemical data can be satisfactorily interpreted in any detailed way. A number of studies have been carried out on aspects of textural, chemical and mineralogical alteration of Archaean greenstone lavas (Williams, 1971, 1972; Williams and Hallberg, 1973; Viljoen and Viljoen, 1969b; Pearce and Birkett, 1974; Condie et al., 1977). Generally these studies have been oriented towards investigating the geochemical changes occurring between samples that show the least textural and mineralogical alteration and those that display extensive development of secondary minerals, such as epidote, chlorite and calcite.

In this paper the effects of alteration (sensu lato) on the geochemistry of mafic komatiite pillows (see Chapter 4 for definition of terms used in this work) from the Komati Formation, Barberton greenstone belt, are reported. Pillows with well preserved igneous texture and mineralogy were selected and variations in the major and trace element compositions in different portions of the pillows were investigated. Massive flows were considered unsuitable material for the purposes of

this investigation as any chemical alteration effects may have been superimposed on chemical heterogeneity caused by crystal fractionation and accumulation in the flows.

Previous studies on the effects of alteration on the volcanic rocks from the Barberton greenstone belt have been carried out by Viljoen and Viljoen (1969b) and Condie et al. (1977). Viljoen and Viljoen (1969b) studied the effects of carbonation on the major element chemistry of pillowed tholeiitic basalts from the Hooggenoeg Formation. They conclude, often from poorly defined trends, that SiO_2, Al_2O_3, MgO, CaO and K_2O contents of the samples have decreased, the FeO, Na_2O, TiO_2 and H_2O contents haveincreased with increasing carbonation. The Fe_2O_3 content of the pillows remained roughly constant. Some of these changes may in part have been due to real igneous variations as the samples were taken from a 2 km stratigraphic section of the Hooggenoeg Formation. Condie et al. (1977),working on massive mafic flows from the Hooggenoeg Formation, found that the development of secondary minerals such as epidote, chlorite or calcite could be correlated with some of the observed major and trace element variations in the samples. Certain of their conclusions will be compared to the results of this study later in the text.

Numerous studies have been published on the effects of sea water and low grade metamorphic alteration processes on the mineralogy and geochemistry of basaltic lavas from the ocean floor (e.g. Melson and van Andel, 1966; Melson et al., 1968; Cann, 1969; Hart, 1970; Hart and Nalwalk, 1970; Hart et al., 1974; Aumento et al., 1976; Scott and Hajash, 1976; Shido et al., 1974; Humphris and Thompson, 1978a,b). It is not the intention here to review these data,although the results of many of these studies are relevant to the alteration of Archaean greenstone belt lavas and will be referred to where appropriate in the text. Investigations

of the effects of metamorphic mineral assemblages on the chemistry of altered basaltic rocks (Smith, 1968; Jolly and Smith, 1972; Smith and Smith, 1976; Harrigan and MacLean, 1976) from the continental environment, will be similarly treated.

2-2 GEOLOGY AND SAMPLE LOCATION

The geology of the Lower Ultramafic Unit of the Onverwacht Group, Barberton greenstone belt, has been described in detail by Anhaeusser et al. (1968) and Viljoen and Viljoen (1969a,c), and consists of pillowed and massive flows of basaltic and ultramafic lavas, particularly well exposed in the Komati Formation. Two sets of samples of mafic komatiite pillows were obtained. The first set (the eastern pillows, SC-3 to SC-13, see Fig. 2-1) was sampled from a well exposed basaltic horizon near the type section established for the Komati Formation (Viljoen and Viljoen, 1969c). In this area the lavas show the best preservation of their original igneous mineralogy. The second set of samples (the western pillows, SC-2 and SD-82 to SD-85) was taken from pillows along the same horizon but 2 kms to the west of the first set. This second set of samples is a lighter green colour than the first, indicating that they are slightly more altered (Viljoen and Viljoen, 1969b). Sampling and analytical techniques are given in Appendix II.

2-3 PETROGRAPHY AND MINERALOGY

The pillows sampled, ranged in diameter from 70-100 cms and are rimmed by a ∿2 cm thick, dark grey-green, chilled margin. The interior of the pillows contained ocelli up to 1 cm in diameter (which tend to increase in number and size towards the centre of the pillows) as well as oblate gas cavities, some of which are quartz-filled. Thin cracks (0.1 mm wide) can be seen extending from the margin up to 2 cms into the

interior of the pillows. A few large cracks (1 mm wide) penetrate into the core of the pillows. This fracture system may represent relict radial cooling cracks as described by Moore et al. (1971) and Moore (1975) for younger pillow lavas.

In thin section the chilled margin consists predominantly of brown cryptocrystalline material (mainly tremolite-actinolite and chlorite as identified by X-ray diffraction) which is interpreted as devitrified glass and three minor microphenocryst phases (<2 volume %):-

1) Euhedral to subhedral pyroxene crystals (up to 0.9 mm long) which are now often pseudomorphed by actinolite and chlorite with variable amounts of epidote, quartz and calcite.

2) Occasional calcic plagioclase crystals which are usually altered to albite with minor quartz, clinozoisite, sericite and calcite.

3) Glomeroporphyritic spinel octahedra that are scattered throughout the devitrified glass and also occur as inclusions in the plagioclase and pyroxene microphenocrysts.

Epidote is present as small discrete grains within the devitrified glass of the SC-2 margin. Small hollow euhedral crystals (now replaced by quartz) occurring in the SC-11 and SC-3 pillow margins are interpreted to have originally been hopper olivines, similar to examples described by Donaldson (1974, 1976).

The interior of the pillows consists of a groundmass of fan-shaped spherulites of actinolite with occasional relict phenocrysts of diopside, which however are usually replaced by actinolite ± chlorite. Chlorite,

quartz, epidote and minor calcite are also present in the groundmass of all the pillow cores, while minor biotite is developed in the groundmass of the intermediate and core samples of the SC-3 pillow. The ocelli which only occur in the interior of the pillows have been adequately described by Ferguson and Currie (1972) and commonly contain relict diopside crystals. The glomeroporphyritic spinel microphenocrysts are scattered throughout the core samples including the ocelli. The normative minerals have been calculated (Table 2-1) assuming a Fe_2O_3/FeO ratio of 0.2 and on this basis all the pillows are quartz normative.

2-4 COMPOSITION OF THE OCELLI

Before the detailed chemical variations within the pillows can be discussed, it is necessary to consider the possible complications that may be caused by the inhomogeneous distribution of the ocelli on the chemistry of the pillow samples. Major and trace element data for the ocelli cut from the SC-13 core sample are given in Tables 2-1 and 2-2. Compared to the whole rock composition, the ocelli are enriched in SiO_2, Na_2O, S and Sr and slightly enriched in Zn, Cu, Ni, Cr and Ga. The other elements are depleted by varying degrees in the ocelli, K_2O, Rb and Ba being particularly low. Additional analyses of ocelli from rocks of mafic komatiite composition are listed in Viljoen and Viljoen (1969c) and Ferguson and Currie (1972). The data given by Ferguson and Currie (1972) show that K_2O is behaving inconsistently, being higher in the ocelli than the matrix in two samples and lower in a third. The inconsistent variations of K_2O may well be related to alteration rather than being a primary igneous feature.

The data (Tables 2-1 and 2-2) for the ocelli-free intermediate

sample (SC-3I) and the ocelli-bearing core sample (SC-3C) of the SC-3 pillow are similar for many elements (e.g. TiO_2, Zr, Co and V). No systematic trends consistent with the addition or removal of ocelli (of similar composition to those in the SC-13 sample) can be observed between the composition of the SC-3 intermediate and core samples. The alkali elements and their inter-element ratios have not been considered here as these elements, have probably been re-distributed during the alteration processes that have affected the pillow lavas. It is therefore concluded that the core sample taken from the SC-3 pillow (and by analogy all the core samples) was sufficiently large to ensure homogeneity with respect to the ocelli. As no ocelli are observed in the chilled margins and the composition of margin-core pairs for key elements such as SiO_2 and TiO_2 are very similar, (see Table 2-1) the time of formation and development of the ocelli must have occurred after the pillows had formed, possibly while the interior was still liquid.

2-5 CHEMICAL VARIATIONS WITHIN INDIVIDUAL PILLOWS

The major and trace element data for all the margin and core samples are given in Tables 2-1 and 2-2 and graphically displayed in Fig. 2-2. All the pillows are Barberton type basaltic komatiites using the characteristics outlined by Viljoen and Viljoen (1969c). However, using the nomenclature given in Chapter 4, these lavas are termed evolved low-Mg mafic komatiites in this work. An average has been calculated for all the pillow samples used in this study and given in Table 2-4. Attention is first drawn to those elements that show no variation in the different portions of individual pillows, within analytical error (AE = $\pm 2\sigma$ i.e. 4σ) and thereafter to those elements that show significant variations.

Examination of the plots (Fig. 2-2) shows that for the pillow margin-core pairs, no variation in the concentrations occurs for TiO_2, P_2O_5, Nb, Zr, Y, Co, V and Sc within AE. However, for some of these elements, the AE is relatively large compared to the concentration of the elements in the samples. The AE has been expressed as a percentage of the average for each element (Table 2-4). From this data it can be seen that variations greater than 5% relative (or less for some elements) in TiO_2, Zr, V and Sc would be considered analytically significant, while larger variations in the concentrations of Co, P_2O_5, Y and Nb (a maximum of respectively 8%, 12%, 13% and 36% relative) could occur and not be considered significant. Bearing these constraints in mind, this complete group of elements is considered to be homogeneously distributed within the pillows and as such, have remained immobile during the alteration processes that the Komati Formation lavas have been subjected to.

TiO_2, Nb, Zr and Y have previously been shown to remain immobile during low-grade metamorphic and sea water alteration processes (Cann, 1970; Pearce and Cann, 1973; Hart et al., 1974; Humphris and Thompson, 1978a,b), while the evidence that P_2O_5 remains stable under these conditions has not been convincing (see for example, Hart, 1970; Hart et al., 1974). In view of the current usage of P_2O_5 in the classification of tectonic environments of basaltic rocks (Pearce et al., 1975; Floyd and Winchester, 1975; Winchester and Floyd, 1977) the consistent behaviour of this oxide in the Komati Formation pillows is additional evidence for its immobile nature during low temperature alteration processes. Co shows very little change with varying degrees of alteration in sea floor basalts (Hart et al., 1974; Aumento et al., 1976), and Condie et al. (1977) conclude that Co is not significantly affected by

carbonization-chloritization processes but is depleted during epidotization of tholeiitic lava flows from the Barberton greenstone belt. Hallberg (1972) has analysed Archaean pillow rim-core pairs from the eastern Goldfields region of Western Australia. Co varies inconsistently in these pillow pairs by small amounts, although the spread in Hallberg's Co data cannot be assessed as the AE's for the trace elements were not reported. V from the core samples of the Western Australia pillows is significantly lower than in the margins, unlike the Komati Formation pillows. However Hart et al. (1974) have also observed very little change of V with varying degree of alteration in sea floor basalts. The use of Nb in this study is limited by the difficulty in obtaining precise data by X-ray fluorescence at the low levels encountered.

Elements that remain immobile during alteration processes of initially homogeneous material may still show variations in concentration between samples depending on the nett addition (i.e. dilution) or removal (i.e. concentration) of components from the samples (Gresens, 1967). Although H_2O and CO_2 have been added to these pillow samples in varying amounts and other elements removed during alteration, the difference in the nett effect between core and margin samples, on immobile minor and trace elements, is often too small to be detected within the precision of the analytical method. These effects however can be noticed on elements that can be determined relatively precisely (see Table 2-4) such as SiO_2. Examination of variations in the concentration of this element can be more constructively carried out using inter-element ratios rather than absolute abundances. SiO_2 and Al_2O_3 show minor changes within pillow pairs (generally within AE), except for SiO_2 in the SC-2 pillow and Al_2O_3 in the SC-3 pillow samples. The Si/Ti and Si/Al ratios (Table 2) of the SC-2 margin and core samples

are essentially identical which suggests that SiO_2 has not been removed from the margin sample, but diluted. Al_2O_3 appears to have been leached from the margin of the SC-3 pillow as this sample has lower Al/Ti and higher Si/Al ratios than the intermediate and core samples. Apart from this one sample, Al_2O_3 shows no significant variations and it is concluded that SiO_2 and Al_2O_3 have generally remained immobile during the alteration processes these lavas have been subjected to; similarly for MgO which shows analytically significant changes in one margin-core pair but no significant variation in the other 4 pillow pairs.

The remaining elements and oxides show changes within the pillow pairs which are often much greater than the AE. Explanation for these changes by crystal fractionation or accumulation are attempted first, thereafter alteration effects are invoked. The important phenocryst phases in this respect are clinopyroxene (electron microprobe analyses of diopside from massive flows of basaltic komatiite of similar composition have MgO and CaO contents of 18-20 wt.% see Chapter 6), calcic plagioclase and spinel.

Probably the best indicator of chemical alteration is the H_2O^+ and CO_2 contents of the samples. No systematic trend can be seen in the H_2O^+ content between the margin and cores of these pillows. The SC-2 margin and core have the highest H_2O^+ content which indicates that the SC-2 pillow is the most altered pillow analysed in this suite of samples. The CO_2 contents of the samples are low and identical within AE for the pillow pairs. It follows therefore that the large but inconsistent variations of CaO within the pillow pairs (see Fig. 2-2) are not related to the formation of calcite in the samples. Processes of diopside or plagioclase accumulation or removal can also be rejected for accounting for the variable contents of CaO in the pillow

pairs, as samples with high CaO should also contain either high MgO or Al_2O_3 contents respectively, which is not observed (see Fig. 2-2). The high CaO (as well as Sr and Ga) in a sample such as SC-2M is probably related to the appearance of epidote grains in the devitrified matrix of this sample. Epidote in low-grade metamorphic rocks has been shown to accumulate Ca, Sr and to a lesser extent Ga (Melson and van Andel, 1966; Smith, 1968; Smith and Smith, 1976; Condie et al., 1977; Willis, 1978). It is therefore suggested that the erratic variations of CaO, Sr and Ga between the pillow pairs, are due to the addition of variable amounts of these elements into the samples to form epidote during metamorphism.

FeO shows a regular decrease in the cores of the pillows compared to the margins and similarly for MnO, although some of the changes are within AE. By contrast Fe_2O_3 remains constant within AE for most of the pillow pairs. The loss of iron from cores of recent marine pillows has been recorded by Scott and Hajash (1976). This feature in the Barberton pillows may be a relict sea water alteration effect and suggests that the cores are more altered than the rims with respect to their FeO contents. Hart et al. (1974) and Scott and Hajash (1976) have shown that the margins of recent pillow lavas are not always the most altered portion of the pillows and may often retain their primary composition better than the more crystalline interiors.

The K_2O, Rb and Ba contents of the pillow pairs show similar variable distribution patterns. No variation for K_2O occurs in the SC-11 and SC-5 pillow pairs within AE. If an analogy can be drawn with the alteration processes in modern abyssal basalts (Hart et al., 1974), the low K_2O, Rb and relatively high K/Rb of the SC-11 pillow indicates that this is a comparatively fresh pillow, while the high

K_2O, Rb and low K/Rb of the SC-2 pillow and SC-8 core suggests they are the most altered samples.

Na_2O also shows variable distribution in the pillow pairs. The calcic plagioclase microphenocrysts have usually been replaced by albite, a feature which is typical of spilitic rocks (Vallance, 1965; Melson and van Andel, 1966; Cann, 1969). These pillows however do not show anomalously high Na_2O compared to other fresh basaltic rocks, neither do they have the characteristic distribution of low Na_2O in the margins and high Na_2O in the cores (Vallance, 1965) and therefore spilitic affinities for the Barberton pillows are discounted.

Cr shows small but significant changes within the pillow pairs; these changes could however be accounted for by minor variations in the content of the spinel microphenocrysts and need not be an alteration effect. A similar feature has been noted by Humphris and Thompson (1978b) for the behaviour of Cr in Mid-Atlantic ridge greenschist facies pillow samples. The remaining elements Zn, Cu, Ni and S show large and variable changes between cores and margins of the Komati Formation pillows. S and Cu show very similar distribution patterns in the pillow pairs (Fig. 2-2) and are lower in all the margins than cores except for the SC-8 pillow pair. Consistently lower S content in the margins of recent ocean floor basalts has been recorded by Hart et al. (1974) and this similar trend for most of the Barberton pillows may be a relict sea water alteration effect.

In summary the elements determined in margin-core pillow pairs from the Komati Formation can be divided into two groups:-

A) An immobile group of elements that show no detectable change within the pillow samples and comprise Si, Ti, Al, Mg, P, Nb, Zr, Y, Co, V and Sc. Cr probably also belongs to this group.

B) A mobile group of elements that show changes in the different portions of the pillows that vary from being small but significant to huge changes of several hundred percent. This group of elements consists of:- Fe, Mn, Ca, Na, K, S, Sr, Rb, Zn, Cu, Ni, Ba and Ga.

2-6 THE ACID LEACHED PILLOW SAMPLES

The data for HCl leached powder samples (methods given in Appendix II) are given in Table 2-3. While it was expected that only carbonate would be removed by the HCl leaching process, comparison of the data in Table 2-3 with the relevant analyses from Tables 2-1 and 2-2 show that this is not the case. The following changes can be observed:-

A) SiO_2 and Na_2O are higher in the leached samples. The changes for SiO_2 are significant for all the samples except the SC-5 margin and only significant for Na_2O in the SC-2 core sample.

B) Al_2O_3, TiO_2, $Fe_2O_3^*$, MnO, MgO, Nb, Zr, Y, Sr, Co, Cr, V and Sc show no change within AE or very minor changes.

C) CaO, P_2O_5, Zn, Cu and Ni are systematically lower in the leached samples.

D) K_2O, Rb and Ga have remained constant in the leached SC-2 pillow samples but show a significant decrease in the leached SC-5 pillow samples, while Ba shows a significant decrease in the SC-2 leached pillow samples but only a slight decrease in the SC-5 leached samples.

The decrease of the C group elements in the leached samples is attributed to the removal of small amounts of calcite, apatite and sulphide in the acid solution. The Sr data are essentially identical in the

leached and unleached samples suggesting an insignificant amount of the total Sr in the samples was incorporated in the leached calcite.

2-7 ALTERATION EFFECTS BETWEEN PILLOWS

In the ensuing discussion on the alteration effects that can be identified between pillows, the assumption has been made that all the samples formed from a homogeneous magma source and that crystal fractionation processes have only played a minor part in modifying the initial lava compositions. Inter-element ratios of the immobile elements (as identified in the previous section) such as Si/Al, Al/Ti, Zr/Y and Ti/P (see Table 2-2) show very little variation between the pillows and tend to support this assumption. On this basis the western pillows (SC-2, SD-82 to SD-85) are more altered than the eastern pillows (SC-3 to SC-13) as they have, on average, higher H_2O^+, Fe_2O_3 and Fe_2O_3/FeO.

Perusal of the data in Tables 2-1 and 2-2 and in Fig. 2-2 shows that the immobile elements such as Ti, P and Zr display very little variation in their contents between pillows, while the mobile elements (e.g. K, Rb, S) show a much greater range of concentrations regardless of whether margins only, cores only or the average of margin-core pairs are considered. An attempt to quantify the range observed for each element has been made in Table 2-4. Here the range observed for an element (all data recalculated volatile free) in all the margin and core samples minus the spread due to AE (4σ) has been expressed as a percentage of the average composition as below:-

$$\text{Real Spread (RS)} = \frac{((\text{Max}-\text{Min})-\text{AE})}{\text{Average}} \times 100 \quad \ldots\ldots\ldots \quad 2\text{-}1$$

The negative RS values for P_2O_5 and Y indicate that the range observed for these elements in all the samples was less than that expected from

analytical error alone. Based on the RS values the elements can be split into three groups:-

1) RS<8% - Si, Ti, Al, P, Nb, Zr, Y, Co, V and Sc. This group of elements is essentially the same as the immobile group of elements identified in the previous section and it is suggested that the variations are sufficiently small to enable this group of elements to be used for detailed modelling of the igneous geochemistry.

2) RS 15-50% - Fe, Mn, Mg, Ca, Cr, Ni and Ga. While the range recorded for Cr within the pillows is probably due to the inhomogeneous content of spinel microphenocrysts, the use of the other elements to interpret the igneous geochemistry of the lavas should be treated with caution(from the Barberton greenstone belt and other similar greenstone belts). However the observed concentrations may nevertheless be useful indicators of the original composition of the lavas.

3) RS>65% - Na, K, S, Sr, Rb, Ba, Cu and Zn. The huge variations recorded for this group of elements makes any estimate of the original content in the lavas extremely unreliable even within the better preserved eastern set of pillows.

Obviously, therefore, the abundance of this last group of elements obtained by the direct analysis of the Barberton greenstone belt lavas should not be used for such subtle purposes as inferring source areas characteristics or classifying rock types. A more valid approach has been adopted by Hart and Brooks (1977) who have attempted to infer the source area characteristics by reconstructing the K, Rb and Ba inter-element ratios of altered Archaean lavas using the K, Rb and Ba contents of high purity clinopyroxene separates and the appropriate

distribution coefficients. Erlank and Kable (1976) have used the Zr/Nb ratio to infer the source area characteristics, which is potentially useful when dealing with altered Archaean lavas. The average Zr/Nb ratio of 20 from these Komati Formation pillows is high and suggests that these lavas were derived from a depleted source area with respect to modern oceanic island basalts, but similar to average mid-ocean ridge tholeiite (MORB). Variations in inter-element ratios are further discussed in Chapter 7.

In view of the varying significance that has been attached to the CaO/Al_2O_3 ratio in classification schemes of komatiites (Viljoen and Viljoen, 1969c,d; Brooks and Hart, 1974; Arndt et al., 1977), the effect of alteration on this ratio will be discussed in more detail. The possibility of plagioclase or diopside accumulation in the SC-2 margin has already been rejected; similar reasoning also eliminates the possible accumulation of these minerals in the more altered (western) pillows to account for the high CaO/Al_2O_3 ratios. The mineralogical control of the chemistry of altered rocks has been well established by Melson and van Andel (1966), Smith(1968), Cann (1969), Jolly and Smith (1972) and Leake (1972); in particular the increase of CaO and Fe_2O_3 with increasing epidote and H_2O^+ with increasing chlorite. A chlorite (ripidolite) analysed by electron microprobe from the interior of the SC-11 pillow has a CaO content of 0.05 wt.% (see Table 5-2). Variation in the chlorite content as such, therefore, will not affect the CaO/Al_2O_3 ratio, as chlorite contains insignificant calcium and because it appears that Al_2O_3 has neither been added to nor removed from the samples. The alteration of calcium rich minerals to chlorite could possibly decrease the CaO/Al_2O_3 ratio if the CaO was removed from the rock. The addition of the required H_2O^+ to form the hydrous

alteration minerals in these rocks will have the effect of diluting the other elements. Again because of the apparent immobility of Al_2O_3, the addition of a likely epidote composition (clinozoisite-epidote) to the samples is impossible without changing the Si/Ti and Al/Ti ratios. If on the other hand CaO and Fe_2O_3 were added and formed an epidote mineral from the Al and Si fixed in the rock, significant changes would be observed in the CaO/Al_2O_3 and Ca/Ti ratios, but no changes would occur for the Si/Ti, Al/Ti and Si/Al ratios. In Fig. 2-3 the CaO/Al_2O_3 ratio has been plotted against Fe_2O_3 for all the samples. The western pillows form a positive trend of increasing CaO/Al_2O_3 with Fe_2O_3. Two trend lines have also been calculated for the addition of CaO and Fe_2O_3 in the correct proportions to form epidote with the Fe_2O_3/CaO ratios indicated. The two epidote compositions were taken from Deer et al. (1962 p. 197 analysis No. 4 and No. 5). The starting point for calculating the trend was chosen rather arbitrarily to be within the lower CaO/Al_2O_3 group of fresher pillow and the SC-5 margin composition was used. The two epidote formation trends enclose the observed trend which indicates that the epidote actually formed should have a Fe_2O_3/CaO ratio between 0.22 and 0.32. Forming an epidote mineral with Fe_2O_3/CaO ratio of 0.27 by addition of 4.52% CaO and 1.22% Fe_2O_3 to the starting composition gives the CaO/Al_2O_3 and Fe_2O_3 content observed in the SC-2 margin. The exact percentage will depend critically on the chosen starting point and epidote composition, but an attempt has been made in Table 2-5 to check whether these are realistic figures using the SiO_2, TiO_2 and Zr data. The data have been calculated volatile free to eliminate the dilution effects of H_2O^+; the SC-2 margin data has in addition been recalculated up by 5.7% after the removal of 4.52 wt.% CaO and 1.22 wt.% Fe_2O_3. The

average fresher pillow composition in Table 2-5 has been obtained by averaging all the eastern pillow samples (volatile free data). The agreement between the recalculated SC-2 margin data with that of the average fresher pillow (Table 2-5) is remarkable and well within analytical precision for the three elements. This data is consistent with the interpretation that these elements have been diluted by ∿5.7% in the SC-2 margin compared to the average fresher pillow. From Fig. 2-3 and associated calculations, it appears that the addition of CaO and Fe_2O_3 to the more altered samples to form epidote is a real possibility, but is difficult to verify modally as the rocks are extremely fine-grained. However, the observation that epidote is more abundant in the SC-2 margin than the other margin samples tends to support the above suggestion. In addition 14-J (analysis in Viljoen and Viljoen, 1969c), a pillow of similar composition to the pillows analysed in this study, contains abundant epidote in thin section as large (0.1-0.2 mm) discrete grains and has a high CaO/Al_2O_3 ratio. From this and the other circumstantial evidence presented, it is concluded that the addition of CaO and Fe_2O_3 to these pillows to form epidote has seriously affected the CaO/Al_2O_3 ratio of the more altered samples. In view of this and because the fresher pillows have CaO/Al_2O_3 ratios as low as 0.83, the suggestion that $CaO/Al_2O_3>1$ (Brooks and Hart, 1974) be used as one of the main parameters for defining the low MgO komatiites, is not valid for these Komati Formation lavas.

2-8 DISCUSSION AND CONCLUSIONS

Among the common suite of elements that have been determined in this study and that of Condie et al. (1977) there are some serious discrepancies with regard to the elements that are considered to be

immobile. Samples used by Condie et al. (1977) show a far greater degree of alteration in terms of secondary mineral development than the samples used in this study, and comparison therefore will be restricted to the elements that, according to Condie et al. (1977), are immobile. They suggest that Ba, Ga and Ni among others do not appear to have been greatly modified by alteration as they show the same variabilities in altered and unaltered flows. These same three elememts are highly variable both within and between pillows from the Komati Formation studied here. The differences could possibly be related to sea water alteration on the pillow samples. However inspection of the so-called "unaltered" Londozi River flow data of Condie et al. (1977) shows Ba to vary from 97 to 236 ppm, while in the carbonated flow Ba varies from 21 to 57 ppm and in the epidotized flow from <15 to 41 ppm. It is suggested that these huge variations are due to alteration, considering the erratic behaviour of Ba found in the Komati Formation pillows. In addition Condie et al. (1977) note an 11 to 17 ppm increase of Ga with increasing epidotization and it is therefore puzzling as to why they consider Ga not to have been greatly modified by alteration processes. This difference of 6 ppm Ga between the relatively fresh sample and the altered sample becomes extremely significant when models for the genesis of these lavas are being tested. Arth (1976) has emphasised the need for precise and accurate analysis of trace element data when testing geochemical models. To this must be added further constraints, when modelling altered rocks, that only elements which have remained immobile throughout the alteration processes, should be used for modelling.

It is apparent from the data obtained in this study that both margins and cores of all the Komati Formation pillows have been altered

to some extent. The variations of Fe^{2+}, Mn, S and Cu between margin and core of these pillows show similar trends to those observed in altered ocean floor pillow lavas (Hart et al., 1974; Scott and Hajash, 1976; Humphris and Thompson, 1978a,b) and may therefore reflect the effects of alteration by Archaean 'sea' water. The addition of small amounts of Ca and Fe^{3+} (probably some Ga and Sr as well) to some of the samples to form epidote during a low grade metamorphic event has been suggested. However the variation in the contents of the other mobile elements cannot be assigned to a specific alteration process as yet and may have been remobilized by more than one alteration event.

The major and trace elements determined in this study can be divided into three groups when all the samples analysed are considered:-

A) Si, Ti, Al, P, Nb, Zr, Y, Co, Cr, V and Sc have remained immobile or relatively so to the extent that these pillow lavas have been altered and are considered to reflect the igneous geochemistry of the rocks.

B) Fe, Mn, Mg, Ca, Ni and Ga. Undoubtedly for this group of elements the selection of samples is going to be the most critical if the data is to be employed usefully for igneous interpretations. It is suggested that sample selection should be based on the following criterion:-

1) Igneous texture preservation
2) Relict mineral content
3) The lack of the development of monomineralic metadomains (Jolly and Smith, 1972) of minerals such as epidote and chlorite.

C) Na, K, S, Sr, Rb, Ba, Cu and Zn. The large variations recorded for this group of elements both within and between pillows

indicate that the concentrations of these elements have been extensively modified by alteration processes and little reliability can be placed on their absolute concentrations or inter-element ratios, determined directly in the rocks, to infer the original igneous character.

Finally, the behaviour of the elements found in this study of the Komati Formation mafic komatiite pillows is specific to this locality and other greenstone belt lavas that have undergone very similar alteration histories. The chemical elements in rocks that have been metamorphosed to higher or lower grades or have different primary and reconstituted mineral assemblages, may behave differently.

CHAPTER 3.

ULTRAMAFIC AND MAFIC KOMATIITE LAVA FLOWS

ULTRAMAFIC KOMATIITE LAVA FLOWS

3-1 INTRODUCTION

Viljoen and Viljoen (1969d) recognised a class of rocks which occur in the Barberton greenstone belt as the crystallisation products of ultramafic magmas and termed them peridotitic komatiites. These ultramafic rocks crop out as massive flows and pillow lavas associated with mafic komatiites and tholeiitic basalts in the lower Formations of the Onverwacht Group (Viljoen and Viljoen 1969c,d). Several ultramafic komatiite lava flows from the type area of the Komati Formation have been sampled in order to determine the extent of compositional variation within these flows and to test whether such variations can be ascribed to crystal fractionation processes, as has been suggested by Pyke et al. (1973), Arndt (1977a), Arndt et al. (1977) and Arth et al. (1977) for similar lava flows from other Archaean greenstone belts. In this section the textural relationships of three spinifex textured ultramafic komatiite flows are reported, together with compositional data for samples taken from two of the flows. The thickest flow (3.1 m) has been investigated in the most detail and analysed mineral compositions have been utilised in modelling systematic compositional variations within the flow. The textural and compositional data are combined to develop a differentiation history and cooling model for this typical ultramafic flow.

3-2 FIELD RELATIONS AND PETROGRAPHY

A well exposed section across the three ultramafic flows occurs

in a stream bed in the type area of the Komati Formation, but the lateral extent of the flows could not be traced due to poor exposure. Fig. 3-1 is a diagrammatic section of the flows showing the sampling points (arrows), textural features and thickness of the individual flows. Major and trace element data for seven samples have been obtained. One analysis is from the basal chilled margin of the central flow and six are from the lower flow (see Fig. 3-1). Following the tradition of Arndt et al. (1977) the lower (3.1 m thick) flow is referred to as Stuart's flow.

These three flows from the Komati Formation show a striking textural resemblance to the ultramafic flows containing spinifex texture from Munro Township (Pyke et al., 1973) and Western Australia (Nesbitt, 1971). Pyke et al. (1973) have divided the Munro Township ultramafic flows into six sub-units on the basis of structure and textural features, while noting that certain sub-units may not be developed in every flow. Their terminology will be adopted in this description of the Komati Formation ultramafic flows, but with minor modifications. It should be noted that where olivine is discussed in the following section, all the olivine crystals have in fact been pseudomorphed by serpentine and magnetite except for the B zone samples of Stuart's flow where kernels of primary olivine are occasionally preserved.

A chilled flow top (Pyke et al.'s A_1 zone) was not observed in any of the flows in Fig. 3-1, however an aphanitic basal contact (lower B_4) zone could be clearly identified in the upper two flows. In thin section the basal contact samples consist of a microcrystalline groundmass of serpentine amphibole and magnetite with occasional euhedral microphenocrysts of olivine. Small spherical vesicles (0.06-1.1 mm

across) occur in the contact zone that are often lined with magnetite and contain large single crystals of calcite.

The spinifex (A_2) zone is well developed in all three flows and in common with other spinifex textured ultramafic flows (Pyke et al., 1973; Nesbitt, 1971; Arndt et al., 1977; Lajoie and Gélinas, 1978) they display two distinct types of spinifex texture:-

1) An upper zone consisting of short (<10 mm long) olivine blades that generally show no preferred orientation. Nesbitt (1971) has described this texture as 'radiating spinifex' (see Plates 12b, 14a and 14c in Nesbitt, 1971). Donaldson (1974) has objected to this term as not all the olivine blades radiate from a common nucleus and prefers the term 'randomly oriented spinifex'. In this thesis Donaldson's term will be abbreviated to 'random spinifex'. Lajoie and Gélinas (1978) have designated this zone as the A_2 division in ultramafic flows from La Motte Township, Quebec. However, as the symbol A_2 has already been used by Pyke et al. (1973) to designate the entire spinifex portion of the flow, the symbol A_{2r} is preferred (see Fig. 3-1). Within the A_{2r} zone the olivine blade length increases from top to bottom, the average length being less than 10 mm. Occasional hopper olivines similar to examples described by Donaldson (1974, 1976) occur between the blades. Small vesicles with the same features as these in the chilled basal contact samples are also found. The olivine blades are set in a groundmass of amphibole (probably originally skeletal pyroxene), serpentine and magnetite.

2) A lower zone consisting of well oriented fan-like arrangements of olivine plates up to 300 mm in length. This type of spinifex texture has been well documented in the literature (Viljoen and

Viljoen, 1969d; Nesbitt, 1971; Pyke et al., 1973). Nesbitt (1971) has described this texture as 'plate spinifex' and this term will be used here. Lajoie and Gélinas (1978) have designated this zone as the A_3 division, but the symbol A_{2p} is preferred. The matrix in the interblade area consists of radiating sprays of amphibole (after pyroxene), serpentine and magnetite. No hopper olivines or vesicles have been observed in the interblade area of the A_{2p} spinifex zone. Although the olivine blade length increases from top to bottom of the A_{2r} zone the boundary between the A_{2r} and A_{2p} zones is marked by a sharp increase in average blade length.

A B_1 zone consisting of foliated skeletal olivine crystals was not obvious in any of the three flows in the field. However inspection of a polished slab of a sample taken from the central flow (see Fig. 3-1) revealed a thin B_1 zone (10 mm thick) below the A_{2p} spinifex zone. Olivine blades in the B_1 zone (up to 5 mm long) are aligned roughly parallel to the flow contacts and set in a matrix of amphibole, serpentine and magnetite. However the major portion of the B zone in all three flows consists of fine grained peridotite (B_2/B_4 zone). In thin section the B_2/B_4 zone has a cumulate texture consisting of closely packed equant olivine grains (0.1 to 0.3 mm across) forming 60 - 80% of the rock. The mesostasis contains serpentine, amphibole, chlorite and magnetite. A 'knobby peridotite' B_3 zone was not observed in any of the flows in Fig. 3-1.

Relict olivine kernels have been analysed by electron microprobe in sample HSS-535 from the B_2/B_4 zone of Stuart's flow. The data are given in Table 3-1. Note that the 'margin' analyses referred to in Table 3-1 represent the edge of kernels and not the edge of original crystals. The olivines are normally zoned (FO_{92} to Fo_{90}) and are characterized by relatively high CaO and Cr_2O_3 contents.

The CaO content is consistent with the olivines having crystallized in an extrusive environment (Simkin and Smith, 1970; Dickey, 1972; Green et al., 1975). The high Cr_2O_3 content appears to be a feature of olivines from Archaean ultramafic lavas (Pyke et al., 1973; Bickle et al., 1975; Nisbet et al., 1977) and is discussed below.

3-3 GEOCHEMISTRY OF STUART'S FLOW

Three spinifex textured samples have been analysed from the A zone and three cumulate textured samples from the B zone of Stuart's flow (see Fig. 3-1). The major element data and normative mineral compositions are presented in Table 3-2 and the trace element data and interelement ratios are listed in Table 3-3. The major and trace element data have been recalculated volatile free. Mineralogical and chemical alteration of the mafic lavas from the Barberton greenstone belt have been investigated by a number of authors (Viljoen and Viljoen, 1969b; Condie et al., 1977; Chapter 2, this work). In Chapter 2 it was shown that the variations in abundance of Na, K, Rb, Sr, Ba, Cu, Zn and S in different portions of mafic komatiite pillows from the Komati Formation could not be accounted for by processes of crystal fractionation or accumulation alone. These elements therefore appear to have been mobile during one or more of the secondary processes (i.e. sea water alteration, greenschist facies metamorphism) that have affected the pillow lavas. These elements will therefore be excluded from the following discussion as their concentrations may well have been disturbed in the ultramafic rocks.

The variations of selected elements with position in Stuart's flow are illustrated in Fig. 3-2. In this plot it can be seen that MgO and Ni are depleted and SiO_2, TiO_2, Al_2O_3, CaO, Zr, Y, V and Sc are enriched in the A zone relative to the B zone samples. The

enrichment of the B zone in MgO and Ni is compatible with the olivine cumulate textures of samples from this zone, and suggests that olivine has settled from the A into the B zone.

A number of authors have proposed that the spinifex textured portion of ultramafic lava flows represents the initial composition of the magma (Viljoen and Viljoen, 1969d; Nesbitt, 1971; Nesbitt and Sun, 1976). Comparison between the three analyses of the spinifex samples shows that the A_{2r} sample (HSS-531) and the A_{2r}/A_{2p} sample (HSS-532) have virtually identical compositions. However the lower A_{2p} sample (HSS-533) is depleted in MgO, Ni and Co and enriched in TiO_2, CaO, Zr, Y, Cr, V, Ga and Sc relative to the other two spinifex textured samples. This difference in composition casts doubt on the proposition that any single spinifex textured sample represents the composition of the initial magma in Stuart's flow; the latter may be better determined by averaging the six samples taken from the flow. A weighted average composition has been calculated for Stuart's flow and given in Table 3-2 and 3-3; the weighting was based on the thickness of the flow each sample represents. This weighted average is very similar in composition to that of the basal chilled margin of the overlying ultramafic flow (HSS-530) indicating that the weighted average does represent a plausible magma composition. Donaldson (1974, 1976) has pointed out that the development of spinifex texture requires a nuclei free zone in the magma. The A_{2r} spinifex textured sample therefore probably represents the initial liquid (phenocryst free magma). Any microphenocrysts present in the original magma are inferred to have settled into the B zone before solidification of the A zone in Stuart's flow.

OLIVINE FRACTIONATION MODEL

Major element variations within Stuart's flow have been modelled assuming the weighted average composition as the parent magma and olivine (Fo_{92}) as the fractionating mineral. A least squares linear mixing method (Bryan et al., 1969) was used to calculate the amount of olivine removed from the A zone and accumulated into the B zone. Note that total iron has been calculated as FeO and that Na_2O and K_2O have been excluded; P_2O_5 was also excluded because of problems in obtaining precise measurements for this oxide at the low levels encountered in these ultramafic rocks. Samples with virtually identical compositions have been averaged (i.e. HSS-531 and HSS-532; HSS-534 and HSS-535) and the parameters for the four resulting calculations are presented in Tables 3-4a to 3-4d. CaO consistently shows the poorest agreement between observed and calculated composition. Inclusion of diopside in the mixing calculations did not improve the CaO error significantly. However the residuals ($\Sigma(Obs.-Calc)^2$) for the four computations are considered satisfactory in view of the altered nature of the flow.

The olivine fractionation model has been tested using the trace elements that are incompatible to the olivine structure (i.e. $D_i \sim 0$ where i = Zr, Y, V, Ga or Sc). The expected trace element content of the residual liquids (A zone) and cumulate (B zone) samples have been obtained from the expression:-

$$C^i = C_o^i/(1+X) \quad \ldots\ldots\ldots\ldots\ldots \quad (3\text{-}1)$$

where X = wt. frac. Fo_{92} added to the initial magma.

The agreement between observed and calculated trace element data for the four compositions from Stuart's flow (also in Tables

3-4a to 3-4d) is generally within the precision (±2σ) with which these elements have been determined and supports the olivine fractionation model developed from the major element data.

In this model the composition of the A_{2r} sample can be obtained by the removal of 4.6 wt.% Fo_{92} and the A_{2p} sample by the removal of 13.1 wt.% Fo_{92} assuming the weighted average represents the initial magma (liquid plus olivine phenocrysts). Similarly the composition of the B zone samples can be accounted for by the accumulation of between 8.9 and 14.4 wt.% Fo_{92} into the initial magma. The removal of larger amounts of olivine from the A_{2p} sample compared to the A_{2r} sample postulated in this model indicates that olivine continued to crystallise and settle from the magma after the formation of the A_{2r} zone.

3-4 FeO/MgO, Ni AND Cr CONTENT OF THE OLIVINES

In the olivine fractionation model developed for Stuart's flow, equilibrium between the most Mg rich olivine cores (Fo_{92}) and the initial liquid (A_{2r} composition) is implied and can be tested using element distribution coefficients between olivine and liquid that have been experimentally determined. Roeder and Emslie (1970) and Duke (1976) have found that for basaltic liquids in equilibrium with olivine $K_D = (X^{Ol}_{FeO}.X^{Liq}_{MgO})/(X^{Liq}_{FeO}.X^{Ol}_{MgO}) = 0.30$. In Stuart's flow the FeO content of the A_{2r} 'liquid' is unknown and the assumption has been made that total iron was present as Fe^{2+}. On this basis a K_D of 0.37 for the Fo_{92}-A_{2r} 'liquid' is obtained which is a minimum value. If the A_{2r} 'liquid' contained a significant Fe^{3+} content the K_D value would be higher still. Compared to the experimentally determined value (∿0.30) disequilibrium is indicated for the Fo_{92}-A_{2r} 'liquid' pair. However other estimates of K_D between olivine and

ultramafic komatiite liquids (minimum values as total iron assumed to be FeO) suggests slightly higher values than for basaltic systems. The experimental data of Bickle et al. (1977) for olivine crystallizing from ultramafic liquids shows a range of K_D values from 0.34 to 0.37. In addition the olivine from the spinifex textured sample 49-J (Green et al. 1975) yields a K_D of 0.39 and Nisbet et al. (1977) have estimated K_D for the Rhodesian ultramafic komatiites to be as high as 0.43. In view of the higher K_D values indicated by these studies in ultramafic liquid systems, equilibrium for the Fo_{92}-A_{2r} 'liquid' from Stuart's flow cannot be discounted using the experimentally determined K_D by Roeder and Emslie (1970) and Duke (1976).

D_{Ni} (wt.% Ni in olivine/wt.% Ni in liquid) has been measured for the same olivine-'liquid' pair in Stuart's flow and is given in Table 3-5 together with the D_{Ni} values calculated, using the equations of Hart and Davis (1978) and Arndt (1977b). The agreement between the two calculated results is good but significantly higher than the measured value.

Hart and Davis (1978) have computed the variation of D_{Ni} as a function of 1/MgO for a wide range of compositions (MgO from 4.8 - 24.7) and liquid structures, as reflected by the Si/O (atomic) ratio (0.27 - 0.37), in iron-free olivine-liquid systems. The higher MgO content of the A_{2r} 'liquid' coupled with the fact that the curve fitted to the experimental data of Hart and Davis (1978) falls above the data points at high MgO, may partly account for the discrepancy between the calculated and measured D_{Ni} in Stuart's flow. Arndt (1977b) on the other hand has used komatiitic compositions for his experimental investigation and computed the variation of D_{Ni} as a function of the normative olivine content of the liquids. The A_{2r} 'liquid' from Stuart's flow again has higher MgO content than the range of liquid compositions

given in Arndt (1977b), although both the A_{2r} 'liquid' and the liquid compositions used by Arndt have similar Si/O (atomic) ratios of around 0.3. The D_{Ni} and MgO data reported in Arndt (1977b) have been replotted as D_{Ni} versus 1/MgO (Fig. 3-3). The observed D_{Ni} and MgO data for the spinifex textured sample 49-J (Green et al., 1975) have also been included for comparison. The correlation between D_{Ni} and MgO for this limited number of data points is good (correlation coefficient = 0.99) over a wide range of komatiite liquid compositions and a linear regression line fitted to all the data points in Fig. 3-3 gives the following equation:-

$$D_{Ni} = \frac{111.33}{MgO} - 1.71 \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (3\text{-}2)$$

The predicted D_{Ni} for the A_{2r} 'liquid' from Stuart's flow using this equation is in good agreement with the measured value (see Table 3-5). The expected D_{Ni} for another ultramafic spinifex textured sample SF-134 (Nisbet et al., 1977) calculated using equation 3.2 (D_{Ni} = 2.2), agrees well with the measured value of 2.3 (assuming the most Ni rich olivine analysed in sample SF-134 is in equilibrium with the bulk rock composition). Although there are insufficient data in the literature to fully test the applicability of equation 3-2 to komatiite rocks, the little there is confirms its usefulness and on this basis it is concluded that the Fo_{92}-A_{2r} pair from Stuart's flow was in equilibrium.

The high Cr content of olivines from ultramafic komatiites compared to other terrestrial olivines has previously been noted (Green et al., 1975; Dickey, 1972; Nisbet et al., 1977). Some lunar olivines also have high Cr contents but Green et al. (1975) have pointed out that at the oxygen fugacities at which these olivines crystallized, all Ni existed in the metal phase, which is clearly not the case in the Archaean ultramafic flows. Computing D_{Cr} for the Fo_{92}-A_{2r} sample

gives a value of 0.5. Recent determinations of D_{Cr} (Leeman and Scheidegger, 1977; Duke, 1976) for basaltic rocks yield values that are greater than unity. As the D_{Cr} determined for Stuart's flow is out of the compositional range of these studies, direct comparisons cannot be made. Duke (1976) has noted that increasing the temperature of crystallization should lead to a decrease in D_{Cr} value. Using the equation developed by French (1971), the one atmosphere liquidus temperature of the A_{2r} sample is estimated to be $\sim 1550^{\circ}C$. It is therefore concluded that the D_{Cr} of 0.5 obtained for Stuart's flow is consistent with the expected high crystallization temperature of the olivines and that their Cr content is due to the very high Cr content in the liquid from which they crystallized.

3-5 COOLING AND CRYSTALLIZATION MODEL FOR STUART'S FLOW

One of the most striking features of Stuart's flow and other similar ultramafic flows is the abrupt change of the olivine morphologies occurring within the flow. Green et al. (1975) have duplicated the A and B zones on a microscale and were able to match the olivine compositions in the experimental charges to a natural sample (49-J). Donaldson (1976) has stated that spinifex textures in Archaean rocks are not due to rapid cooling, but due to rapid olivine growth caused by the high normative olivine content of the magma. In view of the high temperatures of extrusion of ultramafic magmas and the thin nature of the flows it would be unrealistic to construct models whereby rapid cooling was not operative at least for the outer margins of the flows. Using the experimental data of Donaldson (1976, 1977) to account for the observed olivine morphologies, a two stage cooling model is proposed for Stuart's flow.

Stage 1

Rapid chilling of the flow margins after extrusion, forming the basal chilled contact and an upper contact of random spinifex (A_{2r} zone 40 cms. thick). As the olivine blades crystallized in the A_{2r} zone the composition of the trapped inter-blade liquid evolved to lower MgO contents and coupled with slower cooling rates (as the temperature of the chilled margin decreased), hopper olivines and skeletal pyroxenes were able to crystallize in the inter-blade area. Solidification of the upper chilled zone must have occurred sufficiently fast to entrap gas bubbles.

Stage 2

This stage was characterized by much slower cooling rates than the first stage, and commenced when the flow of magma between the contacts was either very slow or had ceased altogether. Olivine phenocrysts in the magma settled rapidly leaving a phenocryst free zone below the upper chilled portion of the flow. Arndt et al. (1977) have estimated that an olivine grain 0.5 mm. in diameter would settle with a terminal velocity of 40 cm. per hour in ultramafic magmas. Donaldson (1977) and Pyke et al. (1973) have noted that olivine blades initially nucleate on solid objects and that a nuclei free zone is required for the development of spinifex texture. It is proposed in this model that the plate spinifex nucleated on the base of the random spinifex zone and grew downwards into the magma along the thermal gradient forming the A_{2p} spinifex textured zone. Drever and Johnston (1972) have described crystal textures that have developed

along thermal gradients as 'thermotactic'. It is suggested that the compositional changes (28% to 25.7% MgO) from top to bottom in the A_{2p} zone reflect the changes in the evolving residual liquid caused by the crystallization and settling of olivine to the developing B zone. Stage 2 terminated once the flow had cooled below its solidus.

In the cooling model proposed, the random spinifex (A_{2r}) has cooled at a faster rate than the plate spinifex (A_{2p}) zone. As the olivine morphology observed in any portion of the flow will depend on the interplay of (inter alia) the MgO content, viscosity, temperature, cooling rate and the density of olivine nuclei present in the magma, it would seem premature at present to attempt to constrain the observed morphologies in the flow to a particular linear cooling rate or degree of supersaturation from Donaldson's experimental work. The contribution of pyroxene to the differentiation of the flow appears to be restricted to the crystallization of this mineral in the olivine inter-blade areas of the A_2 zone.

Finally in conclusion the following points arising from this study of an ultramafic komatiite flow are reiterated:-

1) Not all spinifex textured samples from ultramafic rocks are representative of the initial liquid composition, and it is suggested that random spinifex textured samples give the best compositional estimate of the initial liquid (phenocryst free magma).

2) The compositions of plate spinifex textured samples studied have been greatly modified by olivine fractionation that occurred in the flows after extrusion of the magma.

3) The composition of the B zone of spinifex textured ultramafic flows is consistent with their cumulate texture and indicates that variable amount of olivine settled into the lower portions of such flows.

MAFIC KOMATIITE LAVA FLOWS

3-6 INTRODUCTION

Mafic komatiite lava flows occurring as pillowed horizons or as massive flows make up a major portion of the Komati Formation. In stream sections, where outcrop is well exposed, detailed relationships between the flows can be seen. The flows vary in thickness from 1 - 10 m and individual flow units can often be identified with distinct flow top breccias and basal chilled margins. Several massive flows were sampled in order to examine chemical variations across the flows. The data for one mafic komatiite flow are reported below.

3-7 FIELD RELATIONS AND PETROGRAPHY

The section of one flow that was sampled in detail occurs in the stream bed immediately west of the type section established for the Komati Formation by Viljoen and Viljoen (1969c). This flow is 7.3 m thick and has a flow top breccia (1.7 m thick) and a thin (0.1 m thick) basal contact. This flow is hereafter referred to as

Tony's flow. The main features of the flow are shown in Fig. 3 - 4. The flow top breccia consists of coarse and usually angular fragments (several cms across) which are now cemented by quartz. The basal chilled margin and the flow top breccia are much finer grained than the material in the central portion of the flow. Small vesicles (0.1 to 2 mms across) now filled with quartz occur in the central portion of the flow and in the basal chilled margin. No clinopyroxene spinifex textured zones were observed in Tony's flow. The flow can be traced for several metres along strike before it disappears under a talus covering.

In thin section the flow top breccia is fine grained and contains amphibole needles up to 1 mm long set in a groundmass of amphibole, chlorite and minor opaque material. The amphibole needles probably represent altered clinopyroxene needles. No phenocrysts were observed in this sample. In contrast the basal chilled margin, which is also fine grained, contains scattered euhedral microphenocrysts that have been pseudomorphed by quartz, set in a groundmass of amphibole, chlorite, quartz and minor spinel. From the morphology of the microphenocrysts they are inferred to have been olivine.

The two samples immediately below the flow top breccia are fine grained and have clinopyroxene crystals (now altered to amphibole) set in a fine grained groundmass of amphibole, chlorite, quartz and minor spinel. The two samples above the basal contact contain abundant euhedral clinopyroxene phenocrysts (now altered to amphibole) set in a groundmass of amphibole, chlorite quartz and minor spinel. The sample SC-6C also contains some serpentine. From the above inferences of the original igneous mineralogy of the flow, clinopyroxene appears to have been the major crystallising phase accompanied by lesser amounts of olivine and spinel.

3-8 GEOCHEMISTRY OF TONY'S FLOW

Six samples have been analysed from Tony's flow (sampling points are shown on Fig. 3-4) and the major element and normative mineral compositions are given in Table 3-6a and trace element and inter-element ratios are given in Table 3-6b. The major element and trace element data have been calculated volatile free and the FeO content calculated such that the Fe_2O_3/FeO ratio = 0.2. As in the previous section on the geochemistry of Stuart's flow Na, K, Rb, Sr, Ba, Cu, Zn and S are excluded from the following discussion as their concentrations may have been disturbed by secondary processes.

The most striking feature of the chemistry of Tony's flow is the large compositional variation from 12 to 19% MgO (see Table 3-6a and Fig. 3-4). The range of sample compositions from Tony's flow essentially encompasses all three basaltic komatiite types originally defined by Viljoen and Viljoen (1969c). The compositional variations in the flow were probably brought about by in situ crystal fractionation and accumulation processes (see later discussion). These variations illustrate the importance of detailed textural studies of lavas to identify possible cumulate samples before any inferences can be made as to whether a particular rock is representative of a magma composition or not.

In view of the significant (but contentious) role CaO/Al_2O_3 ratios have played in the characterisation of komatiites (see Chapter 4) it is clearly important to establish if high CaO/Al_2O_3 ratios are due to clinopyroxene accumulation in individual samples or not. The weighted average composition of all the samples in Tony's flow (Tables 3-6a and 3-6b) is essentially identical to the phenocryst free flow top breccia sample SC-6A from which it is assumed that the initial magma composition forming the flow was

similar to SC-6A in composition. If this is correct then the initial magma had a CaO/Al_2O_3 ratio of 1.7 and the observed range of ratios in the flow (1.3 to 2.5) was probably brought about by the fractionation and accumulation of calcium and or aluminum bearing phases. The sample HSS-C2 has a CaO/Al_2O_3 ratio (2.5) which is significantly higher and clearly not representative of the initial magma and from thin section examination this sample is believed to have accumulated clinopyroxene.

Variation of the concentrations of selected elements with depth in Tony's flow are illustrated in Fig. 3-4. MgO is higher and Al_2O_3, TiO_2, Zr and V lower in the lower three samples than the top three samples, suggesting that ferromagnesium minerals have accumulated towards the base of the flow. The variations of CaO, Cr, Ni and Sc with depth in the flow are erratic, while variations in the CaO/Al_2O_3 ratio (see Table 3-6b) show that the two top and bottom samples have similar ratios of $\sim$1.7. The sample HSS-C2 has the highest CaO/Al_2O_3 ratio and the sample above it (SC-6B) the lowest, suggesting that clinopyroxene fractionation and accumulation were most important in the central zone of the flow. The similar CaO/Al_2O_3 ratios of the lower samples (SC-6C and HSS-C3) but higher MgO and Ni contents compared to the flow top breccia, suggest that olivine or perhaps orthopyroxene had accumulated in this zone of the flow. Inter-element ratios such as Al/Ti, Ti/Zr and Ti/V show small or insignificant changes with depth in the flow indicating that an Al bearing phase (such as plagioclase) did not fractionate as the flow cooled and crystallised.

CRYSTALLISATION AND ACCUMULATION PROCESSES OCCURRING IN TONY'S FLOW

Attempts to quantify the absolute amounts of accumulation or crystal fractionation that occurred in Tony's flow are hampered by the lack of mineral compositions as no relict igneous minerals were observed. From thin section examination of the samples it has been deduced that clinopyroxene and olivine were probably important fractionating phases. The geochemical data are compatible with clinopyroxene, olivine and or orthopyroxene as likely fractionating minerals. While no igneous mineral compositions have been determined from Tony's flow, many analyses of phenocrysts from other mafic and ultramafic komatiite samples are available (given in Chapter 5 and 6). As will be shown in Chapter 6 clinopyroxene has a restricted range of compositions regardless of whether this mineral was analysed from ultramafic komatiites or mafic komatiites. The composition of the clinopyroxenes crystallising in Tony's flow were probably similar to the endiopside compositions given in Table 6-1 and for the purposes of the modelling calculations below, a clinopyroxene composition (atomic) Mg/(Mg+Fe) = 0.86 was chosen. An olivine of composition Fo_{87} was selected for the modelling calculations as this gives a K_D (defined earlier) = 0.32 between olivine and assumed initial liquid (SC-6A) which is of the expected order of magnetude (i.e. Roeder and Emslie, 1970).

One further problem with the samples from Tony's flow is the presence of secondary quartz either as thin veins (mainly in the flow top breccia) or as quartz filled vesicles in the central and basal chilled margin samples. This quartz, which is clearly of secondary origin, was not completely eliminated from all the samples during

sample preparation. Quartz was therefore added as an additional mineral to the modelling calculations, although it is emphasised that quartz is not considered to have been a fractionating phase during the cooling and crystallisation of Tony's flow.

The least squares linear mixing method (Bryan et al., 1969) was used to calculate the proportions of initial magma (assumed to be similar in composition to the flow top breccia sample SC-6A), clinopyroxene, olivine and quartz required to generate the observed compositions in Tony's flow. The results of the mixing calculations are given in Tables 3-7a to 3-7e. Note that total iron has been calculated as FeO; K_2O, Na_2O and P_2O_5 have not been included in the calculations. The agreement between the predicted and observed major element compositions as given by $\Sigma(\text{calc.-obs.})^2$ is considered satisfactory. The inclusion of orthopyroxene in the calculations was found to be unnecessary as its fractionation was only required in very small amounts and the residuals were not significantly improved.

The calculation of trace element checks is complicated by several factors including the problem of variable amounts of included quartz in the different samples, and accumulation of minerals into the lower portion of the flow section (i.e. F must be estimated using methods such as those described by Le Roex and Reid, 1978). It is suspected that the flow top breccia sample SC-6A also contained some secondary quartz and therefore the absolute trace element contents in this sample may be too low by an unknown amount but probably less than 5%.

Only one trace element check has been attempted in Tony's flow. This is for the central sample SC-6B where the proportion of minerals removed from the initial magma to get SC-6B composition (as estimated

from the major element modelling) is relatively large ∿20%. Quartz was treated as a fractionating mineral with $D^{qtz/liq} = 0.0$ for the trace elements considered in Table 3-7b. The $D^{mineral/liq}$ values used for clinopyroxene and olivine are given in Table 6-8 and the fractional crystallisation equations given in Arth (1976) were used. The general agreement between the observed and calculated trace element contents are considered acceptable in view of the problems of quartz contamination. Thus, it is considered that the trace element data supports the major element modelling, which indicates that all the compositions in Tony's flow can be related to the initial magma composition by addition or removal of clinopyroxene and olivine.

In detail, the modelling suggests that Tony's flow differentiated in two stages. The first stage consisted of crystallisation and settling of olivine accompanied by lesser amounts of clinopyroxene. These minerals accumulated in the lowest portions of the flow. The second stage was crystallisation and settling of clinopyroxene in the central portion of the flow.

In general these variations are compatible with what can be inferred about the primary igneous mineralogy of the flow. Although the section of Tony's flow sampled is 7.3 m thick, it probably cooled relatively rapidly as it is closely associated with pillow lavas. It was nevertheless able to differentiate into a significant range of compositions by in situ crystallisation and crystal settling.

The most significant conclusions arising from this study of Tony's flow are summarised below:-

1) Clinopyroxene and olivine were the major fractionating silicate phases in the flow.

2) In situ differentiation has developed samples from this flow section with a wide range of compositions from 12 to 19% MgO.

3) Clinopyroxene accumulation in certain sections of the flow has increased the CaO/Al_2O_3 ratio from 1.7 in the presumed initial magma to 2.5 in the sample HSS-C2.

CHAPTER 4.

KOMATIITE NOMENCLATURE

4-1 INTRODUCTION

In 1969 at the Upper Mantle Project Symposium held in Pretoria (South Africa) M.J. Viljoen and R.P. Viljoen presented a series of papers (Viljoen and Viljoen, 1969a-j) on the geology and geochemistry of the Barberton Mountain Land which have proved to be of great significance with respect to igneous rock nomenclature and classification. They identified a hitherto unknown suite of volcanic rocks for which they coined the name Komatiite after the Komati river flowing through the Barberton area. They subdivided the suite into basaltic and peridotitic komatiites with a further subdivision of the basaltic komatiites into Barberton, Badplaas and Geluk types. As all the greenstone belt rocks from this area have suffered varying degrees of metamorphism from amphibolite to greenschist facies, the Viljoens consequently stressed chemical features rather than mineralogy for classifying the rock types. One important aspect of the Viljoens' work is that they listed distinctive features rather than strict chemical limits for each of the groups.

Since 1969 both basaltic and peridotitic komatiites have been identified from many other Archaean greenstone belts while only basaltic komatiites have, so far, been identified from younger tectonic environments. The geochemical characteristics outlined by Viljoen and Viljoen (1969c,d,e) for the komatiite rock types have been extensively modified or often totally rejected by other workers, to suit the particular characteristics of the rocks under investigation. In several cases new names have been proposed for what the present author considers to be broadly equivalent rock types. This has resulted in considerable confusion in the literature as to just what komatiites really are. One

positive aspect that has come out of the confusion is that other parameters have also been shown to be of importance for the characterisation of komatiite lavas such as rock type associations, structures and textures. Obviously a reliable set of characterising parameters for the komatiite suite of lavas must be at hand from Archaean volcanic localities before similar rock types can be identified in younger tectonic environments. As pointed out by Brooks and Hart (1974) classification schemes based on geochemical parameters alone are unsatisfactory as many types of partial cumulate or cumulate rocks may be erroneously classified as komatiites. They advocate inclusion of textural evidence that indicates the rocks are representative of magma compositions. One important consequence of including textural criteria is that unless there is some other compelling evidence, rocks where the igneous textures have been destroyed by metamorphism cannot be unambiguously classified as komatiites.

The whole problem of komatiite nomenclature is at present under investigation. R.A. Binns (then at the University of Western Australia) circulated a questionnaire after the 25th International Geological Congress to earth scientists working on komatiites and greenstone belts, to solicit opinions on the nomenclature problem. A general summary of the 27 replies to the questionnaire was circulated in November 1977. N.T. Arndt (University of Saskatchewan) has taken over the 'Komatiite Nomenclature Project' and compiled and circulated a second questionnaire. Some suggestions made in the replies to these questionnaires have been incorporated in a new definition of komatiite (Arndt, 1980) still to be published.

While it is perhaps premature at the present time to add another komatiite definition to the expanding literature on this topic, it is pertinent to re-examine the Viljoens' classification scheme in the

light of additional data from their type locality. In this chapter some of the more relevant criticism levelled at the Viljoens' characterisation scheme and later efforts to define komatiites are reviewed. The nomenclature used in this work is outlined and differs from that proposed by the Viljoens, but for the present, is only considered to be applicable to the komatiites from the Barberton greenstone belt.

4-2 SOUTH AFRICA

Viljoen and Viljoen (1969c,d) noted three types of basaltic komatiite and two types of peridotitic komatiite from the Barberton greenstone belt. Below is a brief summary of the distinguishing characteristics of each of the rock types as used by Viljoen and Viljoen (1969c,d). The meta-tholeiites will not be discussed here as they are a well established rock class found in most greenstone belts.

BASALTIC KOMATIITES

a) Barberton Type

These have relatively high SiO_2 and alkali contents with the most distinctive feature of this komatiite type being roughly equivalent amounts of CaO, Al_2O_3 and MgO ($\sim$10% each). CaO/Al_2O_3 ratios are within the range 0.8 to 1.2. The Viljoens felt there was a need for more data before this type of lava could be comprehensively characterised.

b) Badplaas Type

These are characterized by high SiO_2, MgO and CaO contents, and low Al_2O_3 and alkali contents. This type of lava has the highest CaO/Al_2O_3 ratios (2-3) of all the

komatiites.

c) Geluk Type

These contain relatively high SiO_2 and very high MgO contents, coupled with very low alkali contents. The contents of CaO and Al_2O_3 are also low and the Al_2O_3 content of this type of komatiite was used to further sub-divide the Geluk type:-

High Al variety with CaO/Al_2O_3 ratios from 1.1 to 1.4

Low Al variety with CaO/Al_2O_3 ratios from 2.4 to 2.6

The low alkali, TiO_2 and FeO contents of the basaltic komatiites were used to distinguish these lavas from other rock types such as ankaramites that show a similar range of CaO/Al_2O_3 ratios and MgO contents.

PERIDOTITIC KOMATIITES

a) Komati and Theespruit Formation peridotitic komatiites.

These lavas have very high MgO contents and high CaO/Al_2O_3 ratios (1.1 - 2.3), relatively low amounts of Al_2O_3 and CaO and very low alkali contents.

b) Sandspruit Formation peridotitic komatiites.

These rocks have higher SiO_2, CaO, FeO* (total Fe as FeO), alkali content, CaO/Al_2O_3 ratios (1.4 - 2.6) and lower MgO contents than the Komati Formation peridotitic komatiites.

The CaO/Al_2O_3 and MgO/FeO ratios were used to distinguish the peridotitic komatiites from other ultramafic rocks of similar MgO content.

The Viljoens' characterisation scheme contains a number of problems such as the limited number of samples used to characterise some of the komatiite types, the rejection of certain analyses on the basis of anomalous oxide contents which may or may not have been due to alteration and the inclusion of rocks which are now considered to be partial cumulates into some of the groups. A further problem in the Viljoens' work is that samples from the Hooggenoeg and Kromberg Formations (e.g. LV-6, M-57 and 526; Viljoen and Viljoen, 1969e) are very similar in composition to some of the basaltic komatiite types, but are termed Mg-rich metabasalts or basalts. This places a stratigraphic (time) restriction on the komatiite occurrence which was probably unintended in the original work.

Rocks of komatiite composition also occur in the less well preserved Murchison Range north of the Barberton greenstone belt (Minnitt, 1975). Anhaeusser (1977) has recorded rocks of peridotitic and basaltic komatiite composition from the Roodekrans Ultramafic Complex and notes continuous chemical gradation from ultramafic komatiites through to rocks of tholeiitic composition.

4-3 WESTERN AUSTRALIA

Nesbitt (1971) has analysed numerous olivine spinifex textured ultramafic rocks from the Yilgarn block and notes the similarity between these ultramafics and those from the Barberton greenstone belt, and in particular the increasing CaO/Al_2O_3 ratio with decreasing MgO content. However this trend is more marked in the Barberton komatiites and Nesbitt (1971) proposed the name 'Archaean greenstone peridotites' as an alternative to the high CaO/Al_2O_3 Barberton 'peridotitic komatiites'. Williams (1972) also notes the lower CaO/Al_2O_3 ratios of the

Mt. Monger ultramafic lenses compared to the Barberton peridotitic komatiites. He further considers the term basaltic komatiite to be unacceptable for mafic rocks in the range 8-24% MgO from the Mt. Monger area. The basis of Williams's rejection of the term basaltic komatiite is due to the limited number of analyses for some of the groups in the original definition, the anomalous chemistry of some komatiite samples and the lack of consideration of textural criteria by Viljoen and Viljoen (1969c). In addition the high CaO/Al_2O_3 ratios are not found in the Western Australian lavas and Williams (1971) therefore proposed the term 'high-Mg basalt' for lavas with MgO in the range 8-24%. Williams (1972) and Hallberg and Williams (1972) stressed the importance of the unusual textures (olivine and clinopyroxene spinifex texture) as one of the identifying parameters for high-Mg basalts from Archaean greenstone belts.

McCall (1973), in a rather confusing criticism of komatiite classification, on the other hand appears to consider high Mg volcanic rocks as 'magnesia-rich associates of the abyssal tholeiites' although he notes that magnesia-rich and tholeiitic suites may form two quite distinct genetic suites.

Nesbitt and Sun (1976) have suggested that the high CaO/Al_2O_3 ratios occurring in the Barberton greenstone belt komatiites may be due to loss of Al_2O_3 during metamorphism and that the original 'definition' should be modified such that komatiitic lavas with CaO/Al_2O_3 ratios close to unity be considered as primary. They further note that high MgO lavas (>20%) from many greenstone belts display quench textures with skeletal olivine and/or pyroxene (spinifex textures) commonly set in a quench matrix. Nesbitt and Sun (1976) termed these high-MgO quench-textured rocks 'spinifex-textured peridotites (STP)

and suggest they belong to the komatiite suite despite many of the samples having low CaO/Al_2O_3 ratios. The whole suite of komatiite compositions as originally selected by Viljoen and Viljoen (1969c,d) show a range of CaO/Al_2O_3 ratios from 0.8 to 2.9 with several ultramafic rocks having CaO/Al_2O_3 ratios close to unity. Therefore the STP rocks discussed by Nesbitt and Sun (1976) should be considered as komatiites as their CaO/Al_2O_3 ratios are within the range recorded for Barberton komatiites. In a later paper (Sun and Nesbitt, 1978) the high CaO/Al_2O_3 ratios from the Barberton ultramafic rocks are considered to represent an original igneous feature and a more complex terminology was introduced. This terminology however will not be discussed here as it has been updated in a more recent paper (Nesbitt et al., 1979). Nesbitt et al. (1979) term peridotitic komatiites with CaO/Al_2O_3 ratios close to unity, AUPK (aluminium-undepleted peridotitic komatiites) while those with high CaO/Al_2O_3 ratios, ADPK (aluminium-depleted peridotitic komatiites). However Nesbitt et al. (1979) still consider the development of spinifex textures to be of prime importance as this texture indicates the development of high magnesian liquids. These authors have emphasised that the geochemical subdivision of the basaltic komatiites from Barberton are of little practical use without further characterisation of the textures and mineralogy of this class of rocks.

Naldrett and Turner (1977) have modified the komatiite classification scheme proposed by Arndt et al. (1977) (see later section) for their classification of komatiite rocks from the strongly metamorphosed Yakabindie greenstone belt. They have used the following criteria:-

Peridotitic komatiites	>30% MgO
Pyroxene peridotites	20-30% MgO
Pyroxenites	20-12% MgO
Magnesian basalts	10-12% MgO
Basalts	<10% MgO

However this scheme is of little general use as the primary textures in the rocks from this area have largely been destroyed.

4-4 RHODESIA

Bickle et al. (1975) and Nisbet et al. (1977) have studied the geochemistry of komatiites from the Belingwe greenstone belt and have grouped lavas into three categories, namely rocks essentially free of phenocryst phases, rocks enriched in phenocrysts and rocks that have undergone high level fractionation. The rocks rich in phenocrysts and the cumulate samples were excluded from the definition used by Nisbet et al. (1977) which is essentially that proposed by Brooks and Hart (1974) (see later) except that samples with CaO/Al_2O_3 as low as 0.8 were included as komatiites. Nisbet et al. (1977) also note that spinifex textures have only been observed in rocks with more than 9% MgO and suggest that spinifex textures can be used as a guide in the field for magnesium rich liquids.

4-5 CANADA

Pyke et al. (1973), in their classical description of Archaean ultramafic lava flows in the Munro Township area, note that these lavas do not have the high CaO/Al_2O_3 ratios common to the Barberton peridotitic komatiites. Arndt et al. (1977) after considering the Viljoen and Viljoen (1969c,d) descriptions of komatiites and the Brooks and

Hart (1974) definition, consider both schemes to be inadequate for describing the Munro Township lavas, as over half of the komatiites analysed from this area have CaO/Al_2O_3 ratios less than 1. They proposed instead a three fold classification scheme: peridotitic komatiites (>20% MgO), pyroxenitic komatiites (12-20% MgO) and basaltic komatiites (<12% MgO) based on the field characteristics (unusual textures and volcanic structures such as spinifex textures and polyhedral jointing) and geochemical parameters. The geochemical parameters include low $FeO^*/(FeO^*+MgO)$ ratios and TiO_2 contents and high MgO, NiO and Cr_2O_3 contents with some samples having high CaO/Al_2O_3 ratios. Arndt et al. (1977) used a plot of Al_2O_3 against $FeO^*/(FeO^*+MgO)$ to distinguish komatiites from tholeiitic lavas. The komatiite rocks have high Al_2O_3 contents at given $FeO^*/(FeO^*+MgO)$ ratios compared to the tholeiites. However the type locality Barberton komatiites plot within the tholeiitic field on their diagram and it is therefore considered totally unsuitable as a classifying parameter even though Arndt et al. (1977) note that the Barberton komatiites can be differentiated from tholeiites on the basis of their CaO/Al_2O_3 ratios. Goldich and Wooden (1978) have proposed that the high magnesium lavas from the Munro Township area be called Munroites to circumvent the definition problems.

4-6 RUSSIA

The high MgO rocks (Meimechites) occurring in the Gulin Ultramafic intrusion in northern Siberia show some features similar to the ultramafic komatiites from the Barberton greenstone belt. These rocks have high CaO/Al_2O_3 ratios and olivine as the predominant phenocryst phase. The olivines have high CaO and Cr_2O_3 contents (Sobolev, 1978) and have been shown to be in equilibrium with the groundmass composition.

Sobolev (1978) considers that the bulk rock composition approximates a melt composition which has not been enriched in MgO by olivine phenocryst accumulation.

An olivine rich sample from a Meimechite dyke was donated by A.V. Sobolev to J.J. Gurney of this Department and was analysed in this work, for major and trace elements, for comparison with ultramafic komatiites from the Barberton greenstone belt. The analysis of this sample is given in Table 4-1 along with an analysis of a phenocryst free komatiite from the LUU of similar MgO content and an ultramafic rock from Upton and Thomas (1973). The meimechite has much higher contents of TiO_2, CaO, K_2O, P_2O_5, Nb, Zr, Y, Sr, Rb, V, Ba and higher Ti/Zr and Ti/V ratios than the komatiite, and lower contents of SiO_2 and Al_2O_3 and a lower Al/Ti ratio. There is little doubt that the meimechites are not komatiites but they do show similar major and trace element compositions to the potassic ultramafic rocks (see Table 4-1) from Greenland analysed by Upton and Thomas (1973).

4-7 BROOKS AND HART (1974)

Brooks and Hart (1974) searched computerized data files and analyses reported in the literature for rocks of komatiite composition, using the criteria of Viljoen and Viljoen (1969c,d). They found numerous rock types that correspond to ultramafic komatiite compositions, particularly those ultramafic rocks that contained significant amounts of clinopyroxene or calcic amphibole. They therefore pointed out that evidence such as quench textures or pillowed structures indicating that the peridotitic komatiite was once a silicate melt, is essential to the definition of peridotitic komatiites.

On the other hand very few samples corresponded to basaltic

komatiite compositions once restrictions had been placed on the upper limit of acceptable TiO_2 and K_2O contents. They selected the following criteria for distinguishing basaltic komatiites:-

SiO_2	46-53%
TiO_2	<0.9%
K_2O	<0.9%
MgO	>9.0%
CaO/Al_2O_3	>1

They therefore abandoned the 3 fold subdivision of the basaltic komatiites proposed by Viljoen and Viljoen (1969c) and eliminated some komatiites from the type area as being komatiites. Brooks and Hart (1974) further noted that there may be a continuum of compositions between peridotitic komatiites and basaltic komatiites but did not offer any limiting parameters to sub-divide the two komatiite types.

Although there are many other localities from which komatiite rock types have been described, these rocks have usually been identified on the basis of the criteria outlined by Viljoen and Viljoen (1969c,d) or Brooks and Hart (1974) and more recently by Arndt et al. (1977).

4-8 KOMATIITE TYPES FROM THE LOWER ULTRAMAFIC UNIT

Although the preceding sections outline some of the confusion surrounding komatiite nomenclature there is general agreement on several critical points:-

a) All the authors agree that the group of high magnesium lavas found in Archaean greenstone belts are distinct

from tholeiitic lavas, except McCall (1973) who considers that they may be an extension of the tholeiitic suite. On the other hand several authors have identified what they consider to be high magnesium tholeiites (e.g. Arndt et al., 1977; Naldret and Turner, 1977) and consider them to be geochemically distinct from komatiites.

b) The term komatiite first suggested for this group of high MgO lavas has been widely accepted in the literature.

c) Textures, especially olivine and clinopyroxene spinifex, are considered important either as a characteristic of komatiites or to indicate that liquids existed with a composition similar to that of the rock. Only in one instance (Jahn and Sun, 1977) have textures been considered to be of secondary importance. However rocks from greenstone belts where the textures (igneous) have been destroyed by metamorphism such as in the Yakabindie area (Naldret and Turner, 1977) and the Finnish greenstone belts (Jahn et al., 1978)pose a problem in this respect. Nevertheless the classification of monomineralic rocks as komatiites, such as the hornblendite from Bon Accord, South Africa (McIver, 1972) illustrates the importance of including textural criteria in the definition of komatiites.

d) The data from most greenstone belts (including Barberton) indicate that not all komatiites have CaO/Al_2O_3 ratios

greater than unity. Although the Barberton komatiites often have ratios greater than 1, other greenstone belt komatiites have ratios which tend to cluster around 1, and as such ratios <1 are common. In contrast to the agreement in the literature on the above points, there is little agreement on what precise geochemical parameters should be used to define komatiitic rocks. Variation in the geochemistry of komatiites from different greenstone belts will be considered in more detail in Chapters 5 and 6; here only those from the Barberton greenstone belt will be considered.

DEFINING PARAMETERS

The komatiites from the Lower Ultramafic Unit have in this work been sub-divided into three types on the basis of texture, mineralogy and geochemistry. As the Onverwacht Group volcanic pile has been subjected to greenschist facies metamorphism, many of the primary minerals have been pseudomorphed by greenschist facies mineral assemblages. Identification of the original igneous minerals from the pseudomorphed grains has been facilitated by the preservation of relict igneous minerals in many of the lavas. It should be noted that only the inferred or observed igneous mineralogy of the lavas is discussed in this section. The geochemical and mineralogical features are specific to each group and will be discussed in the relevant sections below.

The textural features on the other hand are common to all the groups and have been used here in the sense of establishing whether

a particular rock represents an essentially phenocryst free magma composition or if it could have been enriched in phenocrysts. Pillows, chilled margins, olivine spinifex textured samples and aphanitic aphyric lavas have been selected as representative of the phenocryst free magma compositions. (The terms aphanitic and aphyric are used here for very fine grained rocks with $\sim$<2% phenocrysts or microphenocrysts). Only these samples have been used for establishing the defining geochemical parameters. It has been shown in Chapter 3 that relatively large degrees of crystal accumulation can occur in thin ultramafic and mafic komatiite flows and it is predicted in Chapter 5 that komatiite magmas had relatively low viscosities. This enables even small phenocrysts to settle through considerable distances in the melts in relatively short times. Bearing these points in mind it is considered justifiable that coarse grained rocks and phenocryst rich samples be eliminated for defining the rock types. It is realised that in certain instances porphyritic rocks may represent liquid compositions (e.g. magmas that have undergone processes of compensated crystal settling as outlined by Krishnamurthy and Cox, 1977) despite their porphyritic nature. However as these samples are difficult to distinguish texturally from lavas that have been enriched in phenocrysts by accumulation processes, all porphyritic samples have been eliminated from this discussion.

The particular characteristics for each komatiite type are given below:-

a) Ultramafic Komatiites

All komatiites with >24% MgO contain abundant olivine either as spinifex textured blades, microphenocrysts or in the groundmass, while those with less than 24% MgO

contain very minor contents or olivine is completely absent. The lavas with >24% MgO also have ~90% normative mafic mineral contents and therefore conform to the term ultramafic, but not ultrabasic as SiO_2 contents >45% are common. The rocks with greater than 24% MgO are therefore termed ultramafic komatiites and major and trace element compositions of this group of lavas are given in Tables 5-3a and 5-3b. The term ultramafic is preferred to peridotitic as the connotations of the latter term are avoided. The division between the Komati Formation and Sandspruit Formation ultramafic komatiites made by Viljoen and Viljoen (1969c) is no longer considered to be valid as it is shown in Chapter 5 that these two groups have essentially identical compositional ranges.

b) High-Mg Mafic Komatiites

Lavas with compositions in the range 19-24% MgO have clinopyroxene and possibly orthopyroxene as the main phenocryst phases. Olivine has not been observed in these lavas but the minor serpentine content present in some of the rocks may represent altered olivine. In fact no fresh orthopyroxene has been observed either, but it is inferred to have been present from the alteration styles of the pyroxene phenocrysts as outlined by Williams (1971). Plagioclase if present is a late-crystallising phase and very minor in content. These lavas are quartz or olivine normative (olivine less than 15 wt.%). The gap in MgO contents of the komatiitic lavas between 16 and 19% MgO is illustrated in

Fig. 4-1. As the density of sample points on either side of this gap is low,this may represent inadequate sampling. However a similar gap at roughly the same MgO contents has been noted in the komatiitic lavas from Belingwe (Nisbet et al., 1977) and from Munro Township (Arndt et al., 1977). The lack of lava compositions that represent liquids between 16 and 19% MgO provides a convenient lower limit of 19% MgO for the high-Mg mafic komatiites and an upper limit of 16% MgO for the low-Mg mafic komatiites. The significance of this compositional gap is discussed in greater detail in Chapter 6.

The high-Mg mafic komatiites encompass the Geluk type basaltic komatiites proposed by Viljoen and Viljoen (1969c). Their subdivision into high-and low-Al subtypes has not been substantiated by the additional data obtained in this work, as a complete gradation between high and low Al_2O_3 values is found (see Chapter 6 and Fig. 4-1). The major and trace element compositions of this group of lavas are given in Tables 6-2a and 6-2b. It should be noted that some porphyritic samples of this group are considered to represent liquid compositions as discussed in Chapter 6. Data for these latter samples are given in Tables 6-3a and 6-3b.

c) Low-Mg Mafic Komatiites

These lavas have MgO contents ranging from 8 to 16% MgO and have clinopyroxene as the major phenocryst phase. Plagioclase if present is minor while olivine is very seldom observed and orthopyroxene has not been observed. However

indirect lines of evidence suggest that olivine may have been present in greater amounts (and now altered) in the higher MgO lavas of this group (see Chapters 3 and 6) than is indicated by the textural evidence. The low-Mg mafic komatiites are all quartz normative. The Barberton type and Badplaas type of basaltic komatiites originally proposed by Viljoen and Viljoen (1969c) have been incorporated into this single group (as a complete gradation of compositions were found in this work). The lower MgO lavas (evolved) in the low-Mg Mafic komatiites however appear to be more common in the LUU than the higher MgO rocks (basic).

Clinopyroxene spinifex textured rocks in this compositional range appear to be fairly common in other Archaean greenstone belts. Viljoen and Viljoen (1969c) have noted the frequent outcrop of coarsely crystalline amphibolites characterised by actinolite needles up to 10 cms long in some cases. The needles are randomly oriented in some instances and in others show strong parallel alignment. They propose, as one possibility, that this texture could be formed by analogous processes to that forming the crystalline quench structures (olivine spinifex) in the ultramafic komatiites. Williams and Furnell (1979) also consider these horizons to represent mafic komatiite lava flows. In this work these textures are assumed to be metamorphosed clinopyroxene spinifex textured samples with the clinopyroxene having been replaced by actinolite. Only one sample of this type has been analysed and is considered to be representative of a liquid composition.

Many rocks of tholeiitic composition from the LUU fall in the 8-16% MgO range of the mafic komatiites. However the tholeiites are readily distinguished from the mafic komatiites as they have higher Al_2O_3 contents at given MgO (see Fig. 4-1 and Chapter 7). As can be seen from Fig. 4-1 no lavas representing liquids have Al_2O_3 in the range of 12-13.5% Al_2O_3. This compositional break is unlikely to represent inadequate sampling as numerous samples have been analysed of both low-Mg mafic komatiites and tholeiites in the 8-12% MgO range. It is therefore concluded that if rocks representative of liquids occur in the 12-13.5% Al_2O_3 range in the LUU, they must be extremely rare. In addition low-Mg mafic komatiites generally contain >600 ppm Cr while the tholeiitic lavas have 300 ppm Cr or less, again indicating that the two lava types are not simply related. Representative major and trace element data are given for the low-Mg mafic komatiites in Tables 6-4a and 6-4b.

Detailed modelling of the geochemical trends (Chapter 6) shows that clinopyroxene becomes an important fractionating phase in low-Mg mafic komatiite magmas at ~14% MgO. Throughout this work low-Mg mafic komatiite lavas with >14% MgO are referred to as 'basic' while those with <14% MgO are termed 'evolved'.

A summary of the classifying parameters used here for the LUU komatiites is given in Table 4-2. It should be emphasised that this scheme has been developed specifically for the Onverwacht Group volcanics and has not as yet been systematically tested for komatiites in general. In Chapter 7 sub-groups of ultramafic and mafic komatiites

are identified on the basis of inter-element ratios. These differences however are not considered to be significantly large to warrant separate classification schemes.

4-9 DISCUSSION

The main feature of the classification scheme outlined above is that the number of categories of komatiite types have been reduced to three compared to the five fold division given in Viljoen and Viljoen (1969c,d). The inclusion of textural requirements in the komatiite definition has eliminated many problems caused by the compositional similarity of certain other coarse grained ultramafic rock types (see Brooks and Hart, 1974). Several other features of the komatiites stand out and warrant brief discussion:-

a) The sub-alkaline nature of komatiites

Although absolute concentrations of elements such as K_2O may have been seriously perturbed by alteration processes (see Chapter 2) the generally low alkali content of the komatiites is consistent with the low abundance of other less mobile incompatible elements such as Ti and Zr. The sub-alkaline nature of the komatiites is an important feature for distinguishing these lavas from other high MgO magmas such as kimberlites, melilitites and meimechites among others.

b) Low TiO_2 contents

This feature can be used in a more quantitative fashion than the sub-alkaline nature of komatiites as Ti is believed to have remained relatively immobile

during the alteration events that have affected the Barberton greenstone belt lavas (see Chapter 2). Brooks and Hart (1974) have suggested a value of <0.9% TiO_2 for komatiites which distinguishes them from rocks such as ankaramites. This upper limit value of 0.9% TiO_2 for komatiites is supported here and also helps distinguish komatiites from other alkaline magmas (kimberlites etc.) mentioned above. Arndt et al. (1977) have also used the TiO_2 contents of tholeiitic lavas to separate them from komatiitic lavas. However a considerable range in the TiO_2 contents of tholeiitic lavas from the Barberton greenstone belt (see Fig. 4-2) is observed which overlaps the TiO_2 range in the mafic komatiites. The two TiO_2-MgO trends (Fig. 4-2) for the mafic komatiites and tholeiites are however very different, but do not help to uniquely identify individual low-Ti tholeiites.

c) CaO/Al_2O_3 ratios

From Table 4-2 (and Tables 5-3b, 6-2b and 6-4b) it can be seen that while high CaO/Al_2O_3 ratios are prevalent, values as low as 0.8 are not uncommon in non-cumulate komatiite lavas. The variations of this ratio with MgO content in the komatiites is discussed in detail in Chapter 5 and 6. However the limit of CaO/Al_2O_3 >1 placed on komatiites by Brooks and Hart (1974) is discounted by the type area data.

The tholeiitic rocks from the Barberton greenstone

belt have CaO/Al_2O_3 ratios in the range 0.66 to 0.85 (see Table 7-2b) and are thus generally lower than the mafic komatiites. The CaO contents of these two lava types is similar at the same MgO content and therefore the higher CaO/Al_2O_3 ratios in the mafic komatiites are due to their lower Al_2O_3 contents (Fig. 4-1).

The state of komatiite nomenclature in general brings to mind Mark Twain's comment (quoted in Scott and Hajash, 1976). "The researches of many commentators have already thrown much darkness on this subject, and it is probable that, if they continue, we shall soon know nothing at all about it." Hopefully such a pessimistic view can be overcome once it is realised that komatiites from different localities have individual geochemical characters while still preserving a sufficient number of common characteristics to warrant considering them as the same class of rocks.

CHAPTER 5.

GEOCHEMISTRY OF THE ULTRAMAFIC KOMATIITE LAVAS

5-1 INTRODUCTION

Viljoen and Viljoen (1969d) presented convincing evidence for the existence of rocks in the Barberton greenstone belt that represent crystalline ultramafic magmas. Although the existence of such magmas had previously been postulated (see review in Viljoen and Viljoen, 1969d) the evidence was not generally accepted. Ultramafic lavas have now been recorded in many Archaean greenstone belts, and for the purpose of this Chapter the well documented areas from Canada (Naldrett and Mason, 1968; Pyke et al., 1973; Arndt et al., 1977; Lajoie and Gilinas, 1978), Western Australia (Nesbitt, 1971; Williams, 1971; Williams, 1972; Naldrett and Turner, 1977) and Rhodesia (Bickle et al., 1975; Nisbet et al., 1977) will be discussed in relation to the Barberton ultramfic rocks.

In all four of these greenstone belts the ultramafic rocks crop out as massive flows (often containing olivine spinifex textured portions), pillows, polyhedra, and thick sill or lens-like differentiated bodies. Recent experimental work supports the field evidence for the existence of ultramafic lavas by duplicating textures (Donaldson, 1974, 1976, 1977), mineral compositions (Green et al., 1975; Arndt, 1977b; Bickle et al., 1977) and mineral assemblages observed in the rocks.

In this Chapter, only brief descriptions will be given of the field occurrence and petrography of the ultramafic lavas from the Lower Ultramafic Unit (LUU), as much of this data have been presented by Viljoen and Viljoen (1969c,d). The mineral and whole rock chemistry will be discussed in detail and for this purpose the ultramafic samples have been divided on textural grounds into aphyric and porphyritic

groups. The aphyric group is believed to represent essentially phenocryst free lavas and include samples of chilled margins (<2% olivine phenocrysts pseudomorphed by serpentine and magnetite), pillow lavas and olivine spinifex textured samples. The porphyritic group contains olivine and some pyroxene phenocrysts and in many cases are demonstrably partial cumulates. It should be noted that where olivine is referred to in the following descriptions, it has generally been pseudomorphed by serpentine and magnetite, although relict kernels of olivine are often preserved in the porphyritic rocks. The geochemistry and relationships between these two groups are first discussed with respect to low pressure igneous processes that may have contributed to the generation of the trends. The geochemical data of the ultramafic lavas from the Barberton greenstone belt are then compared to similar data from ultramafic rocks from the other greenstone belts mentioned previously. The more speculative interpretation of the high pressure processes that have contributed to the generation of the ultramafic magmas will be discussed in Chapter 8.

5-2 SAMPLE LOCATION, FIELD CHARACTERISTICS AND PETROGRAPHY

The ultramafic samples used in this study are from the three lower Formations of the Onverwacht Group which together make up the Lower Ultramafic Unit (LUU) (Viljoen and Viljoen, 1969c,d). The uppermost Formation of the LUU, the Komati Formation, contains rocks in which igneous structures and textures can be readily identified and consequently the majority of the ultramafic samples analysed for this project are from the Komati Formation type area. The location of the samples within the Komati Formation is shown on Fig. 5-1 and 5-2. The ultramafic rocks from the Theespruit Formation are less abundant than in the

Komati Formation and not as well preserved. However, one sample (HSS-150) from this formation was sufficiently well preserved to warrant analysis. Nine samples from the type area of the Sandspruit Formation were also analysed. The type area was established for this Formation in a detached xenolithic block in the south west portion of the Barberton greenstone belt by Viljoen and Viljoen (1969c). The samples are from near the homestead on the farm Brandybal; no detailed geological maps are available for this area. In general the igneous structures and textures of the rocks from this area are poorly preserved but the samples were analysed mainly to obtain data for comparison with the ultramafic rocks from the Komati Formation.

Fifteen samples analysed for major elements by Viljoen and Viljoen (1969b,c,d) from the Komati and Sandspruit Formations have been analysed for trace elements in this work. Sample locations and brief descriptions of these rocks are given in Viljoen and Viljoen (1969b,c,d). All samples that are not prefixed by HSS or SD in Tables 5-3, 5-4 and 5-5 are rocks sampled by the Viljoens.

The ultramafic rocks in the Komati Formation crop out as flows, pillows, olivine-rich ridges and dykes. Brief field and petrographic descriptions are given in Appendix-1. Samples that are considered to represent phenocryst free liquid compositions are collectively termed aphyric lavas and include the following types:-

1) chilled margins of flows and dykes (HSS-530, HSS-88C, 53-J)
2) pillows (HSS-523, AU-5)
3) Random and plate spinifex textured samples (HSS-95, HSS-531, HSS-532, HSS-533, HSS-88A, HSS-90, HSS-89, HSS-109)
4) Aphanitic aphyric rocks (HSS-87, HSS-14, HSS-15, HSS-150)

Although chilled margins and pillows could be enriched in olivine phenocrysts, the textural evidence suggests that this effect, if it occurred at all, was very limited (<2% microphenocrysts).

As will be demonstrated later in this Chapter, due to the low viscosity of the ultramafic magmas, phenocrysts could settle through considerable distances in the lava in relatively short times. All porphyritic samples are therefore potentially, and in some cases, demonstrably phenocryst enriched and as such are not representative of phenocryst free magma compositions. Rock incorporated into this group are here termed porphyritic lavas and include the following types:-

1) Abundant phenocrysts or microphenocrysts of olivine (HSS-534, HSS-535, HSS-536, HSS-88B, HSS-93, HSS-13, VU-25, V-2, VU-30A, VU-32A, HSS-8)

2) Medium grained olivine-rich rocks (R-13, R-16, 20-J)

3) Fine to medium grained olivine-pyroxene rocks (HSS-33, HSS-92, HSS-27, HSS-31, HSS-1, HSS-28, 52-J).

4) Medium grained pyroxenite (SD-76)

An additional group of samples has been analysed that comprises most of the samples from the Sandspruit Formation. These samples have no igneous texture or mineralogy preserved and are considered to be of unknown composition in terms of the aphyric-porphyritic distinction made above for the Komati and Theespruit Formation rocks. The samples analysed from this group are 86-J, 87-J, 88-J, VU-33, HSS-130, HSS-131, HSS-112, HSS-107 and HSS-105.

APHYRIC LAVAS

Detailed descriptions of the olivine spinifex textured lavas from

the Komati Formation have been given in Chapter 3 and only a brief general description will be given here. Two types of olivine spinifex texture have been recognised in the ultramafic flows from the LUU. The first type being random spinifex forming the upper portion of the flows (10-40 cms. thick) and consisting of randomly oriented olivine blades (usually <2 cms. long) with interblade areas of serpentine, magnetite and amphibole. These interblade areas often contain hopper olivine crystals and spherical vesicles lined with magnetite. The plate spinifex occurs in zones, 10 to 150 cms. thick, below the upper random spinifex in the flows. The olivine plates which form in fan or booklet arrangements are usually >2 cms. long and may be 30 cms. or longer. The interblade areas of the plate spinifex also consist of serpentine magnetite and amphibole. However, no vesicles or hopper olivines have been observed.

The basal chilled margins of the flows (10-20 cms. thick) are essentially aphyric rocks containing spherical vesicles and occasional (<2%) microphenocrysts of olivine. The olivine crystals either form as highly skeletal hopper crystals (∿0.6x0.3 mm.) or as solid euhedral crystals (0.1-0.2 mm. across). The sample 53-J is from the chilled margin of a crosscutting ultramafic dyke in the Komati Formation (Viljoen and Viljoen, 1969d) and is a holocrystalline rock consisting of rounded or elongated olivine grains poikilitically enclosed by clinopyroxene crystals.

The pillow samples are aphanitic aphyric rocks although sample AU-5 does contain small olivine cyrstals (0.2 mm. across) in a groundmass of serpentine, amphibole and magnetite. Arndt (1975) has suggested that the polyhedral jointing of the type described by Pyke et al. (1973) and Arndt et al. (1977) in the ultramafic flows from Munro Township

could generate structures (polyhedra) that have been mistaken for pillows in the Komati Formation. However, re-examination of the outcrop illustrated in Viljoen and Viljoen (1969d, see plate Vb) from the Komati Formation shows very convincing pillow structures. The pillows have chilled margins that are much finer grained than the fine grained interior. It should be noted, however, that outcrop of convincing ultramafic pillow structures is very limited in extent.

For the final group of samples that is considered to represent liquid compositions (aphanitic aphyric rocks), outcrop was not sufficiently good to determine precisely which portion of the flows (or sills) the samples were from. In thin section the samples consist of a groundmass of serpentine amphibole magnetite and chlorite with very occasional crystals of olivine up to 0.3 mm. across.

PORPHYRITIC LAVAS

The porphyritic samples consist of rounded subhedral grains of olivine (0.4-1.0 mm.) across, making up between 50-80% of the rock. The groundmass usually consists of amphibole, magnetite, serpentine and chlorite. In this group of rocks relict olivine cores are often preserved. Many of the samples were taken from clearly identifiable cumulate (B zone) portions of spinifex textured ultramafic flows.

Three samples of the medium grained olivine rich rocks are from the Viljoens' collection, one of which (20-J) is from the Hooggenoeg Formation. These samples consist of large olivine crystals (2-4 mm. in their longest dimension) set in a groundmass of serpentine and magnetite. Amphibole and chlorite, if present, are very minor. This group of samples probably represents olivine-rich cumulates from the flows or sills.

The olivine-pyroxene group of cumulates are holocrystalline rocks with olivine (rounded grains 0.3-1.5 mm. across) making up 60-80% of the rock poikilitically, enclosed by large irregularly shaped clinopyroxene grains. The olivine is usually replaced by serpentine and magnetite while the clinopyroxene grains are usually well preserved in most of the samples. Occasional grains of magnetite and chromite occur as inclusions in the clinopyroxene. Many of the samples are from medium to coarse grained horizons, oriented parallel to the general layering of the basaltic and ultramafic flows and may represent sills or the lower portions of thick flows. Sample 52-J is from the clearly identifiable crosscutting dyke in the Komati Formation and has a texture typical of this group of cumulate rocks (see Plates X1(a) and X1(b) in Viljoen and Viljoen, 1969d).

Of all the ultramafic samples examined in thin section from the LUU, only three were found to contain orthopyroxene. The sample SD-76 is a monomineralic orthopyroxene-cumulate from the Stolzburg body, a layered body consisting predominantly of alternating horizons of dunite and pyroxenite. Many differentiated bodies occur in the Komati Formation and are believed to have formed contemporaneously with the ultramafic and basaltic lavas (Viljoen and Viljoen, 1969h, 1970b). The other samples, in which orthopyroxene has been identified, are HSS-27 and HSS-31. While the fine detail of the contact zone between olivine and orthopyroxene has been obscured by alteration, orthopyroxene does mantle the olivine grains, suggesting a reaction relationship between the two minerals.

From the above discussion of the petrography of the aphyric and porphyritic rocks, it is clear that olivine is the dominant phase crystallising from the ultramafic magmas. Clinopyroxene and opaques crystallise

much later than olivine while orthopyroxene appears to be an unimportant phase in the lavas. This sequence of crystallisation is in agreement with that observed experimentally for ultramafic komatiite lavas, as Green et al. (1975), Arndt (1977b) and Bickle et al. (1977) have shown that olivine is the predominant liquidus phase and at lower temperatures is joined by clinopyroxene (Arndt, 1977b).

SANDSPRUIT FORMATION ULTRAMAFIC ROCKS

Most of the samples in this group consist of serpentine and chlorite with minor variable amounts of amphibole, magnetite, talc and epidote. Two samples (87-J and 88-J), however, consist predominantly of actinolite-tremolite, large flakes of chlorite and some magnetite and lack the major amounts of serpentine present in the other samples. No igneous textures are preserved in any of the samples except the spinifex textured sample HSS-109. This sample is included with the other aphyric group of samples in the following discussions.

5-3 MINERAL CHEMISTRY

OLIVINE

The discussion of the mineral chemistry will be restricted to the relict igneous mineralogy of the ultramafic lavas and in this respect olivine is the most important phase. Olivine compositions have been determined from several porphyritic ultramafic rocks, as relict olivine grains are preserved in these samples. The average olivine composition and range of (Mg x 100)/(Mg+Fe) ratios for each sample are given in Table 5-1. Where core-edge pairs of the relict olivine grains have been analysed they show normal zoning (e.g. see Table 3-1). However, as in Chapter 3, it is again emphasised that the edge analyses do not

represent the true margin of the original olivine crystal. Unfortunately no primary olivine was preserved in any of the aphyric group of samples investigated in this study. However, olivine is preserved in the olivine spinifex textured sample, 49-J, from the Komati Formation and olivine analyses from this sample are given in Green et al. (1975). High Cr_2O_3 and CaO contents of the olivines have also been noted in Chapter 3 and in Fig. 5-3, the Cr_2O_3 content has been plotted against the Fo composition for all the olivine analyses. In this diagram a broad decrease in the Cr_2O_3 content can be seen with decreasing Fo content. NiO shows a similar decrease (Fig. 5-4) with decreasing Fo content.

CLINOPYROXENE

Fresh clinopyroxenes have been analysed from three ultramafic komatiites from the LUU and the average compositions and range of Mg x 100/(Fe+Mg) ratios are given in Table 5-1. Individual analyses have been plotted on a pyroxene quadrilateral (Fig. 5-5) and plot around the line dividing the endiopside and augite fields. The most striking feature of the clinopyroxenes are their sub-calcic compositions similar to clinopyroxenes from kimberlite nodules (e.g. Davis and Boyd, 1966) and quite distinct from the trend observed in large differentiated bodies such as Skaergaard or the Bushveld Igneous Complex (e.g. Atkins, 1969). The diopside solvus temperatures determined at 30 kb. by Davis and Boyd are indicated on Fig. 5-5. Davis and Boyd (1966) have shown that clinopyroxene compositions in the system $Mg_2Si_2O_6$-$CaMgSi_2O_6$ are insensitive to pressure variations as the 1 atmosphere solvus (determined by Boyd and Schairer, 1964) is virtually within experimental error of the 30 kb. data. The sample HSS-27

contains a range of ortho- and clinopyroxene compositions and assuming that the most Mg-rich ortho- and clinopyroxenes were in equilibrium and that the most Ca-rich clinopyroxene was in equilibrium with the most Fe-rich orthopyroxene, a temperature range of pyroxene crystallisation from 1300 to 1000°C is indicated for this sample. The liquidus temperatures of the aphyric rocks is believed to be in the range 1500-1650°C which suggests that pyroxene only became an important liquidus phase at much lower temperatures than olivine. This is consistant with the experimental data of Arndt (1976).

The fields occupied by the clinopyroxenes from Belingwe and Munro Township are shown on Fig. 5-5; the low Fe content and Ca enrichment trend of the LUU clinopyroxenes are distinct from the relatively Fe enriched field of clinopyroxenes from Munro Township and the Fe enrichment trend shown by the Belingwe clinopyroxenes. The clinopyroxenes from the LUU ultramafic rocks have similar high TiO_2 contents to those from the Belingwe (Nisbet et al., 1977) and Munro Township (Arndt et al., 1977) ultramafic lavas. The Cr_2O_3 contents of the LUU pyroxenes are higher than in the latter two areas; the highest Cr_2O_3 content measured in the clinopyroxenes from the LUU rocks is from HSS-27 where values as high as 1.4% Cr_2O_3 have been recorded.

ORTHOPYROXENE

The composition of orthopyroxenes have been determined in two samples from the LUU ultramafic rocks (Table 5-1 and Fig. 5-5). The sample SD-76 is a monomineralic pyroxene cumulate from the Stolzburg Body (Viljoen and Viljoen, 1969h, 1970b). The enstatite crystals from this sample are unzoned, uniform in composition and contain very low contents of Al_2O_3 and TiO_2. In contrast the orthopyroxenes from

HSS-27 (Bronzite) show a spread in composition and contain high contents of Al_2O_3. The difference between the orthopyroxene compositions in these two samples could be related to the different crystallisation rates in the two rock types. The orthopyroxenes in the Stolzburg sample probably crystallised more slowly than the flow sample.

5-4 GEOCHEMISTRY OF THE ULTRAMAFIC KOMATIITES

The major element compositions and normative mineral contents are given in Tables 5-3a, 5-4a and 5-5a for the aphyric lavas, porphyritic lavas and altered Sandspruit Formation samples respectively. Similarly the trace element data and inter-element ratios are given in Table 5-3b, 5-4b and 5-5b. All the data reported in these tables have been calculated to a volatile free basis,although the volatile content and the original totals of the major element data are reported (Tables 5-3a to 5-5a). Major element data for the samples from the Viljoens' collection (Viljoen and Viljoen, 1969b,d) that have been analysed for trace elements in this work, are also given in Tables 5-3a to 5-5a. Data for sample 49-J analysed by Green et al. (1975) are also given in Tables 5-3a and 5-3b. Note that the data given in tables 5-3a to 5-5a has had Fe apportioned so that $Fe_2O_3/FeO = 0.20$.

ALTERATION EFFECTS

No systematic investigation of the effects of alteration have been carried out on the ultramafic rocks as was done for the mafic lavas (Chapter 2). This is partly due to the difficulty of selecting a suitable suite of ultramafic samples that have been altered to differing extents, but for which the assumption that all the samples initially had the same composition, is valid. Nevertheless, some

attempt to assess the effects of alteration on the ultramafic lava compositions has been made and it is pertinent to discuss these matters here, before dealing with possible igneous processes that controlled the lava compositions. It is obvious from examination of thin sections of these rocks that the major alteration process affecting the samples is that of serpentinisation. Viljoen and Viljoen (1969b,d) have noted two stages of serpentinisation in the Komati Formation ultramafic rocks. The first stage they suggest was incipient alteration (affecting the olivine grain boundaries) followed by a second stage of serpentinisation of the olivine cores, with antigorite being the serpentine mineral formed in both stages. They further suggest that Ca along with small amounts of Si, Al, Na and K were lost from the samples during this process and that at H_2O^+ contents greater than 10%, the rocks are completely 'de-natured'.

The problem of the effects of serpentinisation on the geochemistry of rocks has been extensively studied (Thayer, 1966, 1967; Page, 1967, 1976; Condie and Madison, 1969; Martin, 1971; Gülacor and Delaloye, 1976; Moeskops, 1977) but there is little consensus of opinion as to whether serpentinisation is a constant volume (change in composition) or volume increase (addition of H_2O) process. Several authors have attempted to apply chemical screens to the geochemical data (Viljoen and Viljoen, 1969b; Nesbitt et al., 1979) to aid in the selection of relatively unaltered ultramafic lava compositions. However, chemical screens alone (such as maximum or minimum contents of certain constituents or restricted ranges of selected inter-element ratios) are considered unsatisfactory as the data selected already conform to some preconceived behaviour. Such subtleties as the serpentinisation of clinopyroxene, which according to Williams (1972), strongly

affect the original chemical characteristics of the parent rock, may pass undetected through the screening test. Consequently, in this Chapter, samples that are used for detailed geochemical interpretation have been selected on the basis of textural preservation, lack of complete serpentinisation and in particular, lack of serpentinisation of the clinopyroxene, although samples where pyroxene has been converted to amphibole have been analysed.

The effects of recalculating the major and trace element data volatile free are illustrated in Fig. 5-6, where CaO, Al_2O_3 and V have been plotted against MgO contact of the aphyric lavas before and after correction. It can be seen from this diagram that while the correlation of these three elements with MgO has not been visibly improved by the correction procedure, the slope and intercept of the three corrected trends have been significantly changed. It is stressed that it is equally important to correct the trace element data as it is to correct the major element data for the added volatile content.

The possible alteration effects on the calcium concentration must, however, be considered in more detail as CaO is one of the key oxides in the interpretation of ultramafic komatiite geochemistry. It has been suggested (Nesbitt and Sun, 1976) that the CaO content and CaO-MgO trend in the Barberton ultramafic lavas could have been affected by alteration processes in such a way that reasonable trends were developed, but did not result directly from igneous processes. From Fig. 5-6 an excellent correlation between CaO and MgO is observed for the aphyric lavas from the LUU. Good correlations are also observed for many other oxide/element-MgO plots (Fig. 5-8) from the aphyric lavas. Therefore there appears to be no reason to suspect

only the CaO-MgO trend of having been generated by alteration processes and the possibility that all the trends (Fig. 5-8) have been systematically rotated by alteration processes must be considered. In Fig. 5-7a the MgO, CaO and Al_2O_3 concentrations (before correction for volatile content) have been plotted against H_2O^+ (only data obtained in this laboratory are used) for the group of aphyric ultramafic lavas. MgO correlates positively with H_2O^+ while CaO and Al_2O_3 show negative correlations with H_2O^+. Two possible explanations are advanced for these trends: a) they represent trends controlled by the secondary mineralogy (greenschist facies assemblages) of the samples, or b) the trends represent igneous trends in which the predominant alteration effect is the dilution of the oxide concentrations by the addition of volatiles.

If possibility a) were true then the major phases to be considered are serpentine, amphibole (actinolite), chlorite, unresolved groundmass and minor magnetite. The unresolved groundmass probably consists of the same minerals. The calcium bearing minerals occurring in the samples considered in this study, are actinolite and very small amounts of calcite and epidote. As calcite and epidote are present in very small amounts they contain an insignificant portion of the total CaO present in the samples and will, therefore, be disregarded from further discussion.

The compositional fields allowed for the aphyric lavas, if they represented mixtures of serpentine, actinolite and chlorite, have been outlined in Figs. 5-7a and 5-7b. The compositions of these secondary minerals have been analysed in komatiite lavas from the Komati Formation and are given in Table 5-2. Chlorite and serpentine were assumed to contain 13% H_2O^+ and actinolite 3% H_2O^+. The H_2O^+-CaO

trend plots along the serpentine plus chlorite-actinolite control line (Fig. 5-7a) but the possibility that the ultramafic lava compositions are controlled by these three minerals is discounted as the H_2O^+-MgO, H_2O^+-Al_2O_3 and CaO-MgO trends (Figs. 5-7a and 5-7b) fall outside the fields of allowed compositions. However, the possibility that the trends could be developed by partial alteration of primary minerals or glass is not precluded, but since the degree of alteration varies between the samples (as reflected by H_2O^+ contents from 4.8-9.6%) such an incomplete process would not be expected to generate trends (e.g. CaO-MgO) with such good correlations.

Possibility b) above is the explanation that is favoured. Here it is considered that the secondary mineralogy was controlled by the composition of the rock, i.e. more serpentine could be formed in the higher MgO samples and more actinolite could be formed in the higher CaO samples, than samples with lower contents of these oxides. If CaO-MgO and Al_2O_3-MgO were correlated in the unaltered igneous rocks, these correlations would be preserved in the metamorphosed samples. In order to reproduce the original trends, the data should be recalculated volatile free.

Evidence for the primary nature of the trends displayed in Fig. 5-8 is largely circumstantial and outlined below:-

1) The range of compositions in Stuart's Flow (Chapter 3) were successfully modelled assuming that they were generated by the igneous processes of crystallisation and accumulation of olivine. This model is consistent with what can be inferred or observed about the way in which the flow differentiated, and it would therefore appear to be a valid assumption that the compositions

of the samples (for the elements considered) reflect the range of pre-alteration compositions occurring in the flow. As this flow is typical of many of the flows from which samples were taken and used to define the trends in Fig. 5-8, it follows that the trends also reflect the pre-alteration trends in the lavas.

2) On the basis of Al/Ti ratios (among others), absolute TiO_2 abundance and CaO-MgO trends, two distinct types of ultramafic komatiite lavas have been identified from the LUU (as discussed below) which nevertheless show broadly similar ranges of compositions (e.g. ∿24 to ∿33% MgO for the aphyric lavas). As lava flows of each type are intercalated in the Komati Formation, have similar primary (where preserved) and secondary mineral assemblages and textures, they were probably affected by the same alteration events. If the effects of alteration were to systematically modify the igneous composition of the ultramafic lavas, the differences outlined above between the two types would not be expected. The only reasonable explanation is that the differences reflect a real difference in the compositions of the magmas. The differences in the CaO-MgO trends between the two groups are therefore believed to represent a real igneous difference.

Returning to the contention of Nesbitt and Sun (1976) that the CaO-MgO trends in the ultramafic lavas could have been systematically rotated by alteration processes, it is here considered an unlikely possibility. From the evidence considered above, the interpretation that the

CaO-MgO trends (and other trends in Fig. 5-8) reflect real igneous trends is favoured and the following discussion of the geochemical data from the LUU ultramafic komatiites will be based on this premise. It should, however, be borne in mind that while the ultramafic samples are relatively well preserved, they have nevertheless been serpentinised to variable extents and occasionally aberrant concentrations of an element may occur. In addition the variations in the P and Nb contents of the lavas will not be considered, due to the poor analytical precision obtained for these element concentrations at the low levels encountered in the ultramafic rocks.

APHYRIC LAVAS

The major and trace element data for the aphyric ultramafic lavas are given in Tables 5-3a and 5-3b respectively. The samples have been arranged in order of increasing MgO and display a wide range of MgO contents from 24-33% with an average value of 28.95% (see Table 5-6). The other oxides and elements such as SiO_2, TiO_2, Al_2O_3, FeO, MnO, CaO, Zr, Y, V, Ga and Sc show a decrease in concentration with increasing MgO content. These variations are illustrated in Fig. 5-8, where the above elements as well as Cr_2O_3 and Co have been plotted against MgO content. All the data have been recalculated volatile free. As can be seen from the plots, many of the elements show a regular and generally linear decrease in content with increasing MgO. However, two of the spinifex samples (HSS-95 and 49-J) have significantly lower TiO_2 contents than the rest of the samples comprising the TiO_2-MgO trend and will be discussed in a later section.

Linear regression equations have been calculated for the data and the best fit lines plotted on the diagrams (Fig. 5-8) and

extrapolated to 50% MgO. As the major element data of the samples reported in Viljoen and Viljoen (1969b,d) are from different laboratories using a variety of different analytical techniques, the regression equations have been calculated using only data from this laboratory, to avoid complications caused by any inter-laboratory variation that may exist. Three further samples have been rejected, HSS-95, 49-J and HSS-88C, the first two due to their low TiO_2 contents and the latter, as it plots off the general trends for many elements. The slopes, intercepts, correlation coefficients as well as compositions calculated from this data at various MgO contents, are given in Table 5-7. Two sets of values have been calculated for the Zr-MgO, Y-MgO and Cr_2O_3-MgO trends. The second set of values has been calculated after further samples have been rejected (listed in Table 5-7) that appear to bias the calculated trend line away from, what in the author's subjective opinion, appears to be the best fit. This subjective refinement has made little difference to the Zr-MgO and Y-MgO data, but has changed the slope and intercept of the Cr_2O_3-MgO trend significantly. For the Cr_2O_3-MgO trend this approach is partly justified by the difficulty in detecting any enrichment of small chromite microphenocrysts that are present in the samples.

Except for the CaO-MgO trend all the major element-MgO trends pass through or very close to the field of olivine compositions determined in the ultramafic samples. The MgO intercept value for the CaO-MgO trend is well defined due to the good correlation between these two elements and the relatively short extrapolation of the trend to the MgO axis. The intercept value of 40.66% MgO is clearly displaced from the field of olivine compositions and eliminates the possibility that olivine was the only mineral controlling the evolution of the ultramafic magmas.

In Fig. 5-9 the predicted olivine differentiation paths are shown in relation to FeO^*-MgO data of the aphyric lavas. The fractional and equilibrium crystallisation curves have been calculated (using the equations given in Pearce, 1978) assuming a parent liquid composition of 33% MgO and 11.24% FeO and K_D values of 0.30 (Roeder and Emslie, 1970) and 0.37 (from Chapter 3).

where $K_D = (X^{ol}_{FeO}/X^{ol}_{MgO})/(X^{liq}_{MgO}/X^{liq}_{FeO})$

All the calculated trends form more iron rich compositions than the regression line trend. Although the scattered nature of the whole rock data do not enable firm conclusions to be made, it appears unlikely that the range of aphyric ultramafic lava compositions were derived by fractional crystallisation of olivine from a 33% MgO parental magma, which is consistent with the CaO-MgO trend discussed above.

As outlined in previous sections, examination of the ultramafic liquids in thin section shows that olivine (occurring as spinifex textured blades, quenched hopper olivines or euhedral microphenocrysts) is the first mineral to crystallise, followed at a later stage by pyroxene in the interblade areas of the olivine spinifex textured samples. The ultramafic cumulate samples also show olivine as the main phase being accumulated and clinopyroxene, if present, only as late interstitial crystals. Detailed modelling of Stuart's flow also demonstrated olivine to be the only important fractionating phase for accounting for the range of compositions in a cross-section of the flow. In addition, the experimental work of Green et al. (1975) and Arndt (1976, 1977b) shows that olivine is the only liquidus mineral for a wide range of temperatures in ultramafic komatiite liquids at low

pressures. It therefore appears from the field, thin section and experimental evidence that olivine is the only mineral likely to be involved in the control of the ultramafic liquid compositions in the low pressure environment. However, as the geochemistry of the LUU aphyric ultramafic lavas does not reflect this expected olivine control, the liquid trends must be developed by melting or high pressure fractionation processes.

The trace elements in Fig. 5-8 show two distinct trends with increasing MgO content, those that increase, i.e. Co and Ni (Fig. 5-12) and those that decrease such as Zr, Y, V, Ga and Sc (Sc not plotted). No curved trends have been detected. The trace elements that are believed to be incompatible in the olivine structure such as Zr, Y, Ga and Sc (i.e. $D^{ol/liq} << 1$) do not project through likely olivine compositions. This tends to support the major element data, that olivine fractionation alone does not control the ultramafic aphyric lava trends, although for some trace element data the regression lines are less well controlled, due to the scatter of the data (Zr), limited amount of data (Sc) or very low concentrations (Y, Ga). For the Cr_2O_3-MgO data the selected trend (solid line on Fig. 5-8) suggests an increase of Cr_2O_3 with decreasing MgO. However, because of the scatter in the data no firm conclusions can be made regarding the behaviour of Cr with increasing MgO content of the lavas. Nevertheless, the selected trend is similar to that in the Belingwe lavas (Nisbet et al., 1977).

From consideration of the inter-element ratios, further constraints can be placed on the processes involved in the production of the ultramafic aphyric lava trends. Perusal of the Al/Ti data (Table 5-3b) shows restricted range of values for this ratio with most samples having

values close to 9.6, while the samples HSS-95 and 49-J have considerably higher values of 16.3 and 16.6 respectively. An equivalent feature is displayed by the Ti/V ratio where most of the samples have values close to 20 while values for HSS-95 and 49-J are considerably lower (14 and 12 respectively, see Table 5-3b). These features of low Ti and different Al/Ti and Ti/V ratios in HSS-95 and 49-J (and several porphyritic samples, as identified in the next section) set them apart from the major group of ultramafic lavas and as such will be discussed in a following separate section, and will not be considered again in this discussion.

From consideration of all the inter-element ratios in Table 5-3b the following observations can be made:-

A) The following inter-element ratios show no systematic change with increasing MgO content and have relatively restricted ranges of values:-
Al/Ti, Ti/Zr, Ti/Y, Ti/V, Ti/Ga and Ti/Sc.
Therefore, any process postulated for the generation of the LUU magmas must do so without fractionating the elements Ti, Al, Zr, Y,V, Ga and Sc to any marked extent.

B) The ratios Si/Ti, Fe/Ti and Mg/Ti show a systematic, although somewhat erratic increase, and Ti/Ni and Ti/Co show an erratic decrease, with increasing MgO content. Processes postulated for the generation of these ultramafic komatiites must account for the systematic increase of Si, Fe, Mg, Ni and Co relative to Ti and the other group (A) elements with increasing MgO content of the magmas.

C) The Ca/Ti, Ca/Mg and CaO/Al_2O_3 ratios show a regular decrease with increasing MgO content of the lavas. Petrogenetic processes must account for the decrease in CaO content relative to both the group (A) and (B) elements with increasing MgO content.

These observations place severe constraints on the minerals and initial mantle composition that could have been involved in the development of the ultramafic liquids at high pressures. Models for the processes possible in the genesis of the ultramafic liquids will be tested in Chapter 8.

PORPHYRITIC LAVAS

The major and trace element data are given in Tables 5-4a and 5-4b for the ultramafic porphyritic lavas. The data have been arranged in three groups, namely the olivine phyric samples, the olivine rich samples and the olivine-clinopyroxene phyric samples. The samples within each group have been arranged in order of increasing MgO content. Both the major and trace element data have been calculated volatile free and with Fe_2O_3/FeO ratio set to 0.2. The variations of selected elements against MgO content are shown in Fig. 5-10. Regression lines have been calculated for the data and, as for the aphyric lavas, only samples analysed in this laboratory have been used. In addition three samples from this subset of data (HSS-93, HSS-92 and HSS-1) have been rejected as they are relatively low Ti rocks (high Al/Ti and low Ti/V ratios). These three samples will be discussed in conjunction with the low-Ti aphyric lavas (HSS-95 and 49-J) in a following section. The equations of the regression lines and MgO axis intercept values (where applicable) are given in Table 5-8.

In Fig. 5-10 the plots generally show more scatter than the same plots for the aphyric lavas (as reflected by lower correlation coefficients in Table 5-7). This is in part, due to the cumulate rocks being able to form from a range of liquid compositions, as will later be illustrated from CaO-MgO and FeO-MgO plots. The regression lines are therefore less well constrained, however for the major element-MgO plots all the trends, on extrapolation, pass through or very close to the field of olivine compositions determined from the ultramafic cumulate samples. The Cr_2O_3-MgO plot is an exception and may be due to the accumulation of chromite crystals in many of the porphyritic samples.

The CaO-MgO trend in the porphyritic lavas contrasts strongly with that in the aphyric lavas. The best fit regression line to the selected data for the porphyritic samples (dotted line in Fig. 5-10) extrapolates through the low MgO portion of the olivine composition field. As there is a range of MgO magma compositions from which cumulate rocks could form, the area in CaO-MgO space in which the cumulate samples should plot, would be triangular, with the highest and lowest MgO aphyric lavas defining two of the apices and olivine the third. The resulting cumulate trend would not necessarily lie on a single straight line. In addition, as the aphyric lavas show a regular decrease in the CaO/Al_2O_3 and Ca/Ti ratios with increasing MgO content, the partial cumulate and cumulate samples would also be expected to show a range of values for these ratios, as olivine only contains very minor amounts of these elements. The fact that olivine accumulation does not fractionate these ratios, provides a fingerprint method for identifying which cumulates have been derived from specific liquids. Porphyritic samples that have similar CaO/Al_2O_3 ratios

should lie on a single olivine control line and the trend should intercept the CaO-MgO liquid trend (of the aphyric lavas) at a point where the liquids have similar CaO/Al_2O_3 ratios to the cumulates. Two additional regression lines have been calculated for the CaO-MgO data. All the samples in Table 5-4b that have CaO/Al_2O_3 ratios in the range 1.0-1.3 have been used to calculate the first regression line and the second was calculated from all samples with CaO/Al_2O_3 ratios in the range 1.6-1.9. The extrapolated trends (solid lines in Fig. 5-10) from both these regression lines pass through olivine. The 1.0-1.3 (CaO/Al_2O_3 ratio range) trend intercepts the aphyric lava CaO-MgO trend at 32.69% MgO and 4.59% CaO and the calculated Al_2O_3 content of the "liquid" at this point (3.27% Al_2O_3) gives a CaO/Al_2O_3 ratio of 1.40. The intercept of the 1.6-1.9 trend with the aphyric lava trend occurs at 28.46% MgO and 7.03% CaO and at this point the "liquid" contains 4.02% Al_2O_3 and has a CaO/Al_2O_3 ratio of 1.75. The good agreement between the predicted and observed CaO/Al_2O_3 ratios in the aphyric lavas at the interception points tends to support the conclusion made in the previous section that the CaO-MgO trend in the ultramafic aphyric rocks was not generated by alteration processes.

The field of cumulate compositions in the ultramafic porphyritic lavas has been modelled in detail for variations in FeO and MgO content in Fig. 5-11. The compositions of the olivines in equilibrium with the liquids were calculated assuming a $K_D = 0.37$. For small fractions of olivine crystallisation, the cumulate rocks must lie on a tie-line between the liquid with which the olivine is in equilibrium and the olivine composition. The field of possible olivine cumulate compositions is indicated in Fig. 5-11. If the cumulates formed from the progressive crystallisation of an initial liquid (i.e. the

FeO/MgO ratio of the liquid changed) then the cumulates would plot along curved lines. The scatter in the cumulate data obscures any such trends, although the majority of the porphyritic lavas plot on the low FeO, high MgO side of the aphyric lava FeO-MgO trend. This is consistent with most porphyritic lavas having been formed by the accumulation of olivine. A representative range of olivine compositions determined in the ultramafic cumulate samples is also plotted on the diagram. Not surprisingly the range of observed olivine compositions span the expected range, but do tend towards higher FeO contents. The olivines of higher FeO content than predicted, probably reflects the continuing crystallisation of olivine in the rocks after crystal movement in the magmas had ceased.

The element that should potentially be the most sensitive to olivine fractionation is Ni and in Fig. 5-12 the fields of aphyric and porphyritic lava compositions are outlined. The large fields indicate considerable scatter in the data. Regression lines (from Tables 5-7 and 5-8) have been plotted for the aphyric and porphyritic rocks. Three situations have been modelled in Fig. 5-12:-

Model A The first model assumes an initial magma composition of 11.24% FeO, 33.00% MgO and 1973 ppm Ni. Olivine was assumed to fractionate in batches of 5 wt.% with $D_{Ni}^{ol/liq}$ calculated from the equation given in Chapter 3 ($D_{Ni}^{ol/liq} = \frac{111.33}{MgO} - 1.71$). The FeO-MgO contents of the derived liquid and olivine were calculated assuming a $K_D = 0.37$. The derived liquid and olivine trends are also given in Fig. 5-12. This model has already been rejected as a means of obtaining the aphyric lava trends on the basis of the CaO-MgO and FeO-MgO contents of these rocks (Fig.

5-8, 5-9), but has been included here as two interesting features can be noted. Firstly,the evolving liquid trend is very slightly curved (this trend has been modelled down to liquids that contain 15% MgO) and as a first approximation could be considered a straight line. This approximate linear trend of liquid descent has interesting consequences for other lavas that have been proposed as high MgO liquids, such as those from Kilauea (Wright, 1971), and Baffin Island (Clarke, 1970). Hart and Davis (1978) have rejected the possibility that the linear correlation of the Ni-Mg data from Kilauea (given in Gunn, 1971) lavas, 8-20% MgO,could be generated by olivine fractionation from the postulated 20-25% MgO parent magma (Wright, 1971). The Ni-Mg trend that Hart and Davis (1978) modelled using their equation to calculate $D_{Ni}^{ol/liq}$ were markedly curved and displaced from the observed trend. They further state that the observed trend could be obtained by assuming an increase in the bulk D_{Ni} from ∿3 at the beginning of the trend to 10-15 at the end. The $D_{Ni}^{ol/liq}$ calculated for a magma with 20-25% MgO from the equation developed in Chapter 3 is 2.7-3.9 and at 8% MgO is 12.2, well within the range suggested by Hart and Davis (1978) for the Kilauea tholeiites. It therefore appears that the postulated high MgO liquids in this area cannot be discounted on the Ni-Mg trends alone and these rocks require further investigation to prove or disprove their existence. The second point of interest is that the most magnesian olivine does not contain the highest Ni content. This point will be discussed in model C.

Model B The second model assumes the same parameters as model A, except the equation to calculate $D_{Ni}^{ol/liq}$ from Hart and Davis (1978) has been used. The derived liquid trend in this model is curved and displaced to considerably lower Ni contents at given MgO, compared to model A. The olivine crystallising from the 33% MgO liquid should contain ∿5400 ppm Ni, considerably higher than the observed compositions and the olivine compositions calculated in model A. The olivines in model B show a regular decrease in Ni content with decreasing MgO. This model is however rejected on the basis of the Ni-Mg trend in the ultramafic aphyric lavas.

Model C This model differs from A and B in that the liquid compositions are not calculated but are constrained along the observed path. The composition of the olivines have been calculated at intervals of 1.0% MgO along the observed aphyric lava trend. The FeO and Ni contents of the presumed liquids at each MgO composition have been calculated from the regression equations (Table 5-7) and the compositions of the olivine calculated assuming a FeO-MgO K_D = 0.37 and $D_{Ni}^{ol/liq}$ calculated from the equation given in Chapter 3. The olivine has been removed in 5 wt.% batches. Tie-lines, along which the cumulates forming from the individual liquids should plot, have only been drawn in for the highest and lowest MgO liquid olivine pairs to avoid confusion on the diagram.

The olivines in model C show an initial increase of Ni content with decreasing MgO, reaching a maximum of 3540 ppm Ni in olivines crystallising from a liquid with ∿27% MgO and thereafter both Ni and MgO decrease in the olivines, crystallising from liquids with lower MgO contents. The olivines

analysed for Ni in the cumulate samples are also plotted for comparison. The Ni contents of the analysed olivines agree with the calculated olivines of similar MgO content to within 150-250 ppm Ni. Using smaller olivine fractionation steps of 1 wt.%, increase the Ni content of the calculated olivines by ~40 ppm Ni. The decrease of Ni content in the higher MgO olivines cannot be tested due to insufficient data, but overall the agreement between observed composition and calculated Ni content is considered to be reasonable.

Model C is therefore considered to be the best model for accounting for the porphyritic lava compositions and is consistent with the other major element data. However, these lavas show a considerably greater spread in composition than allowed in model C. This spread could in part be accounted for by the greater range of Ni compositions in the liquids from which the cumulates formed, than was accounted for in model C. Small amounts of sulphide minerals could also lead to some spread in the data and alteration processes that have affected the ultramafic rocks, may have redistributed Ni to some extent.

Throughout this section reliance has been placed on the equation from Chapter 3 ($D_{Ni}^{ol/liq} = \frac{111.33}{MgO} - 1.71$) relating the Ni concentration between olivine and liquid. Cawthorn and McIver (1977) have modelled the change in the partition coefficient ($D_{Ni}^{ol/liq}$) from 35-6.4% MgO in the lavas from the Barberton greenstone belt. Unfortunately many of the compositions they used are from rocks with partial cumulate textures (see Appendix 1) and not liquids as assumed in their models. Nevertheless good agreement is obtained between the $D_{Ni}^{ol/liq}$

calculated from the equation above and the model values from Cawthorn and McIver (1977) for high MgO liquids, as shown below. This provides additional support for the validity of the equation derived in Chapter 3 when dealing with komatiite lavas.

	This work	Cawthorn and McIver (1977)
35% MgO	1.5	1.7
24% MgO	2.9	3.2

The trace element-MgO plots for the porphyritic lavas generally show much more scatter and poorer correlation coefficients (Table 5-8) than the equivalent plots for the aphyric lavas. The variations are in general compatible with the olivine accumulation models developed for the major and minor elements. The Co-MgO trend shows a positive correlation indicating that $D_{Co}^{ol/liq}$ is greater than 1 for the ultramafic komatiites.

One feature that stands out in the porphyritic lavas is that the Rb content of the olivine-clinopyroxene phyric rocks is on average much higher than the olivine phyric and olivine rich rocks. It is unlikely that the Rb is contained in the clinopyroxene, as Hart and Brooks (1977) have recorded a range of 0.015-0.081 ppm Rb in clinopyroxenes extracted from Archaean igneous rocks. The reason for the high Rb contents in this class of ultramafic porphyritic lavas is not understood.

The inter-element ratios in general are consistent with the olivine accumulation model. The CaO/Al_2O_3 and Ca/Ti ratios show a wide range of values, but as recorded earlier in this section, variations in the CaO/Al_2O_3 ratios of the porphyritic rocks are consistent with the olivine accumulation model for the porphyritic lavas. The

Al/Ti and Ti/V ratios show a restricted range of values close to the average value for the aphyric lavas. The average porphyritic rock was calculated from the same data used to compute the regression lines and is given in Table 5-6. The average aphyric and porphyritic rock have been equally weighted to calculate the composition of the average ultramafic rock, and these data are also given in Table 5-6.

RELATIONSHIP BETWEEN APHYRIC AND PORPHYRITIC LAVAS

The two main points of the preceding sections are summarised below:-

1) Aphyric lava trends are generated by high pressure processes.
2) The porphyritic lavas formed from the range of aphyric lava compositions by olivine accumulation.

Several papers have appeared on the geochemistry of the ultramafic rocks from the LUU since the initial work carried out by the Viljoens. However, most of these have dealt with the problems of genesis of the ultramafic komatiites and not with their surface relationships.

Viljoen and Viljoen (1969d) have noted that flows varying in thickness from 12-30 meters, form the broad ultramafic zones in the Komati Formation type area. They suggest that some type of flowage differentiation, similar to that demonstrated experimentally, (Bhattacharji and Smith, 1964; Bhattacharji, 1967) concentrated the larger olivine crystals at the base of the flows. They further consider the possibility of invoking substantial settling of small olivine crystals in the flows an unlikely mechanism for accounting for the olivine rich cumulates. However, in this work it has been shown that olivine phenocrysts are sparsely distributed in the chilled margins

of flows and insufficient in content to account for the high MgO portion of some of the flows (e.g. Stuart's flow). Crystallisation and gravity settling of olivine crystals are considered the most likely processes for accounting for the porphyritic rocks.

Densities and viscosities of magmas can be calculated at different temperatures using the data given in Bottinga and Weill (1970, 1972). This enables the settling rates of olivine phenocrysts to be calculated from Stokes equation. Before olivine settling rates can be computed for the aphyric lavas, (Table 5-3a) estimates of their extrusion temperatures are required. From the experimental data given in Arndt (1976) and Green et al. (1975) olivine liquidus temperatures have been plotted (not shown) against MgO content of the liquid. A linear trend is observed (correlation coefficient = 0.99) between T and MgO and the equation of a least squares regression line fitted to these data is given below:-

$$T^{o}C = 18.79 \times MgO + 1042 \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (5\text{-}1)$$

A similar linear correlation has been noted by Thompson (1973) for published one-atmosphere liquidus temperatures plotted against MgO content for lavas with $\geq$13% MgO. The linear relationships between T and MgO as well as between $\frac{1}{MgO}$ and $D_{Ni}^{ol/liq}$ (from Chapter 3) imply a regular relationship between T and $D_{Ni}^{ol/liq}$. A plot of $\ln D_{Ni}^{ol/liq}$ against $\frac{1}{T^{o}K}$ gives a linear trend with a correlation coefficient of 0.99 for the data of Arndt (1977b) and Green et al. (1975). The equation of the least squares regression line is given below:-

$$\ln D_{Ni}^{ol/liq} = \frac{10691}{T^{o}K} - 4.93 \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (5\text{-}2)$$

It is beyond the scope of this work to consider the arguments as to whether $D_{Ni}^{ol/liq}$ is composition (Hart and Davis, 1978) or temperature dependent (Leeman and Scheidegger, 1977). However, it does appear from equations 3-2, 5-1 and 5-2 that the three parameters $D_{Ni}^{ol/liq}$, T and MgO content of the lavas are inter-dependent.

Equation 5-1 allows the one-atmosphere liquidus temperatures to be estimated for the aphyric lavas and these data are given in Table 5-3a. The multiple regression equations developed by French (1971) relating liquidus temperatures to liquid composition also allow the one-atmosphere olivine liquidus temperatures of the ultramafic rocks to be calculated. The temperature data obtained from the French (1971) equations have been compared to the equation 5-1 data in Fig. 5-13. Obviously, under ideal conditions, both calculations should give the same temperature for any particular sample and a plot of the temperature data should have a slope of 1.0. In Fig. 5-13 it can be seen that at high MgO contents ($T^{o}C$ >1400) equation 5-1 yield temperatures roughly 15^{o}C higher than the equations of French (1971), while at lower temperatures (<1250^{o}C) the data points scatter evenly around the ideal trend. The agreement between the two data sets is encouraging. French (1971) has indicated that his equations permit the liquidus temperature of the lavas to be obtained with a mean error of 11^{o}C.

The densities and viscosities for the aphyric lavas were calculated from the data of Bottinga and Weill (1970, 1972) and are also given in Table 5-3a. While the densities of the lavas can be calculated at any temperature in the range 1200-1700^{o}C, Bottinga and Weill (1972) have presented data for the viscosity calculations at 50^{o}C intervals. Therefore from the temperatures obtained for the

ultramafic lavas the viscosities and densities were calculated at the nearest 50°C interval.

The data for the ultramafic aphyric lavas (Table 5-3a) show that these lavas had very low viscosities at the one-atmosphere liquidus temperatures. In Fig. 5-14 the terminal settling velocities of spherical olivine crystals of various diameters have been calculated from Stokes' equation and plotted for a range of lava compositions. The composition of the olivines in equilibrium with each of the lavas was calculated assuming a FeO-MgO $K_D = 0.37$. The densities of these olivine compositions were calculated at 20°C assuming $\rho Fo_{100} = 3.222$ and $\rho Fa_{100} = 4.392$ from the equation:-

$$\rho_{(olivine)} = Fa \times 0.0117 + 3.222 \quad \ldots\ldots\ldots\ldots \quad (5\text{-}3)$$

The final density of the olivine was calculated at the liquidus temperature from the thermal expansion coefficient data given in Clark (1966) for $Fo_{89.9}$ at 1000°C.

For the range of ultramafic lavas presented in Fig. 5-14 very rapid settling rates of olivine phenocrysts are indicated. It is well known that particle shape affects settling velocities (Graf and Acaroglu, 1966). Flemming (1978) has reviewed the effects of particle shape on settling velocity of well defined geometric objects (such as spheres, cubes, octahedrons, tetrahedrons and so on) and notes that at Reynolds Numbers (Re)

($Re = \frac{WsD}{V}$ where Ws = settling velocity of sphere of diameter D and V = kinomatic viscosity = D/ρ)

less than one, the drag coefficient/Reynolds Number relationship follows Stokes' Law of settling irrespective of particle shape.

The Reynolds Number for sample HSS-109 becomes greater than 1 for olivine crystals with a diameter >3.0 mm. For the range of grain sizes and settling velocities given in Fig. 5-14 the Reynolds Number for all the samples is less than 1 and therefore the approximation that the olivine crystals are spherical will not significantly affect the calculated settling velocities.

The common olivine microphenocryst size range given for the chilled margins and pillow samples (Appendix 1) is from 0.1-0.3 mm. in diameter, allowing settling rates from 3.3 to 80 cm/hr. in ultramafic liquids. Olivine crystals in the cumulate samples approach 1 mm. in diameter in size and the settling velocity of these crystals in the liquids would be in the range of 330 to 890 cms/hr. The average settling velocity of a growing crystal in an ultramafic magma would obviously be between these two extreme ranges, nevertheless even small olivine crystals could settle through considerable distances in ultramafic flows in relatively short times. It is therefore concluded, that regardless of whether mechanisms of flow differentiation were in operation in the ultramafic flows, once the magma had come to rest, any olivine phenocrysts in the flow could settle to near the base of thin flows in very short periods of time. The rapid settling rates indicated for olivine phenocrysts in ultramafic magmas provide an adequate mechanism for obtaining the porphyritic lavas by olivine accumulation in flows, dykes and sills from the range of initial magma compositions (as represented by the aphyric lavas).

The viscosity and settling velocity data should enable rough estimates of the time it took for flows to cool to a stage where crystal movement in the flow had ceased. This has been modelled for Stuart's flow based on the following assumptions:-

1) The initial diameter of olivine microphenocrysts was 0.2 mm. and final diameter of the settling olivine crystals was 0.6 mm.

2) ρ and γ change linearly with MgO and that the initial liquid had a viscosity of 2.81 poises (HSS-531/2, MgO = 27.88) and the final liquid had a viscosity of 3.54 poises (HSS-533, MgO = 25.65).

3) The volume of the crystals increased linearly with the distance settled. The initial crystallising composition of olivine being Fo_{92} and the final portion of olivine to crystallise being Fo_{91}.

In Stuart's flow there is a 40 cm. thick upper chilled margin, underlain by 100 cms. of plate spinifex. Donaldson (1977) has noted that a nuclei free zone is required in order for spinifex texture to develop. Any phenocrysts in the magma at the base of the chilled margin must have settled faster than the advancing front of crystallising spinifex textured olivines, in order to leave a nuclei free zone in the spinifex growth area. The time required, therefore, for a growing olivine phenocryst to settle from the base of the upper chilled margin to the upper portion of the B zone (a distance of 100 cms.) will indicate very roughly how long it took the flow to crystallise under closed system conditions with the magma remaining stagnant. The settling rate of the olivine crystals has been calculated for 5 cm. intervals with the average crystal radius, density of liquid and crystal and viscosity of the magma for the 5 cm. interval being used in Stokes' equation. From this average settling velocity the time taken to settle through each

5 cm. interval was calculated. The integrated time required under these conditions for the crystal to settle through 100 cms. is 6.8 hours. This crude estimate of the crystallisation time of the flow is consistent with the ultramafic magmas being extruded into a sub-aqueous environment and therefore being cooled rapidly by the water.

LOW-Ti ULTRAMAFIC LAVAS

This group of lavas includes the two spinifex textured samples (HSS-95 and 49-J) and the three porphyritic rocks (HSS-93, HSS-92 and HSS-1). The geochemical parameters that set these lavas apart are their low TiO_2 contents and high Al/Ti and low Ti/V ratios compared to other ultramafic samples of equivalent MgO content. The two spinifex textured samples have very similar CaO/Al_2O_3 ratios in marked contrast to the trend of decreasing CaO/Al_2O_3 ratio with increasing MgO content of the major aphyric lava trend. Nesbitt et al. (1979) have also pointed out that 49-J is not typical of the main body of ultramafic komatiite data from the LUU, based on inter-element ratios and REE patterns.

In Fig. 5-15 selected elements have been plotted against MgO content for this group of lavas. An olivine control line from 50% MgO passes through the two spinifex textured samples (including the CaO-MgO data) suggesting that olivine could be the only mineral controlling the compositions of this group of lavas. The three porphyritic samples also plot near to the olivine control line trend and could have been generated by olivine accumulation from liquids similar in composition to the two spinifex samples.

The data in Tables 5-3b and 5-4b show that in addition to the low TiO_2 contents of these lavas, Zr and Y are also lower than the more abundant group of samples of equivalent MgO. The lower contents

of these incompatible elements indicate that the low-Ti group of lavas was derived from a source area depleted in Ti, Zr and Y, compared to the source area of the 'normal' group of ultramafic rocks. In Chapter 7 the relatively rare low-Ti ultramafic komatiites are referred to as group I rocks while the more abundant relatively higher Ti ultramafic komatiites are referred to as the group II rocks. Further implications of the differences in chemistry between these two groups for mantle heterogeneity are discussed in Chapters 7 and 8.

SANDSPRUIT FORMATION ULTRAMAFIC ROCKS

The major and trace element data (recalculated volatile free and with the $FeO/Fe_2O_3 = 0.2$) are given in Tables 5-5a and 5-5b respectively. The samples have been arranged in order of increasing MgO content and no aphyric-porphyritic division has been made for these samples, as no original igneous textures are preserved. The average of the 9 Sandspruit Formation ultramafic rocks is given in Table 5-6. This average composition is similar to that obtained for the average ultramafic rock from the Komati and Theespruit Formations. Because of the limited number of samples from the Sandspruit Formation, small differences such as slightly higher MgO and lower CaO and Al_2O_3 contents cannot be considered as significant. However, the Al/Ti ratio is lower and the Ti/V ratio higher than the average ultramafic rock. These differences in the ratios are in part due to the very high TiO_2 contents of three of the Sandspruit Formation ultramafic rocks (87-J, 88-J and HSS-131).

The variations in chemistry of the Sandspruit Formation ultramafic rocks are illustrated in Fig. 5-16 where selected elements have been plotted against MgO content. The regression lines obtained

for the aphyric lavas (24-33% MgO), and porphyritic lava (30-45% MgO) from the two upper Formations of the LUU have also been drawn in for comparison. The Sandspruit Formation ultramafic rocks show very similar trends to those from the upper Formations of the LUU. Several of the Sandspruit Formation ultramafic samples however do have much higher TiO_2 contents, but these same samples do have very similar Al_2O_3, FeO, CaO and V contents to the rest of the LUU samples.

McIver and Lenthal (1973) and McIver (1975) have concluded from consideration of the ultramafic komatiite data in Viljoen and Viljoen (1969b,d), that the Sandspruit Formation ultramafic rocks are primary and those from the Komati Formation are enriched in olivine. However, from Fig. 5-16 it appears that there is little justification for this conclusion, as the Sandspruit Formation ultramafic rocks show similar ranges in composition and define similar trends, even though the former have been metamorphosed to a greater extent. It is therefore concluded that although a few ultramafic samples from the Sandspruit Formation have higher TiO_2 contents compared to samples of similar MgO content from the Theespruit and Komati Formations, the compositions of the two groups of ultramafic lavas are essentially identical with respect to the trends displayed and contents of individual elements.

5-5 LUU KOMATIITES COMPARED TO OTHER GREENSTONE BELT ULTRAMAFIC ROCKS

RHODESIA

Recent work on the Belingwe greenstone belt (Bickle et al., 1975; Bickle et al., 1976; Nisbet et al., 1977) has revealed good exposures of ultramafic lavas inter-bedded with komatiitic and

tholeiitic volcanics of lower MgO content. The outcrop of the ultramafic lavas show many similar features to those from the LUU and include pillows, massive flows, spinifex (olivine) textured horizons and ultramafic ridges. In thin section olivine appears to be the first and major crystallising phase followed by clinopyroxene (Nisbet et al., 1977), while orthopyroxene is rarely observed in the ultramafic lavas. Nisbet et al. (1977) consider the ultramafic ridges to have formed by accumulation of olivine towards the base of thick flows or sills although they consider olivine fractionation in thin flows (2-12 m. thick) to be minimal or not to have occurred at all.

Bickle et al. (1975) and Nisbet et al. (1977) have presented major and trace element data for the ultramafic lavas from the Belingwe greenstone belt and suggest that phenocryst free magma compositions containing 20-32% MgO had occurred. The similar outcrop and textural forms, mineralogy and range of liquid compositions are convincing evidence for considering the Belingwe greenstone belt ultramafic rocks as belonging to the same suite of rocks as those from the LUU. However, there are differences in many of the geochemical parameters (Nisbet et al., 1977; Arndt et al., 1977) when the data are examined in detail, such as the lower CaO/Al_2O_3 ratios of the Belingwe ultramafics.

In order to compare the LUU and Belingwe data, ultramafic lava compositions (which are considered to represent phenocryst free magmas) have been selected from Bickle et al. (1975) and Nisbet et al. (1977) based on the petrographic descriptions of the Belingwe ultramafic rocks in Nisbet et al. (1977). Pillows and spinifex textured rocks were automatically selected. However, many samples are described as spinifex textured with some equant (filled) olivine crystals present.

As no indication was given of the proportion of the latter morphological type of olivine in the samples, they were rejected as representing liquid compositions. Eleven samples were selected (sample nos. given in Table 5-9) and the three key elements TiO_2, Al_2O_3 and CaO plotted against MgO in Fig. 5-17 with the regression line trends from the LUU aphyric lavas for comparison.

The Belingwe ultramafic lavas are depleted in TiO_2 and CaO and enriched in Al_2O_3 compared to the LUU trends. The most striking differences between the two CaO-MgO sets of data are the significantly different slopes and MgO intercept values. The low-Ti ultramafic lavas from the LUU, on the other hand, show very similar slopes and MgO intercept values as well as absolute CaO content to the Belingwe ultramafic lavas, however they are depleted in TiO_2 and Al_2O_3 compared to the Belingwe rocks.

Regression lines have been fitted to the Belingwe data and the slopes and intercept values given in Table 5-10. The data for the CaO-MgO trend differ markedly from that given in Nisbet et al. (1977). The reasons for this are unknown, particularly as the TiO_2-MgO and Al_2O_3-MgO regression line data are similar. The MgO intercept values for the three trends vary from 47.3 to 55.7% MgO making the possibility, that olivine-only fractionation controls the trends, unlikely. Nisbet et al. (1977) have also come to the same conclusion based on the Cr_2O_3, TiO_2 and Al_2O_3 variations in the Belingwe ultramafic lavas.

The LUU and Belingwe ultramafic lavas therefore show broadly similar ranges in MgO contents and a similar lack of olivine control on the 'liquid' trends, but considerable variation in detail when the geochemistry of the lavas are compared. As these two groups of

lavas formed in similar Archaean volcanic environments the processes involved in the genesis of the magmas were probably similar. The detailed geochemical differences therefore probably reflect regional variations in mantle composition; however this aspect will be considered in more detail in a later section.

WESTERN AUSTRALIA

The occurrence of ultramafic and mafic volcanics from Western Australia has been well documented in the literature although the exposure of the lavas, apparently, is not as good as found in Barberton (Anhaeusser, 1971b). Nevertheless many spinifex textured ultramafic flows have been described (Nesbitt, 1971; Naldrett and Turner, 1977) which have similar features to those from the LUU, such as fine grained random (or radiating, Nesbitt's 1971 term) spinifex zones overlying plate spinifex zones, the latter being abruptly terminated by the underlying olivine peridotite. No pillow structures have as yet been recorded in rocks of ultramafic composition from Western Australia, although sill and lens like bodies have been investigated (Hallberg and Williams, 1972; Williams, 1972; McCall, 1973) and show many similar features to those occurring in the Barberton greenstone belt. The crystallisation sequence of olivine followed by clinopyroxene with a scarcity or total absence of orthopyroxene (Williams, 1972; Nesbitt and Sun, 1976; Naldrett and Turner, 1977) is also the same as that observed in the LUU flows and dykes of ultramafic composition.

Analyses of spinifex textured ultramafic rocks from Western Australia have been selected from the literature (see Table 5-9 for source references and sample nos.) for comparison with the LUU aphyric

lavas. The Western Australian rocks display a range of compositions from 24-35% MgO while two samples (that have been described as spinifex textured peridotitic komatiites from the Marshall Pool area, Nesbitt et al., 1979) have lower MgO contents of 19.2 and 21.2% MgO. All the analyses of the selected Western Australian samples have been calculated volatile free and with total Fe as FeO. The three elements TiO_2, Al_2O_3 and CaO have been plotted against MgO (Fig. 5-18) and the aphyric lava trend lines from the LUU, drawn in for comparison.

The TiO_2-MgO trends are similar in both sets of lavas while the LUU lavas have significantly lower Al_2O_3 contents than the Western Australian rocks. However, three samples from the Pilbara area (Nesbitt et al., 1979) show a similar low Al_2O_3 content to the LUU lavas and in addition, these same three samples plot close to the LUU trends for TiO_2-MgO and CaO-MgO. These samples also have similar CaO/Al_2O_3, Al/Ti and Ti/V ratios, Zr, Y, Ni, V and Sc contents to LUU rocks of similar MgO content. The CaO-MgO trends for the LUU and Western Australian lavas show broadly similar contents of CaO at given MgO, but the LUU trend has a significantly steeper slope.

Nesbitt and Sun (1976) and Naldrett and Turner (1977) have suggested that olivine is the only mineral controlling the evolution of the ultramafic liquid compositions. However, regression lines fitted to the Western Australian data (Table 5-10 and Fig. 5-18) show that the Al_2O_3-MgO trend does not extrapolate through likely olivine compositions and therefore, as with the LUU and Belingwe ultramafic lavas, the trends are not simply controlled by olivine fractionation alone. Although it should be noted that because the Al_2O_3-MgO data for the Western Australian rocks show more scatter than the other two areas the MgO intercept value is not as well constrained.

MUNRO TOWNSHIP

Ultramafic komatiites from the Munro Township area in northeast

Ontario have been well documented (Pyke et al., 1973; Arndt, 1975; Arndt et al., 1977; Arndt, 1977a; Arth et al., 1977; Arndt, 1979) and occur in close association with flows and sills of mafic komatiites and tholeiitic basalts. The detailed structures and textures in the ultramafic flows from Munro Township are very similar to those found in the LUU (see Chapter 3 for discussion). Settling of olivine phenocrystsin the ultramafic flows has been suggested (Pyke et al., 1973; Arndt et al., 1977) as the main process for accounting for the cumulus enriched B portion of the flow sections. The crystallisation sequence of the ultramafic rocks (at low pressure) has been well established both in the natural samples and in the laboratory (Arndt, 1975; Arndt, 1976; Arndt et al., 1977) and shows the now familiar sequence of olivine first, joined by clinopyroxene at a much lower temperature, with orthopyroxene playing little or no part in the sequence.

Once again ultramafic rocks representative of phenocryst-free magma compositions have been selected from the literature; sources of data and sample nos. are given in Table 5-9. TiO_2, Al_2O_3 and CaO have been plotted against MgO content (Fig. 5-19) after the data had been calculated volatile free and with total Fe as FeO. The LUU aphyric lava trends have been drawn in for comparison. Regression lines have also been calculated for the trends from the Munro Township samples and the equations given in Table 5-10.

The LUU lavas are enriched in TiO_2 and depleted in Al_2O_3 compared to the Munro Township samples. One spinifex textured sample (SA-1160 from Pyke et al., 1973) has a significantly higher Al_2O_3 content compared to the main trend and was eliminated from the Al_2O_3-MgO regression line calculation. However, the most outstanding

feature of Fig. 5-19 is that the CaO-MgO trends for the two groups of lavas are very similar with respect to both the absolute CaO contents and MgO intercept values and demonstrates that this trend is not unique to the Barberton greenstone belt ultramafic rocks. The three trends for the Munro Township ultramafic rocks in Fig. 5-19 all intercept the MgO axis at considerably lower values than the likely olivine compositions in equilibrium with the liquids and therefore olivine control (by crystal fractionation or melting) on the composition of the Munro Township lavas is considered unlikely.

REGIONAL VARIATIONS OF ULTRAMAFIC LIQUID COMPOSITIONS

The gross features of the ultramafic komatiites from the different areas such as geological setting, structures, textures, mineralogy and geochemical composition are compelling evidence for considering the lavas from these areas as belonging to the same suite of rocks. Now that the major element geochemistry from other greenstone belts is reasonably well characterised and the search for 'super' komatiites waning, it may be more constructive to consider the similarities as well as emphasising the differences between the LUU and ultramafic komatiites from other greenstone belts.

To summarise the foregoing discussion, the following similarities have been noted:-

a) Ultramafic lava compositions that are considered to represent phenocryst free magmas have MgO contents mainly in the range 20-34% MgO.

b) Olivine is the only mineral that is observed to crystallise and accumulate in the ultramafic lava flows.

c) The presumed liquid trends are not controlled by olivine as the only fractionation or melting phase and therefore the liquid trends must be generated by high pressure crystal fractionation or melting processes involving additional phases.

d) The TiO_2-MgO, Al_2O_3-MgO and CaO-MgO trends are essentially linear within the scatter of the data from the four areas considered.

This last point has played a major role in eliminating olivine-control on the evolution lines of the composition of the ultramafic magmas. If olivine was the only mineral crystallising from the liquids in the range of say, 33-24% MgO, a spread of olivine compositions would be involved and the resulting incompatible element versus MgO trend would be slightly curved. The effects of this slight curvature on a straight line regression have been modelled assuming a starting composition of 33% MgO, 11% FeO and incompatible element content of 4.4% ($X_{inc.}$). Olivine was removed in 5% batches ($K_D^{FeO-MgO} = 0.37$) and the MgO and $X_{inc.}$ content of the liquid calculated after each batch removal. The process was modelled down to MgO = 23.8% ($X_{inc.} = 6.63\%$) and the MgO composition of the olivine ranged from 52.05 - 49.30%. A straight line fitted to the $X_{inc.}$-MgO data points gave a correlation coefficient of 0.9998 (i.e. a straight line for all practical purposes) and an MgO intercept value of 51.0% MgO. Clearly the slight curvature of the trend would not obscure the olivine control lines if this process had been in operation in the ultramafic magmas.

The geochemical differences for key elements between the

ultramafic magmas have been summarised in Fig. 5-20 where the TiO_2-MgO, Al_2O_3-MgO and CaO-MgO regression lines for all the trends have been plotted. The Belingwe and Western Australia rocks show similar trends in all the diagrams while the Munro Township and LUU rocks have very similar CaO-MgO trends which have much steeper slopes than the CaO-MgO trends from the other two areas. The feature that singles out the LUU lavas is their very low Al_2O_3 contents which accounts for the high CaO/Al_2O_3 ratios for this group compared to the other areas (Table 5-11). However, comparing average CaO/Al_2O_3 ratios for the ultramafic lavas from the four areas is an over-simplification, as the CaO/Al_2O_3 ratio decreases with increasing MgO content for all the areas (as calculated from the regression equations for CaO-MgO and Al_2O_3-MgO). This effect is most marked for Munro Township where the CaO/Al_2O_3 changes from 1.1 to 0.58 in the MgO range from 20 to 35% MgO, the change in CaO/Al_2O_3 ratio for the LUU aphyric lavas in the same MgO range is from 2.2 to 1.1. This change in ratio with increasing MgO content is considered to be as important as the absolute CaO/Al_2O_3 ratio for characterising the ultramafic komatiites and in this respect the LUU lavas are not unique.

Further differences between the groups of ultramafic lavas are illustrated in Table 5-11 where several average inter-element ratios are compared. The average values for Belingwe, Western Australia and Munro Township have been computed from the same sources of data (where available) as used for the plots and given in Table 5-9. The average Al/Ti, Ti/Zr and Ti/V ratios for the Belingwe, Western Australia, Munro Township and the low-Ti LUU (group I) ultramafic lavas show very similar values suggesting the lavas formed from

similar source compositions. The common (group II) LUU ultramafic lavas have lower Al/Ti and Ti/Zr ratios and higher Ti/V ratios. From consideration of these inter-element ratios and the trends in Fig. 5-20, the group II LUU lavas are depleted in Al and enriched in Ti and Zr compared to the other groups of lavas.

The geochemical features shown in Fig. 5-20 and Table 5-11 serve to demonstrate that, in detail, the geochemistry of the ultramafic komatiites differs markedly between the different greenstone belts, except perhaps for the Belingwe and Western Australia areas. Furthermore, individual samples such as those from the Pilbara area or the low-Ti group I lavas from the LUU indicate that certain aspects of the regional geochemical variations can be duplicated in a single greenstone belt. If the ultramafic komatiite liquid trends as represented by the aphyric lavas are generated by partial melting in the mantle, then this variation indicates significant mantle inhomogeneity in the Archaean, both on a regional and local scale. Further evidence for mantle heterogeneity is considered in Chapter 7.

5-6 CONCLUSIONS

In conclusion of this section the following points resulting from the interpretation of the textural, mineralogical and geochemical data of the LUU ultramafic komatiites are stressed:-

1) The ultramafic rocks are readily divided into samples representing phenocryst free magma compositions (aphyric lavas) or partial cumulates (porphyritic lavas) on the basis of their petrography.

2) Olivine is the dominant crystallising phase from the magmas at low pressure and joined by clinopyroxene

at lower temperatures.

3) The aphyric lava trends are not developed by low pressure olivine fractionation and as such must be developed by high pressure crystal fractionation or melting processes.

4) The porphyritic rocks are developed by olivine settling and accumulation from the range of aphyric lava compositions.

5) The calculated range of temperatures at which the ultramafic magmas were extruded varies from 1500-1660^{o}C and from the predicted viscosities of these lavas the terminal settling velocity of the olivine phenocrysts is in the order of meters/hour.

6) Two groups of ultramafic lavas have been identified in the LUU. The more abundant group (referred to as group II rocks in Chapters 7 and 8) and the low-Ti group (group I rocks).

7) Metamorphism has destroyed the igneous textures in the Sandspruit Formation ultramafic rocks, but the range of compositions and inter-element trends show them to be similar to the better preserved samples from the Theespruit and Komati Formations.

8) Comparison of the LUU aphyric lavas with similar lavas from other greenstone belts shows that the LUU rocks are depleted in Al_2O_3 and slightly enriched in TiO_2 compared to the Belingwe, Western Australia and Munro Township lavas.

9) The ultramafic lavas from the four areas considered, all show decreasing CaO/Al_2O_3 ratios with increasing MgO content, although this effect is most marked in the Munro Township and LUU lavas.

10) Regional and local variations in some aspects of the geochemistry of the ultramafic lavas probably reflects inhomogeneous source areas for the lavas.

CHAPTER 6.

GEOCHEMISTRY OF MAFIC KOMATIITES

6-1 INTRODUCTION

Viljoen and Viljoen (1969c) identified three types of basaltic komatiites from the Barberton greenstone belt which covers a range of compositions from 9 - 22% MgO and with CaO/Al_2O_3 ratios generally, but not exclusively, greater than unity. In this work (see Chapter 4) a two-fold division of the 9 - 24% MgO komatiites from the LUU is proposed, viz. high-Mg mafic komatiites and low-Mg mafic komatiites. Details of the chemical parameters used to sub-divide the lavas, are given in Chapter 4. The low-Mg mafic komatiites incorporate the Barberton type and Badplaas type of basaltic komatiites proposed by Viljoen and Viljoen (1969c) while the high-Mg mafic komatiites are essentially equivalent to their Geluk type of basaltic komatiite.

In the present Chapter, field, textural and mineralogical features of the lavas are briefly described. The geochemical data are discussed in relation to the low pressure processes that may have contributed to generating the observed geochemical trends. Discrepancies are noted between the observed phenocryst assemblages and the liquidus phases predicted for the lavas at one atmosphere pressure. These differences have an important bearing on proposed models of genesis for the mafic komatiites. Finally the Barberton greenstone belt komatiites are compared to similar rock types from Belingwe, Western Australia and Munro Township areas.

6-2 SAMPLE LOCATION, FIELD CHARACTERISTICS AND PETROGRAPHY

The majority of mafic komatiite samples are from the LUU (specifically from the type areas of the Komati, Theespruit and Sandspruit Formations)

but others are from widely scattered areas within the Barberton greenstone belt, such as the Helshoogte pass, Kaapmuiden-Malelane area and Steynsdorp valley. Sample localities in the Komati Formation type area are shown in Figures 6-1 and 6-2. Several samples have also been analysed from the Upper Mafic to Felsic Unit (UMFU).

Sample localities and brief petrographic descriptions for all the mafic komatiites analysed for this study are given in Appendix 1. These samples include rocks analysed by Viljoen and Viljoen (1969b,c,e) for major elements. Rocks collected by the author are prefixed by HSS, SC, SD or HK while the balance of the samples are from the Viljoen's collection. A sample of olivine gabbro (HK-4) from the Koedoe body (Viljoen and Viljoen, 1970b) was analysed for olivine, pyroxene and plagioclase mineral compositions by electron microprobe.

Both types of mafic komatiite crop out as massive flows, sometimes with a flow top breccia or as pillowed horizons (detailed descriptions of pillows are given in Chapter 2). Ocelli varying in size from a few mms. to several cms. across, are common in the pillows and flows. These features are discussed in detail in Chapter 2, where it is suggested that they were formed by liquid immiscibility. In addition several coarse grained rocks are from outcrops where clear field relationships could not be discerned. These samples may be portions of sills, dykes or lower zones of thick flows.

Clinopyroxene blades and needles forming macroscopic spinifex textured layers in basaltic komatiite flows (similar to that formed by olivine in ultramafic komatiite flows) are frequently observed from other greenstone belts (Arndt et al., 1977; Arndt, 1977a; Nisbet et al., 1977; Sun and Nesbitt, 1978; Nesbitt et al., 1979), but they have not been recorded in lavas from the Barberton greenstone belt. However,

Viljoen and Viljoen (1969c) have suggested that zones, lying parallel to the general layering of the Komati Formation lavas, which contain long (several cms.) amphibole blades, could represent recrystallised quench textures. Williams and Furnell (1979) consider that these zones represent clinopyroxene spinifex textured portions of mafic komatiite flows. One sample (SD-42C) from a zone bearing coarse-grained amphibole blades from the Komati Formation, has been analysed. This sample contains relatively short (<2 cms.) actinolite blades and is considered to represent a metamorphically recrystallised pyroxene spinifex textured rock. The observed or inferred igneous mineralogies and textures of the mafic komatiites, are discussed below.

HIGH-Mg MAFIC KOMATIITES

Fine-grained phenocryst-free rocks (aphyric), containing 20 - 24% MgO, have been classified as high-Mg mafic komatiites. These rocks contain abundant pyroxene crystals as needles or blades, often aligned, and forming a micro-spinifex texture. The pyroxene crystals have usually been altered to amphibole. Some samples contain small amounts of pyroxene microphenocrysts that have been altered to amphibole and talc. Spinel microphenocrysts are also present and in some instances have been replaced by magnetite. These aphanitic samples are considered to be quenched liquids and to be representative of essentially phenocryst-free magma compositions.

Samples that have porphyritic textures and MgO contents between 20 - 25% MgO have also been analysed. These samples contain abundant equant pyroxene phenocrysts (now altered to amphibole) that often enclose small rounded inclusions of serpentine. The matrix is usually fine grained amphibole, chlorite, serpentine, minor epidote, sphene

and opaques. Actinolite crystals (twinned) have pseudomorphed phenocrysts, inferred to have been clinopyroxene by analogy with the alteration styles of pyroxenes in mafic rocks from the Mount Monger area (Williams, 1972). Rarely the amphibole replacing the pyroxenes is of a distinctive fibrous nature; these crystals may originally have been orthopyroxene (cf. Williams, 1972). Serpentine pseudomorphs occurring as inclusions in pyroxene and occasional serpentinized crystals in the matrix may originally have been olivine. However, although these pseudomorphs contain minor associated magnetite, they lack the conspicuous dusty magnetite rims that are observed in altered olivines from the ultramafic rocks.

While the similar range of MgO contents of the aphyric and porphyritic high-Mg mafic komatiite lavas might be taken to indicate a close association of these texturally distinct groups, certain differences in chemistry suggest that they are not simply related. These differences will be illustrated in a later section.

LOW-Mg MAFIC KOMATIITES

Following Chapter 4, fine-grained aphyric lavas with MgO contents in the range 9 - 16% MgO, are termed low-Mg mafic komatiites. These lavas are believed to represent compositions of phenocryst-free magmas and consist of samples from pillows, flowtop breccias and basal chilled margins. In thin section the aphyric lavas contain clinopyroxene as small equant crystals or laths, both of which have commonly been replaced by actinolite. Clinopyroxene blades forming microspinifex texture are also common. The sample (SD-42C) discussed previously, with relatively short actinolite blades (<2 cms.), is considered to represent an altered clinopyroxene spinifex textured

rock. Plagioclase is occasionally observed as a minor phenocryst phase. The groundmasses of the aphyric samples consist of fine-grained amphibole, chlorite, minor epidote and sphene. Small microphenocrysts of spinel are scattered through the samples and often occur as inclusions in the clinopyroxene cyrstals. In many of the samples, grains of fresh or only partly altered clinopyroxene occur. Olivine has only been identified as occasional small hopper crystals in some of the chilled margin samples.

Many samples have been analysed which have similar compositions to the aphyric low-Mg mafic komatiite lavas, but they cannot be considered as representing quenched liquid compositions. Some of these rocks are medium to coarse grained, while others are porphyritic and in some instances demonstrably enriched in phenocrysts. The porphyritic samples show a gradation of phenocryst types from being predominantly clinopyroxene-phyric through to having clinopyroxene and plagioclase as the major phenocryst phases. Spinel microphenocrysts are commonly scattered throughout the samples, and the groundmass consists of fine-grained amphibole, chlorite, plagioclase, minor quartz and very minor epidote and sphene. The coarse-grained samples consist of clinopyroxene crystals (up to 1 mm. long) with large, 3 - 4 mm. interstitial plagioclase grains, usually altered. Skeletal ilmenite crystals are present along with minor amounts of spinel and occasionally brown hornblende. For ease of reference, both the porphyritic and coarse-grained samples will be referred to as porphyritic low-Mg mafic komatiites.

The final group of rocks with compositions in the low-Mg mafic komatiite range are mainly from the Sandspruit Formation or areas close to granite contacts. These rocks are recrystallised but pillow

structures and ocelli, although strongly deformed, can often still be recognised. The mineralogy of the samples consists of interlocking grains of fresh actinolite with minor amounts of quartz, plagioclase, pyroxene and sphene. The quartz and plagioclase contents decrease with increasing MgO content. Chlorite, if present, is very minor. As with the ultramafic rocks that display metamorphic textures, the geochemistry of these samples will be discussed separately.

6-3 MINERAL COMPOSITIONS

Mineral compositions determined from the mafic komatiite lavas and the sample HK-4, are presented in Table 6-1. The only olivine compositions determined are from the olivine gabbro sample HK-4, taken from the thin (12 m.) anorthositic gabbro-norite zone in the Koedoe body. This horizon is overlain by the 610 m-thick zone of the upper dunite sequence (see Viljoen and Viljoen, 1970b). The olivines from HK-4 have a restricted range of compositions and slightly lower Fo contents on average, than those from the ultramafic komatiite lavas, but do have significantly lower Cr_2O_3 and NiO contents (compare olivine analyses in Tables 5-1 and 6-1). The lower NiO and Cr_2O_3 contents probably reflect the more differentiated nature of the residual magma in the Koedoe body at the time the HK-4 olivines crystallised, compared to magmas from which olivines in the ultramafic flows crystallised.

Much effort was expended on attempting to obtain pyroxene compositions from porphyritic high-Mg mafic komatiites, but the poor totals of analyses (not reported) reflect the altered nature of the pyroxenes in these samples. However, one satisfactory analysis of a clinopyroxene was obtained from sample HSS-10 (see Table 6-1). This

plots in the augite field on the pyroxene quadrilateral (Fig. 6-3) and is quite distinct from clinopyroxene compositions found in the ultramafic rocks, HK-4, or the mafic komatiites. The significance of this pyroxene of relatively unusual composition (for komatiites) cannot be assessed at this stage due to a lack of data. In general clinopyroxenes from mafic komatiites and HK-4 have similar compositions to those found in ultramafic komatiites and mainly plot in the endioside field on the pyroxene quadrilateral (Fig. 6-3), although clinopyroxenes from the ultramafic komatiites have slightly higher TiO_2 and NiO contents. Perhaps the most significant feature of the clinopyroxenes is the very restricted range in compositions observed in the komatiite lavas, regardless of the MgO content of their host rocks. The Al_2O_3 content of the mafic komatiite clinopyroxenes are significantly lower than those found in clinopyroxenes from similar lavas in the Munro Township area (Arndt et al., 1977).

Plagioclase crystals analysed in HK-4 are uniform in composition with a high An content (Table 6-1). No plagioclase compositions have been determined from the mafic komatiite lavas as this mineral is usually altered. However, as the olivine and clinopyroxene compositions in HK-4 are similar to those found in the lavas, it is expected that the plagioclase compositions would be of a similar calcic nature.

6-4 GEOCHEMISTRY OF THE MAFIC KOMATIITES

The major element compositions, normative mineral contents, trace element concentrations and inter-element ratios are given for the mafic komatiites in the following tables:

Aphyric high-Mg mafic komatiites	- Tables 6-2a and 6-2b
Porphyritic high-Mg mafic komatiites	- Tables 6-3a and 6-3b
Aphyric low-Mg mafic komatiites	- Tables 6-4a and 6-4b
Porphyritic low-Mg mafic komatiites	- Tables 6-5a and 6-5b
Texturally altered mafic komatiites	- Tables 6-6a and 6-6b

The major and trace element data have been recalculated volatile free and the FeO and Fe_2O_3 contents calculated such that the Fe_2O_3/FeO ratio = 0.2. The original totals, loss on ignition (LOI) and $H_2O^+ - CO_2$ contents (where determined) are also reported. The major element compositions from Viljoen and Viljoen (1969c,e) have also been given for samples from their collection that have been analysed for trace elements. The compositions of the better preserved pillow samples discussed in Chapter 2 have been averaged and these data presented in Tables 6-4a and 6-4b.

From the study into the effects of alteration on the geochemistry of mafic komatiite pillow lavas (Chapter 2), the elements determined have been placed into three groups. Those elements in group A (Si, Ti, Al, P, Nb, Zr, Y, Co, Cr, V and Sc) are believed to have remained relatively immobile during the various alteration processes that have affected the mafic komatiite pillow lavas. The group B elements (Fe, Mn, Mg, Ca, Ni and Ga) are also thought to have remained relatively immobile in samples that still have some relict igneous mineralogy and texture. While the concentrations of the group C elements (Na, K, S, Rb, Sr, Zn, Cu and Ba) are believed to have been changed by the alteration processes that affected the LUU lavas. The group A and B elements will be used here to interpret the geochemistry of the mafic komatiites, as these elements should reflect the original

igneous characteristics of the lavas. In the following discussion the terms high-Mg mafic komatiites and low-Mg mafic komatiites will be abbreviated to high-Mg MKs and low-Mg MKs.

APHYRIC HIGH-Mg MAFIC KOMATIITES

The number of analyses of the high-Mg MKs is limited due to the generally poor textural and mineralogical preservation of these rocks and to their scarcity in the better preserved areas of the LUU. The compositions of the aphyric high-Mg MKs are in the arange of 20.4 to 22.9% MgO with relatively high SiO_2 contents. In contrast large variations, roughly by a factor of 2, are observed for other oxides and elements such as TiO_2, Al_2O_3, CaO, Zr, Ni, V and Ga (see Table 6-2a and 6-2b). The variation of these elements are also coupled with large variations in their inter-element ratios such as Al/Ti and CaO/Al_2O_3. Neither the element contents nor the inter-element ratios show any systematic changes with MgO content.

The extrusion temperature of the aphyric high-Mg mafic MKs has been calculated from equation 5-1 and are given in Table 6-2a. The limited range of temperatures (1425 - 1473^{o}C) reflects the small spread in MgO composition of the lavas. The viscosities and densities (calculated at 1450^{o}C from data given in Bottinga and Weill, 1970, 1972) for these samples are presented in Table 6-2a. The viscosities of the lavas show a two-fold increase from 6.7 to 13.2 poises.

The variations of the aphyric high-Mg MK compositions with MgO content are illustrated in Fig. 6-4 along with the trends of the aphyric ultramafic komatiite (UK) lavas. In contrast to the regular variations of many elements with MgO content displayed by the UK lavas, the aphyric high-Mg MKs show very scattered trends.

One possibility which must be considered for the origin of the aphyric high-Mg MKs is that they represent residual magma from the olivine fractionation processes that occurred in the UK lavas. Modelled liquid trends generated by the fractionation of olivine from the average UK aphyric lava (given in Table 5-6) are shown in Fig. 6-4. The equations given in Pearce (1979; eqn. 11 and eqn. 18) were used to calculate the MgO and FeO^* (K_D = 0.37) composition of the crystallising olivine and the residual liquids. SiO_2 was assumed to enter the olivines in stochiometric amounts while the content of the other major elements(except Ni) in the olivine were assumed to be negligible. Both fractional and equilibrium crystallisation of olivine have been considered and for most of the plots (Fig. 6-4), these two processes define very similar trends. Consequently only the fractional crystallisation trend has been indicated, except on FeO-MgO and SiO_2 - MgO diagrams where both trends are shown. The distribution coefficients ($D^{ol/liq} = C_{ol}/C_{liq}$) are given in Table 6-8. Fractional crystallisation was assumed for the trace elements and F (fraction of liquid remaining) obtained from the major element data. As illustrated in Fig. 6-4 the field of aphyric high-Mg MKs does not fall on the calculated olivine control lines for many of the elements (e.g. SiO_2, Al_2O_3 and FeO) and therefore these rocks do not appear to have been generated from the average UK magma by olivine fractionation.

Clinopyroxene is the main phenocryst phase in the high-Mg MK lavas while orthopyroxene and olivine phenocrysts may also have been present. Fractionation of clinopyroxene (either augite or more commonly observed endiopside - see Table 6-1) from a magma of similar composition to the average aphyric high-Mg MK lava, would generate residual liquids of higher MgO content than the presumed parent magma. However examination of the CaO-MgO plot (Fig. 6-4) for the aphyric lavas shows a rough increase of

CaO content with increasing MgO content. This indicates that fractionation of clinopyroxene alone could not generate the CaO-MgO trend in the aphyric high-Mg MK lavas.

The more complex situation of clinopyroxene and orthopyroxene (and or olivine) crystallising from the average aphyric high-Mg MK magma, could cause the residual liquid to evolve to lower MgO compositions. As the compositions of and proportions of possible fractionating phenocryst phases are unknown several situations have been modelled in Fig. 6-5:

X Control line for residual liquid evolution removing augite (analysis from HSS-10, Table 6-1) from the average lava (see Table 6-7);

Y Control line for residual liquid evolution removing augite and bronzite (see Table 5-1) in the ratio 4:1 from the average lava.

Z Control line for residual liquid evolution removing augite and olivine (Fo_{87} analysis in Tables 6-1) in the ratio 4:1 from the average lava.

The arrow heads of the control lines represent the compositions of the residual liquids after 50 wt.% of the phenocrysts had been removed (i.e. F = 0.5). The augite-only vector (X) and the augite-bronzite (Y) fractionation are discounted, as residual liquid trends developed, are not compatible with the poorly developed trends in the lavas. The combination of augite-olivine (Z) appears to fit best the trends developed in the lavas. The olivine composition used, has a higher FeO* content (Fo_{87}) than that predicted for the average lava composition, calculated by assuming a K_D of 0.37 (Fo_{92}) or 0.3 ($Fo_{93.5}$). Attempts to model the

clinopyroxene-olivine fractionation in the lavas by using selected or assumed mineral compositions and the least squares mixing calculations (mix) of Bryan et al. (1969) did not satisfactorily account for both major and trace element compositions. The postulated augite-olivine fractionation model for the aphyric high-Mg MKs therefore cannot be confirmed with the available data. However, it is re-emphasized that only five samples are present in this category, and such trends that are present are poorly developed.

PORPHYRITIC HIGH-Mg MAFIC KOMATIITES

The compositional similarity between the aphyric (Table 6-2a,b) and porphyritic high-Mg MKs (Table 6-3a, 6-3b) is illustrated by the similar MgO contents of the lavas. However, the porphyritic lavas appear to be more closely related to the ultramafic komatiites than to the aphyric high-Mg MKs. The relationships between these rock types are shown in Figs. 6-6, 6-7 and 6-8. Three of the four porphyritic high-Mg MKs plot close to the olivine control lines from the average ultramafic 'liquid' and have very similar inter-element ratios (i.e. Al/Ti, Ca/Ti and Ti/V) to the aphyric ultramafic rocks.

The porphyritic lavas (HSS-12, HSS-10, HSS-9) have compositions consistent with the model that they were derived from an ultramafic (Group 11) parent liquid by olivine fractionation. One porphyritic sample (HSS-54) does not show any close association with the other lava types.

In the olivine fractionation model proposed for the three porphyritic high-Mg MKs, there is a discrepancy between the inferred fractionating phase (olivine) and the dominant phenocryst phase (clinopyroxene) observed in the rocks. Cox and Bell (1972) and Krishnamurthy and Cox (1977) have noted a similar 'lack of correspondence between the phenocryst present and the mineralogical controls of fractionation deduced

from chemical data' in picritic lavas. In the model developed by Cox and co-workers to account for the anomaly, a series of magmas related by olivine fractionation, are subjected to a process of 'compensated crystal settling.' The magmas became phenocryst enriched without significant changes of bulk composition. The essence of this model is that a series of highly porphyritic lavas with two or more abundant phenocryst phases may be compositionally related by the fractionation of one phase only. A compensated crystal settling model could account for the discrepancy between the observed and predicted phenocryst phases in the porphyritic high-Mg MK samples. Details of such a model are discussed later, in relation to the compositions of all the mafic komatiite lavas.

APHYRIC LOW-Mg MAFIC KOMATIITES

The low-Mg MKs range in composition from 9 - 16.3% MgO (Table 6-4a) and most of the samples are quartz normative. Chemical characteristics of these samples are relatively high SiO_2 contents with general but erratic increase of TiO_2, Al_2O_3, Zr, Y, Ga and V contents with decreasing MgO. FeO^* and Sc show no clear trends with change in MgO content. The variation of the oxide and element contents are shown in Figs. 6-4 and 6-7. The CaO/Al_2O_3 ratio (Fig. 6-11) shows an irregular but marked decrease with decreasing MgO content. The inter-element ratios of the aphyric low-Mg MK lavas show a close association with those from the group II UK lavas and will be discussed in more detail in a following section and in Chapter 7.

The predominant phenocryst phase in the low Mg-MK is endiopside which is present with plagioclase in the lower MgO samples. The crystallisation and settling of olivine in Tony's flow has been proposed (see Chapter 3), although pseudomorphed olivine microphenocrysts are

only observed as a minor phase in the basal chilled margin. However, aphyric low-Mg MK lavas with compositions from 13 - 16.3% MgO (prefixed by the term basic), plot close to the calculated olivine fractionation trends for the oxide/elements in Fig. 6-4 and 6-7. On the FeO^*-MgO diagram, these lavas plot between the two calculated olivine control line trends. If the basic low-Mg MK lavas were derived from the average aphyric UK rock by olivine fractionation, this would indicate that the compositions of the olivine crystallising from the magma were between the limits of either being perfectly zoned crystals (fractional crystallisation) or uniform in composition (equilibrium crystallisation). However, the calculated FeO^*-MgO trends can only be considered as an approximation, as no allowance has been made for iron as Fe^{3+}, either in the postulated parent or the derived liquids.

The olivine control line for the Ni-MgO plot (Fig. 6-7) was calculated by removing olivine by fractional crystallisation in 5 wt.% batches. The $D_{Ni}^{ol/l}$ was calculated from the residual liquid composition (Eqn. 3-2) before the removal of the next olivine batch. The basic low-Mg MKs plot close to the calculated olivine fractionation trend. A rough estimate of the amount of olivine that must be added to the average of the 13 - 15% MgO lavas (see Table 6-10 for this average composition) has been obtained by the least squares mixing calculations. Fo_{87} (Table 6-1) was chosen as no olivine composition has been determined in the MK lavas. The results of the mix are given in Table 6-9a. An upper limit of 15% MgO was chosen for calculating the average, rather than 16%, as it is within the 13 - 15% MgO range that the MK trends start to deviate from the olivine control lines. Approximately 45 wt.% olivine must be removed from the average UK liquid to obtain the average basic low-Mg MK lava.

The trace element content that would be expected has also been calculated and compared to the observed composition in Table 6-9b. The $D^{ol/l}$ that have been used for incompatible elements are given in Table 6-8. The agreement for Zr, Y, V, Ga and Sc between the observed and calculated data is good. The Ni content expected has been calculated as outlined above. Although the agreement is not good it is considered satisfactory, because of the variable content of Ni in the ultramafic lavas. Experimental data for $D_{Co}^{ol/l}$ in liquids with high MgO contents are lacking and a value of 1.9 was chosen.

Overall the good agreement between the calculated and observed major and trace element data is interpreted as support for the olivine fractionation model from the 28.95% MgO average UK liquid to the average 13.88% MgO MK lava. There are however problems with this model in that there are no residual liquids in the compositional range from ∿20 - 16% MgO and clinopyroxene is observed as the major phenocryst phase in the MK lavas. The experimental data of Arndt (1976) demonstrates unexpected crystallisation sequences in the komatiites and it is pertinent to consider his data here, as it is relevant to the olivine fractionation model proposed for the basic low-Mg MK lavas.

From the low pressure crystallisation (1 atmosphere) of an ultramafic komatiite, Arndt found (1976) olivine to be the only liquidus phase down to 1200°C at which point the composition of the residual liquid (glass) was 9% MgO. At this composition, clinopyroxene appeared on the liquidus and crystallised with olivine or plagioclase. In the natural mafic komatiites from Munro Township, clinopyroxene occurs as the major phenocryst phase in the lavas with ∿14% MgO and the lavas plot along a clinopyroxene control line to lower MgO compositions. The discrepancy between the experimental data indicating olivine as the main phenocryst phase in the lavas down to 9% MgO and the natural rocks

showing that clinopyroxene becomes the dominant phenocryst phase at 14% MgO, was further investigated. Arndt (1976) found that natural clinopyroxene spinifex samples gave the same crystallisation sequence as the ultramafic sample. However, a synthetic mafic komatiite liquid, enriched in CaO (CaO/Al_2O_3>1.5) crystallised clinopyroxene as the only liquidus phase from ∿13% MgO. Arndt (1976), in attempting to account for the difference between the experimental and petrographic evidence, has suggested that alteration of the Munro Township lavas had depleted these rocks in CaO, causing clinopyroxene to crystallise later than it should in the experimental charges.

The basic low-Mg MKs from the LUU all have CaO/Al_2O_3 ratios ≥1.5 and, from Arndt's experimental data, would be expected to have clinopyroxene as the major liquidus phase at atmospheric pressure. The predicted 1 atmosphere phase relationships for the basic low-Mg MK lavas from the normative minerals show (Fig. 6-9) that they plot close to the olivine-clinopyroxene phase boundary, but still in the olivine field. Although the precise location of the phase boundary from Cox and Bell (1972) may not be strictly valid for komatiite magmas, it nevertheless indicates that olivine should be a phenocryst phase. The petrographic evidence for the occurrence of olivine in the low-Mg MKs is limited to quench crystals and microphenocrysts (pseudomorphed) in the pillow margins and the basal chilled margin of Tony's flow. Olivine cannot therefore be considered a common phenocryst phase in the low-Mg MK lavas.

If the basic low-Mg MKs were derived from a parental ultramafic liquid by olivine fractionation, the apparent lack of olivine phenocrysts implies one or more of several possibilities:

(a) Efficient removal of olivine phenocrysts by settling in the magma chamber before extrusion of the residual lava.

(b) Resorption of any entrained olivine phenocrysts as the lava was extruded.

(c) Subsequent metamorphic alteration has destroyed the textural evidence for olivine phenocrysts in the lavas.

Calculated viscosities of liquids of similar composition to the basic low-Mg MK lavas are low (see Fig. 6-10). Settling rates of olivine phenocrysts would be relatively high (Fig. 5-14) in these liquids. Possibility (a) may therefore may be a plausible explanation for the lack of olivine phenocrysts in the basic low-Mg MKs.

Possibility (b) would require that the magma be extruded rapidly from depth so that the magma arrived at the surface above its liquidus temperature and was thus able to resorb olivine phenocrysts. Possibility (c) is also likely, particularly if the olivine phenocryst content was of the order of a few percent. Any of these possibilities either alone or combined, could thereforeaccount for the deficiency of olivine phenocrysts in the lavas.

The trend of the aphyric low-Mg MK lavas deviates from the olivine control line (Fig. 6-4) in samples that contain <13% MgO. The phenocryst phase observed in these samples is mainly clinopyroxene, with minor plagioclase. Trends developed in the 9 - 13% MgO aphyric low-Mg MKs (evolved) are in general compatible with clinopyroxene fractionation (Fig. 6-4). Plagioclase and olivine may also have fractionated in minor amounts.

In Table 6-10 the composition of a residual liquid has been calculated after 40 wt.% endiopside has been removed from the average basic low-Mg MK composition. The endiopside composition used is also given in Table 6-10, along with the average composition of the evolved low-Mg MK lava. The trace element content of the modelled composition

has been calculated using the $D^{cpx/liq}$ values given in Table 6-8. This composition is similar to the average evolved low-Mg MK except for lower CaO and higher FeO^* and Zr contents in the former. Better agreement between the modelled composition and the evolved low-Mg MK could probably be obtained if plagioclase and olivine were also considered as fractionating phases. However, the latter two minerals have not as yet been analysed in the MK lavas (as no fresh mineral relics are presented in the samples studied) and their precise composition is unknown. Nevertheless it appears that the evolved low-Mg MKs can be derived from the basic low-Mg MK lavas when clinopyroxene is considered to be the predominant fractionating phase.

PORPHYRITIC LOW-Mg MAFIC KOMATIITES

The major and trace element data for the porphyritic low-Mg MK lavas are given in Tables 6-5a and 6-5b. The variation of oxides and elements with MgO content are illustrated in Fig. 6-6. The compositional fields occupied by the aphyric low-Mg MK lavas are also outlined in these diagrams. The porphyritic samples have a slightly wider range of compositions (9.3 - 18.7% MgO) than the aphyric lavas (9.2 - 16.3% MgO), but there is considerable overlap in composition of the two groups. Some porphyritic samples are from phenocryst enriched portions of MK flows, while others have a high percentage of clinopyroxene phenocrysts, suggesting that they are also phenocryst enriched.

In Fig. 6-11 the CaO/Al_2O_3 ratios of the MK samples have been plotted against MgO contents. The aphyric lavas define a trend of decreasing CaO/Al_2O_3 with decreasing MgO, which probably represents the line of liquid descent of the residual magma. The porphyritic lavas plot on both the high and low CaO/Al_2O_3 sides of the postulated liquid

trend. Samples such as R-14 and VB-2 have high CaO/Al_2O_3 ratios, CaO contents and abundant model clinopyroxene (Appendix 1). The composition and mineralogy of these samples is consistent with their derivation from the magma by accumulation of clinopyroxene.

In order to obtain an estimate of the amount of phenocryst accumulation that has occurred, curves for liquid plus clinopyroxene have been drawn (Fig. 6-11) for various amounts of accumulated clinopyroxene. The clinopyroxene composition used to calculate the trend lines was the endiopside composition from Table 6-10. Obviously the accumulation of clinopyroxene into the lavas generates higher CaO/Al_2O_3 ratios and a sample such as R-14 (CaO/Al_2O_3 = 3.2) has accumulated in excess of 60% clinopyroxene. The amount of clinopyroxene accumulation indicated can only be used as a guide, as not all the fractionating phases (either observed or suspected) have been accounted for in this diagram.

There are a number of porphyritic samples with lower CaO/Al_2O_3 ratios than the postulated liquid trend. These samples could not have obtained their compositions by clinopyroxene accumulation from the range of liquid compositions. Several possibilities have been indicated on Fig. 6-11 for obtaining these compositions. The first is that they represent residual liquids from the process of clinopyroxene fractionation from the main liquid trend. The field of compositions that these liquids could develop is indicated by the dashed lines on the low CaO/Al_2O_3 side of the liquid trend line. Samples plotting in this area could represent such residual liquids. However, several samples with relatively low CaO/Al_2O_3 ratios and high MgO contents could not have obtained their composition by this mechanism.

Some of these samples such as SC-6C from Tony's flow are of

indisputable cumulate origin (see Chapter 3). Accumulation of olivine or orthopyroxene is indicated by the high MgO compositions. From consideration of the SiO_2-MgO trend (Fig. 6-6), the possibility of olivine fractionation is favoured over orthopyroxene. The trend lines for olivine accumulation into the postulated liquids are shown in Fig. 6-11. The Fo_{87} olivine composition (Table 6-1) was used to calculate the trend lines. Fig. 6-11 shows that a maximum of 13 wt.% olivine needs to be accumulated into the liquids to give the high MgO compositions of the low CaO/Al_2O_3 rocks.

As indicated previously, samples that are believed to represent magmas enriched in cumulus clinopyroxene, have mineralogies and chemical compositions that are consistent with the clinopyroxene accumulation model. However, samples postulated to be enriched in cumulus olivine, pose a dilemma, similar to that noted earlier for the aphyric low-Mg MK lavas, in that olivine is not observed in most samples or if it occurs at all, is very minor in content. The amount of olivine accumulation proposed, however, is relatively small (<13 wt.% in all cases) and one possibility proposed previously to account for the lack of olivine in the liquid samples, namely that evidence of olivine has been destroyed by metamorphic overprinting, is also tenable for the cumulate samples.

The porphyritic low-Mg MK rocks appear to have been formed by at least three different processes:

(a) Crystallisation of liquid, giving rise to porphyritic or coarse grained samples which have similar compositions to the aphyric lavas.

(b) Clinopyroxene accumulation giving rise to samples with high CaO/Al_2O_3 ratios and CaO contents.

(c) Olivine accumulation developing samples with 'normal' CaO/Al_2O_3 ratios and high MgO contents.

The average aphyric low-Mg MK (lava composition) has been calculated from all the data in Tables 6-4a and 6-4b for selected elements and given in Table 6-7. The average porphyritic low-Mg MK rock has also been calculated from all the data in Tables 6-5a and 6-5b and presented in Table 6-7. The average low-Mg MK rock (Table 6-7) has been calculated by equally weighting the above compositions.

TEXTURALLY ALTERED MAFIC KOMATIITE ROCKS

The texturally altered MK rocks have been obtained from widely scattered areas of the LUU for the purposes of investigating any spacial or temporal variation in the MK compositions. The major and trace element data for these rocks are given in Tables 6-6a and 6-6b respectively. The major and trace element data have been plotted against MgO content in Fig. 6-6 along with the porphyritic low-Mg MK samples and the compositional fields occupied by the aphyric MK lavas. Examination of these plots shows that there are no systematic variations of composition of the texturally altered MK's, compared to the better preserved MK samples. As a number of samples are from the Sandspruit Formation (see Appendix 1), the similar chemical composition indicates that there was no discernable temporal change in the MK lava composition. The average composition of the texturally altered MK rocks has been calculated from the data in Table 6-6a and 6-6b and given in Table 6-7. The average texturally altered rock is very similar in composition to the average MK rock and no temporal variations in the low-Mg MK lavas can be identified within the scatter of the data.

6-5 INTER-ELEMENT RELATIONSHIPS BETWEEN KOMATIITE LAVA TYPES

In Chapter 7 inter-element ratios of komatiite and tholeiite lavas are discussed with respect to variations found in the Barberton lavas and lavas from other greenstone belts. In this section inter-element ratios of komatiite lava types will be discussed briefly in connection with the fractionation models proposed for the mafic komatiites. Only samples that are considered to represent phenocryst-free magma compositions are considered.

In Fig. 6-8 Al/Ti, Mg/Ti, Ca/Ti, and Ti/V ratios have been plotted against MgO content for all the komatiite samples. The Al/Ti and Ti/V ratios of many of the lavas have restricted values of around 10 and 20 respectively. In Chapter 7 these lavas are referred to as the group II rocks. The porphyritic high-Mg MKs and basic low-Mg MKs are believed to have been derived from the average UK liquid by olivine fractionation and variations of the inter-element ratios (Fig. 6-8) are compatible with this model. The evolved low-Mg MKs are considered to have been derived by further amounts of olivine plus clinopyroxene fractionation and the lower Ca/Ti ratios in these samples are consistent with the expected trend.

In contrast to the consistent relationships displayed by the group II lavas, the aphyric high-Mg MKs have comparatively large ranges in these inter-element ratios and in some cases are significantly different to the group II lavas. On average the high-Mg MKs have similar inter-element ratios to the two low-Ti ultramafic lavas from the LUU and to ultramafic komatiites from other greenstone belts (see Chapter 7). The high-Mg MKs and the low-Ti UK samples are referred to as group I rocks. In Chapter 7 it is proposed that the two different groups of lavas are derived from separate source compositions.

6-6 CRYSTAL FRACTIONATION MODELS FOR THE MAFIC KOMATIITES

A summary of possible processes proposed for the genesis of the mafic komatiites is given below:

(a) Porphyritic high-Mg and basic low-Mg MKs could be derived from an ultramafic parent liquid by olivine fractionation.

(b) Further fractionation of olivine plus clinopyroxene could generate the low-Mg MK lavas.

(c) The aphyric high-Mg MKs do not appear to be simply related to UK or other MK magmas by olivine or clinopyroxene fractionation . In an earlier section it was tentatively suggested that fractionation of augite and olivine was responsible for the generation of the range of aphyric high-Mg MK compositions. However, because the trends displayed by the few available samples are very scattered, further possible petrogenetic processes will not be considered for this lava type.

In addition to the above models certain porphyritic low-Mg MK samples are believed to have obtained their compositions by olivine or clinopyroxene accumulation while other porphyritic samples have essentially identical compositions to their aphyric counterparts.

The lavas in (a) (b) could all ultimately have originated by crystal fractionation processes from an ultramafic parent magma. A single model of genesis connecting these lavas to an ultramafic parent magma, is therefore a reasonable consideration and must satisfactorily account for the following features:

(1) Lava compositions (14 - 22% MgO) consistent with olivine fractionation from a parental ultramafic liquid.

(2) Mainly clinopyroxene fractionation producing the lavas with composition from 14 - 9% MgO.

In addition the model must take into account the expected low viscosities of the high magnesium liquids (Fig. 6-10) and the relatively high settling rates of olivine (and clinopyroxene crystals as shown in Fig. 5-14).

The model proposed, includes compensated, partially compensated and uncompensated crystal settling in a magma chamber at moderate pressures (<10 kb) and has been adapted from compensated crystal settling models outlined by Cox and Bell (1972) and Krishnamurthy and Cox (1977) for picritic and basaltic lavas. The model is graphically illustrated in Fig. 6-12. The assumptions on which the model is based are listed below:

(1) Olivine (50% MgO) is the liquidus phase until the residual magma has attained a composition of 14% MgO.

(2) In residual magmas with 14% MgO or less, both olivine (50% MgO) and clinopyroxene (20% MgO) crystallise in the ratio 1:4 (ratio from Krishnamurthy and Cox, 1977).

(3) The parent liquid has a similar composition to the average aphyric ultramafic komatiite (∿29% MgO).

(4) Assumptions 1 - 4 from Krishnamurthy and Cox (1977, p. 70) are also incorporated.

(5) Crystallisation takes place in a stepwise fashion, the amount of crystallisation varying at each step as shown in Fig. 6-12.

The model is characterised in Fig. 6-12 where in step 1, a magma

chamber is filled with phenocryst free ultramafic magma (29% MgO). Four cells of uniform volume are considered, with the uppermost cell (w) at the top of the chamber. In step 2, a portion of the magma crystallises as olivine and the crystals settle (step 3) through one cell depth. These crystallisation and settling steps are repeated as indicated in Fig. 6-12. The MgO content, description of the magma and crystal-liquid proportions of the magma in each cell after each step, are also given.

Magmas C,D, H and M have the same composition, but vary significantly in the proportion of phenocrysts and liquid contents. These magmas are broadly similar in composition to the porphyritic high-Mg MKs. Magmas H and M contain abundant clinopyroxene as well as olivine phenocrysts and are related to the ultramafic parent liquid by olivine fractionation. Magmas extracted from cell X in step 6 or cell Y in steps 7 and 8 could give rise to lavas that are similar to the porphyritic high-Mg MKs. Magmas F, E and L are similar in composition to the basic low-Mg MKs and are related to the ultramafic parent liquid by olivine-only fractionation. Magma F is phenocryst free and could give rise to aphyric lavas while magmas E and L have significantly more clinopyroxene than olivine and could give rise to porphyritic lavas. Magmas J and K are similar in composition to the evolved low-Mg MKs and are related to the ultramafic parent liquid by olivine and clinopyroxene fractionation. Extraction and extrusion of these magmas could give rise to aphyric (J) and porphyritic (K) lavas, similar in composition to the evolved low-Mg MKs.

In step 8 all the MgO in the residual liquid has been crystallised as clinopyroxene and olivine (4:1). This step, while not being petrologically reasonable, provides an estimate of the maximum clinopyroxene

and olivine phenocryst contents of the prophyritic magmas. The model outlined in Fig. 6-12 has been grossly simplified, but nevertheless does provide, in principal, a crystallisation mechanism that accounts for the features of the mafic komatiite lavas (as listed earlier, 1-3). A wide range of magma compositions can be produced with differing ratios and amounts of phenocryst phases and still satisfy the requirement of olivine control (23 to 14% MgO lavas) and predominantly clinopyroxene control in the lower MgO lavas (14 - 9%).

In detail however, the phenocryst assemblage inferred for the porphyritic high-Mg MKs, is not in accord with what is thought to be present in these lavas. Magma H has roughly equal amounts of clinopyroxene and olivine (∿20 wt.% each) increasing to a maximum of ∿36% clinopyroxene and 26% olivine in magma M. No attempt has been made to quantify modal portions of clinopyroxene and olivine (as these phenocrysts are extensively altered) in the porphyritic high-Mg MKs, but visual inspection indicates clinopyroxene > olivine. Krishnamurthy and Cox (1977) have noted from an assessment of data in O'Hara (1968) that the clinopyroxene stability field expands relative to that of olivine (in dry systems) as pressure increases. The clinopyroxene/olivine ratio in the hypothetical magmas H and M would therefore be expected to increase, if the magma chamber was at higher pressures, and result in a phenocryst assemblage which is in better accord with what is thought to be present in the porphyritic high-Mg MKs.

6-7 COMPARISON OF BARBERTON KOMATIITES WITH KOMATIITES FROM OTHER GREENSTONE BELTS

The occurrence of ultramafic komatiites from greenstone belts has been discussed in Chapters 1 and 5. MKs are invariably present

in close association with these ultramafic lavas. Tholeiitic rocks and more differentiated lavas are often present as well. As many of the researchers on greenstone belts have either defined their own komatiite classification scheme and nomenclature or modified existing schemes, broadly equivalent rock types will simply be called komatiites in this section. As for the UKs, comparison will be restricted to lavas from Belingwe, Western Australia and Munro Township.

Data chosen for comparison are from sources listed in Table 5-9 and also from Hawkesworth and O'Nions (1977) and Williams and Hallberg (1973). The petrographic descriptions of MK lavas in these references usually do not enable such definitive selection of aphyric lava compositions as was possible for the ultramafic lavas. Therefore only samples with obvious cumulate origins were excluded from the diagrams illustrated in this section (such as samples from Fred's flow, Arndt, 1977a; Arndt et al., 1977).

MgO-CaO-Al_2O_3

MgO, CaO and Al_2O_3 data have been plotted on ternary diagrams in Fig. 6-13, -14, -15, -16, for Barberton, Belingwe, Western Australia and Munro Township areas respectively. UKs and tholeiitic basalts from these areas have also been plotted on the diagrams for comparison. For the Barberton data a line joining the average UK aphyric lava and average basic low-Mg MK lava passes through the MgO apex, consistent with the olivine fractionation mechanism proposed for the basic low-Mg MK lavas. However, lavas with lower MgO contents show an abrupt change from olivine control towards Al_2O_3 enrichment. The lowest extent of the komatiite field is indicated by the lines drawn parallel to the CaO-Al_2O_3 and MgO-CaO axes from the point $MgO_{(29)}$-$CaO_{(31)}$-$Al_2O_{3(40)}$.

These are also drawn on the other diagrams (Figs.6-14 to 6-16). The tholeiites plot on the high Al_2O_3 side of the diagram (Fig. 6-13) and do not overlap with the komatiites. The komatiites from Belingwe (Fig. 6-14) and Munro Township (Fig. 6-16) show decreasing MgO content consistent with olivine fractionation as the CaO/Al_2O_3 ratios remain close to unity. As with the Barberton rocks, however, the lower MgO MKs deviate from the olivine control line to relatively higher Al_2O_3 contents.

Nisbet et al. (1977) have proposed (as one of several possibilities) that MKs could represent residual liquids after precipitation of olivine from UK liquids. They also note that clinopyroxene becomes an important fractionating phase in liquids with 15-16% MgO. A similar model of genesis has been proposed for MKs from the Munro Township by Arndt et al. (1977) to that proposed here for the Barberton MKs. In Arndt et al.'s model olivine is considered the main fractionating phase in liquids down to 12% MgO whereafter both olivine and clinopyroxene crystallise.

The boundary of the komatiite field derived from the Barberton data, also indicates the limits of the komatiite fractionation trends from the Belingwe, Western Australia and Munro Township komatiites. Most samples that the various authors have termed tholeiites plot on the high Al_2O_3 side of the boundary line. The usefulness of this division between komatiites and tholeiites is however limited as some greenstone belts have a complete gradation in compositions from mafic komatiites to tholeiitic basalts (Naldrett and Turner, 1977).

The data from Western Australia (Fig. 6-15) show a more scattered trend than that from the other greenstone belts. Naldrett and Turner (1977), considering only lavas from the Upper Greenstones in the Yakabindie area, consider that olivine was the only crystallising phase

from the ultramafic komatiites until the residual liquids had attained a composition of 13% MgO. At this point the magmas crystallised olivine and clinopyroxene. Nesbitt et al. (1979) however suggest from REE data that mafic komatiites are not the products of fractional crystallisation of ultramafic komatiites. They consider a process of sequential melting of mantle peridotite a more plausible process for obtaining the spectrum of komatiite lava compositions in Archaean greenstone belts. No REE data have yet been obtained on a key suite of samples from Barberton, but data have been reported in the literature and will be considered in Chapter 7.

Finally the compositional 'gap' in MgO between 16 - 20% warrants brief discussion before leaving this section. This phenomenon has already been stated in an earlier section as one of the objections to the olivine fractionation mechanism for obtaining the basic low-Mg MKs from the average UK lava. It appears that similar compositional 'gaps' have been noted for lava compositions from other greenstone belts. Nisbet et al. (1977) have noted a break in MgO content in the lavas from the Belingwe greenstone belt between 16 - 20% MgO. Arndt et al. (1977) have noted a similar abundance gap in MgO contents (15 - 20% MgO) from the Munro Township lavas. The gap in komatiite compositions from the LUU occurs between 16.3 - 20.4% MgO. While this abundance gap in composition from the three areas outlined above may merely reflect inadequate sampling, the fact that it occurs in the same MgO range may be genetically significant. On the other hand Naldrett and Turner (1977) in their study of the Yakabindie area in Western Australia note that there is a continuous range in MgO contents down to 6.9% MgO. However their observation was based on data from all the samples they analysed and no cognizance was taken of whether lavas

represent possible partial cumulates or phenocryst-free magma compositions.

If this compositional gap is real it poses a serious problem to fractionation models for deriving thelow-Mg MKs from ultramafic magmas. It is reasonable to expect that ultramafic magmas, because of their very high liquidus temperatures, would have to progress rapidly from their site of generation in the mantle to the surface in order to be extruded in a liquid state. It is of interest to speculate that if ultramafic magmas filled relatively shallow magma chambers (as has been suggested in the previous section) and on cooling crystallised olivine, these crystals would have relatively high settling rates (see Fig. 5-14). As can be seen in Fig. 6-10 the viscosity of the residual liquids would increase slowly with decreasing MgO content, until 15 - 16% MgO. Hereafter small changes in composition of the residual liquid caused by further olivine crystallisation would lead to very rapid increase in viscosity of the liquid and consequent lowering of the crystal settling velocities. This suggests that if the magma chamber was tapped from time to time, the most likely lava composition would have $\lesssim$16% MgO and possibly be porphyritic. Lavas with compositions between that of the ultramafic parent liquid and 16% MgO, would be expected to be relatively scarce, but not necessarily totally absent from the volcanic pile. Attempting to model the precise mechanics of the process, however, is complicated by such factors as the effects of pressure or viscosity, cooling and crystallisation rates and turbulence in the magma chamber. However, in broad outline the variations of viscosity (on phenocryst settling rates) with composition, probably have an important bearing on the compositional "gap" between $\sim$16 - 20% MgO in komatiite lavas.

6-8 CONCLUSIONS

In Chapter 4 lavas with MgO contents in the range 9 - 24% have been subdivided into the high-Mg mafic komatiites (20 - 24% MgO) and the low-Mg mafic komatiites (9 - 16% MgO). Interpretation of the geochemical data obtained for the mafic komatiites has shown the following:

(1) The aphyric high-Mg MKs are not related to the ultramafic komatiite lavas by olivine fractionation. Neither are they simply related to the porphyritic high-Mg MKs.

(2) Olivine and clinopyroxene (augite) have tentatively been suggested as important fractionating phases in the aphyric high-Mg MKs.

(3) The basic low-Mg MKs and porphyritic high-Mg MKs have compositions consistent with their derivation from an ultramafic komatiite parent magma by olivine fractionation. However, the dominant phenocryst phase in these lavas is clinopyroxene.

(4) The evolved low-Mg MKs could have been derived from the basic low-Mg MKs if clinopyroxene was the dominant fractionating phase. The fractionation of minor amounts of olivine and plagioclase could also have occurred.

(5) The compositions of the porphyritic low-Mg MK samples suggests that at least three processes could have been in operation:

(a) accumulation of clinopyroxene giving rise to samples with high CaO/Al_2O_3 ratio and CaO contents;

(b) accumulation of olivine developing samples with relatively high MgO contents and "normal" CaO/Al_2O_3 ratios;

(c) closed system crystallisation of magmas developing coarse-grained rocks of similar composition to the aphyric lavas.

(6) A mechanism of partially compensated and uncompensated crystal settling in a magma chamber has been proposed to account for the discrepancy between observed phenocryst phases and those inferred to have fractionated from the lavas to account for the geochemical trends developed in the mafic komatiites.

(7) Comparison of trends developed in mafic komatiites from different greenstone belts show that the lavas initially undergo MgO depletion forming trends which are consistent with olivine fractionation. The trends deviate towards Al_2O_3 enrichment in lavas with 12 - 15% MgO consistent with the fractionation of clinopyroxene as the major phenocryst phase.

CHAPTER 7.

THOLEIITIC BASALTS AND RELATED ROCKS

7-1 INTRODUCTION

Viljoen and Viljoen (1969c) identified tholeiitic basalts in the suite of volcanic rocks developed in the Komati and Theespruit Formations. The tholeiites were distinguished from the komatiites on the basis of chemical composition (i.e. MgO-CaO-Al_2O_3 and MgO-FeO relationships). While tholeiitic lavas were found to be a minor component of the LUU volcanics they form a major part of the volcanics in the UMFU (Upper Mafic-Felsic Unit). Here the volcanic rocks crop out in well developed cycles consisting of basalt at the base, grading upwards into dacitic lavas and terminated by a chart cap. The general nature of the UMFU volcanics is comparable to many other Archaean greenstone belts (Glikson, 1971; Goodwin, 1972, 1978; Condie and Harrison, 1976) and distinctly different from the komatiitic character of the lavas in the LUU.

Additional analyses of tholeiitic basalts have been obtained from the LUU as well as trace element analyses of selected samples from the Viljoens' collection from both the LUU and UMFU. The purpose of obtaining these data were for (a) comparison with mafic komatiite compositions, (b) comparison with tholeiitic basalts from other greenstone belts and (c), comparison with basaltic lavas developed in various modern tectonic environments. No attempt has been made to provide detailed sample coverage of sections through the UMFU. Thorough investigation of the geochemistry of lavas in the UMFU is planned in future research projects on the Barberton greenstone belt.

7-2 SAMPLE LOCATION, FIELD CHARACTERISTICS AND PETROGRAPHY

Samples of tholeiitic basalts have been obtained from all three Formations of the LUU. The location of samples from the type area of the Komati Formation are shown in Fig. 7-1. Selected samples from the Viljoens' collection from the UMFU have been analysed for trace elements in this work. Rocks sampled by the author are prefixed with HSS-,SC- or DC-.

In outcrop, the basalts occur as flows and pillowed horizons. Outcrops also occur that consist of massive fine to medium-grained rocks for which clear field relationships could not be established. These outcrops could be portions of flows, sills or dykes. Further details of the samples from the Viljoens' collection can be found in Viljoen and Viljoen (1969e). Brief petrographic descriptions and sample locations are given in Appendix 1. On the basis of texture and composition (discussed in the next section) the tholeiitic basalts from the Onverwacht Group have been sub-divided into four types:-

1) low-Ti basalts
2) high-Ti basalts
3) high-Mg basalts
4) texturally altered rocks.

In thin section the low-Ti and high-Ti basalts are similar, consisting of fine grained aphyric samples with laths of amphibole, that probably represent altered clinopyroxene needles. Altered plagioclase needles are also common. The groundmasses of these samples consist mainly of amphibole with minor amounts of plagioclase, chlorite, quartz epidote and spinel. Several samples have microphenocrysts of clinopyroxene. Skeletal ilmenite crystals are common in the high-Ti basalts

but are absent from the low-Ti basalts.

The three samples that have been termed high-Mg basalts are from the Komati Formation and are medium-fine grained rocks consisting of clinopyroxene and plagioclase. Both minerals are usually altered. The small amount of interstitial material present consists of amphibole, chlorite, calcite and epidote.

The fourth group of samples consists of the texturally altered rocks. Samples HSS-39 and HSS-111 have metamorphically reconstituted mineral assemblages of plagioclase and actinolite. Other samples in this group have very high volatile contents (V-13, DC-1, HSS-52B) relative to the other basalt types discussed above. The pillow sample HSS-52B, from the Komati Formation, is dacitic in composition and in thin section is an aphyric rock with a very fine grained groundmass.

7-3 GEOCHEMISTRY OF THE THOLEIITIC BASALTS AND RELATED ROCKS

Major element and normative mineral compositions are given in Tables 7-1a to 7-4a for the low-Ti basalts, high-Ti basalts, high-Mg basalts and the altered samples respectively. Trace element data and inter-element ratios are given in Tables 7-1b to 7-4b for the above group of rocks. The major and trace element data have been recalculated volatile free and the total Fe content has been partitioned such that $Fe_2O_3/FeO = 0.2$.

Most of the samples are hypersthene normative and either quartz or olivine normative. Two samples (34-J and LV-4) of low-Ti basalts (Table 7-1a) are nepheline normative. However as these two samples have similar Ti, Zr, Y, Ni, Co, Cr, V, Ga and Sc contents to the rest of the samples in this group, the appearance of nepheline in the norms of 34-J and LV-4 is probably due to the effects of alteration on the

alkali element contents of these lavas. One sample in the altered basalt group is also nepheline normative (HSS-111, Table 7-4a) but as this sample has been extensively altered (LOI = 19.08) little reliance can be placed on its present normative mineral composition.

RECLASSIFICATION OF SAMPLES

Some of the samples from the Viljoens' collection have been reclassified in this work. Sample TS-1 that Viljoen and Viljoen (1969c) considered to be a metatholeiite from the Theespruit Formation is considered here to be a mafic komatiite (see Tables 6-4a and 6-4b). Other samples LV-6 (Mg-rich metabasalt from the Hooggenoeg Formation), M-57 and 526 (metabasalts from the Kromberg Formation) with their designation from Viljoen and Viljoen (1969e) in parentheses, are also considered to be mafic komatiites. Samples were reclassified on the basis of their similar major and trace element compositions to the mafic komatiites from the LUU. The recognition of mafic komatiites in the UMFU shows that komatiitic magmas were produced throughout the development of the Onverwacht Group although at variable rates.

Viljoen and Viljoen (1969e) grouped basalts from the UMFU under several headings:- metatholeiites, metabasalts and Mg-rich metabasalts. Samples of low-Ti basalts in Table 7-1a include rocks from the Hooggenoeg Formation which Viljoen and Viljoen (1969e) termed metatholeiites or Mg-rich metabasalts. The HSS- samples in this group are from the Theespruit and Komati Formations. Samples of low-Ti basalts have not yet been analysed from the Kromberg Formation while the sample HSS-111 (Table 7-4a and 7-4b) from the Sandspruit Formation, although very altered, is similar in composition to the low-Ti basalts.

The high-Ti basalts (Tables 7-2a and 7-2b) contain two samples

LV-9 and 21-J from the Kromberg Formation that the Viljoens' called metabasaltic rocks. The samples SC-9 and HSS-224 are from the Komati and Hooggenoeg Formations respectively. The low-Ti and high-Ti basalt types both occur in the LUU and the UMFU.

The three high-Mg basalts (Tables 7-3a and 7-3b) are from the Komati Formation. No lavas of similar composition have as yet been found in the other Formations of the Onverwacht Group. The sample HSS-52B (Tables 7-4a and 7-4b) is of intermediate composition and from the Komati Formation. It is similar in composition to some of the intermediate lavas from the Hooggenoeg and Kromberg Formations (see Viljoen and Viljoen, 1969e). As lavas of intermediate composition have hitherto not been reported from the Komati Formation the analysis of HSS-52B is reported for record purposes only and is not further discussed.

RELATIONSHIP BETWEEN THOLEIITIC BASALTS AND MAFIC KOMATIITES

The variations of selected elements with MgO content in the tholeiitic basalts are illustrated in Fig. 7-2. The fields occupied by the aphyric high-Mg and low-Mg MKs have been drawn in for comparison. The high-Ti basalts are clearly distinguished from the low-Ti basalts on some of the plots as the former have significantly higher TiO_2, FeO^*, Zr, Y and V contents. The high-Mg basalts are distinguished from the other basalt types by their higher MgO contents. They are also separated from the low-Mg MKs by their lower TiO_2 and CaO contents and higher Al_2O_3 contents. Some of the altered basalts (not plotted) have compositions transitional between the high-Ti and low-Ti basalts. The high-Ti basalts are also readily distinguished from the low-Mg MKs by their higher TiO_2, Al_2O_3, FeO^*, Zr, Y and V contents and lower

Ni and Cr contents. The low-Ti basalts have compositions which plot close too the the low-Mg MK fields for many elements. However the higher Al_2O_3 and lower Cr contents of the low-Ti basalts show that they are not simply related to the low-Mg MKs.

As has been outlined in Chapter 4 the Al_2O_3 content is a useful parameter for distinguishing tholeiitic basalts from mafic komatiites in the Onverwacht Group volcanic rocks. It appears that this parameter may be useful for distinguishing between the tholeiitic and komatiitic suites of lavas from other greenstone belts as well. In Fig. 7-3 histograms for lavas with Al_2O_3 contents in the range 9 - 18% have been plotted. Sources of the data are referenced in Chapter 6. The Al_2O_3 contents in lavas from the LUU, Belingwe greenstone belt and Western Australia show a clear bimodal distribution. Samples on the low Al_2O_3 side of the divide have been referred to either as komatiites or as lavas similar to komatiites in composition by the various authors. Samples on the high Al_2O_3 side have generally been referred to as tholeiitic basalts. The data from Munro Township show a similar bimodal distribution although the divide between komatiites and tholeiites is not as distinct as for the lavas from the other areas. However, the majority of samples on the low Al_2O_3 side have been referred to as komatiites by Arndt (1975) and Arndt et al. (1977) and most of the lavas on the high Al_2O_3 side have been termed tholeiites. While the possibility of transitional rock types has not been discounted the Al_2O_3 content appears to be a useful parameter for distinguishing komatiitic and tholeiitic lavas.

The compositional break in Al_2O_3 content between tholeiitic basalts and low-Mg MKs in the LUU suggests that processes of crystal fractionation did not form a continuous series of magma compositions

between the two types of lavas. However apparent compositional breaks do not preclude one magma type being derived from another and possible crystal fractionation processes relating the tholeiitic basalts to low-Mg MKs are discussed below.

Clinopyroxene is the dominant phenocryst phase in the low-Mg MKs (see Chapter 6) with minor plagioclase. Olivine is an inferred phenocryst phase but seldom observed. It is considered likely that clinopyroxene fractionation accompanied by minor plagioclase and olivine would control the composition of the residual liquids derived from the low-Mg MK magmas. The most evolved aphyric low-Mg MK lava (average of the 5 pillow samples given in Tables 6-4a and 6-4b) is considered a likely parent magma composition. Fractionation of clinopyroxene, plagioclase and olivine in any combination and proportions from this parent composition would generate residual liquids with higher TiO_2 and Zr contents. The low-Ti basalts and high-Mg basalts are therefore eliminated as possible lavas derived from low-Mg MKs by fractionation of these phenocryst phases. The possibility that the high-Ti basalts were derived from a low-Mg MK magma (such as the average pillow composition) by fractionation of clinopyroxene, plagioclase and olivine has been tested using the least squares approximation of Bryan et al. (1969). Numerous compositions of the phenocryst phases were used (both analysed and assumed compositions) but no plausible combinations of fractionating phenocrysts and residual liquids were obtained that satisfactorily accounted for the major and trace element composition of the average high-Ti basalt. It therefore appears unlikely that any of the basalt types were derived by fractionation of clinopyroxene, plagioclase and olivine from the low-Mg MK lavas.

As discussed in an earlier section the main phenocryst phases in the basalts are clinopyroxene and plagioclase. Clinopyroxene has been analysed from a low-Ti basalt (HSS-56) and the average composition of eight determinations is given below:- SiO_2-52.84, TiO_2-0.18, Al_2O_3-2.26, Total Fe as FeO-6.15, MnO-0.15, MgO-18.33, CaO-19.44, Na_2O-0.21, and Cr_2O_3-0.56. The clinopyroxenes from this sample have a restricted range of compositions (Mg/(Mg+Fe) = 84.0±0.8) and are similar in composition to clinopyroxenes from the mafic and ultramafic komatiites.

The predicted one atmosphere phase relationships (from Cox and Bell, 1972) from the normative mineralogy of the basalts are illustrated in Fig. 7-4. This diagram indicates that the low-Ti basalts (the two nepheline normative samples have been omitted) should contain olivine and plagioclase as phenocryst phases while some samples may in addition have clinopyroxene. The phenocryst assemblage indicated for the high-Ti basalts is olivine or olivine, plagioclase and clinopyroxene. Olivine is the only phenocryst phase indicated for the high-Mg basalts. An anomalous situation exists, similar to that noted for the mafic komatiites, in that olivine is predicted as being an important phenocryst phase in all basalt types but has not been observed in thin section. This anomaly may be partly due to the altered nature of some samples with respect to the alkali elements and the importance they assume in the normative mineral calculations (i.e. samples that have lost Na_2O will have less normative diopside than they should and may now plot in the olivine stability field of Fig. 7-4). However, the elements that are believed to be relatively insensitive to alteration processes (groups A and B in Chapter 2) should reflect the effects of the phenocryst phases fractionating and accumulating in the lavas.

Attempts have been made to test whether any of the basalt types are related by fractionation or accumulation of olivine, clinopyroxene and plagioclase using the least squares approximation calculations (Bryan et al., 1969). As mineral composition data are generally lacking for the basalts a range of compositions for each phenocryst phase was chosen. However, no satisfactory major and trace element compositions were calculated either by removal or addition of phenocryst phases (in various combinations) that enabled the derivation of one type of basalt from another (average compositions of the basalt types given in Table 7-5).

Fig. 7-5 illustrates graphically the difficulties encountered when attempting to derive one basalt type from another by fractionation or accumulation of likely phenocryst phases. For example, the SiO_2-MgO plot shows that the three basalt types could be related by clinopyroxene fractionation. However, the TiO_2-MgO, Al_2O_3-MgO and CaO-MgO relationships eliminate this possibility. Similar difficulties arise with other possibilities and it therefore appears unlikely that any of the basalt types are related by low pressure fractionation processes.

7-4 INTER-ELEMENT ASSOCIATIONS AND MANTLE HETEROGENEITY

Although it has been demonstrated in the previous section that basalts and mafic komatiites are not simply related by crystal fractionation processes, consideration of inter-element ratios suggest a common source for some of the komatiite and tholeiitic basalt types. Average values of inter-element ratios for previously identified types of komatiitic and tholeiitic lavas from the Onverwacht Group are given in Table 7-6. The data have been divided into three groups.

The group I rocks from the Onverwacht consist of the low-Ti ultramafic lavas, aphyric high-Mg mafic komatiites and the low-Ti basalts. While there is some variability of the inter-element ratios both within individual lava types and between the lava types, average inter-element ratios of the group I rocks are generally similar to the chondritic ratios (Table 7-6) listed by Nesbitt and Sun (1976). These chondritic ratios were computed from the data given in van Schmus and Hayes (1974) and Wänke et al. (1974). The range and average $Ga/Al x 10^5$ ratio for 72 chondrites was taken from Willis (1978) and the Ti/Sc ratio was computed from the data given in Larimer (1971) and Taylor (1964). Also listed for comparison with the group I Onverwacht lavas are inter-element ratios obtained from olivine spinifex textured ultramafic rocks from the Yilgarn Block and Munro Township areas (Nesbitt and Sun, 1976). Inter-element ratios from these ultramafic rocks are essentially identical to the group I Onverwacht lavas.

Inter-element ratios similar to the group I Archaean lavas are found in young lavas from the ocean floor, such as from the Mid-Atlantic Ridge (e.g. from $29^{\circ}N$, Sun et al., 1979, and $36^{\circ}47'$ N, the FAMOUS project area, Langmuir et al., 1977) and from back-arc basins (e.g. Saunders and Tarney, 1979). Le Roex (1980) in a detailed study of the basalts sampled in the FAMOUS project has computed a primary magma composition from melt inclusions in picrite basalt spinels. Inter-element ratios for this postulated primary magma are given in Table 7-6 and are similar to inter-element ratios computed for other lavas analysed by Le Roex (1980) and other workers (e.g. Langmuir et al., 1977) from the FAMOUS area. Data from back-arc basins will be considered in more detail in the next section

but it is noted here that certain lavas from back-arc basins (Saunders and Tarney, 1979; Dietrich et al., 1978) have similar inter-element ratios to the FAMOUS lavas and to the Archaean group I rocks.

All the lavas in group I have Al/Ti, Ca/Ti, Ti/Zr, Ti/Y, Ti/Sc and Zr/Y ratios which are similar to those in chondrites suggesting that the source material for these lavas also had chondritic ratios for these elements. If this is correct then this group I source composition has yielded lavas since the early Archaean (Onverwacht volcanics have been dated at 3540 Ma by Hamilton et al., 1979) and possibly continuously up to the present time (FAMOUS lavas are believed to be <21,000 years old, Hekinian et al., 1976).

The Onverwacht lavas in group II are volumetrically the most abundant type of lava in the LUU and consist of ultramafic komatiites, high-Mg mafic komatiites, low-Mg mafic komatiites and high-Ti basalts. Inter-element ratios in individual lava types and between average lava types show a very restricted range of values. In general the average inter-element ratios of group II lavas are distinctly different to the chondritic values (Table 7-6). They are however similar to those computed for typical Mid-Ocean Ridge basalts (MORB) using the data given in Melson and Thompson (1971) and Erlank and Reid (1974). Specific examples are given in Sun et al. (1979) from the East-Pacific rise and the Mid-Indian Ocean Ridge.

In the group II komatiites the Ca/Ti ratio remains constant from 33% to 14% MgO (22±1) and decreases to 16 in the evolved low-Mg mafic komatiites (9% MgO). This decrease in Ca/Ti ratio is consistent with the onset of clinopyroxene fractionation at 14% MgO in the komatiite magmas as proposed in Chapter 6. The average high-Ti basalt has a lower Ca/Ti ratio (Fig. 7-6) than the evolved low-Mg mafic komatiites and is similar to that of MORB. As other inter-element

ratios (e.g. Al/Ti, Ti/Zr and Ga/Al) are essentially constant through the MgO range of the group II rocks, the decrease in the Ca/Ti ratio and increase in the Ti/Sc ratio in both the high-Ti basalts and typical MORB are attributed to:-

either some clinopyroxene being left in the residue after melting the source material,

or clinopyroxene fractionating from the basaltic magmas as they travelled from source to surface.

The consistency of most of the inter-element ratios (e.g. Al/Ti, Ti/Zr, Ti/V, Ga/Al, Zr/Y and V/Zr) in the group II lavas is believed to be due to the derivation of these lavas from similar source compositions in both the Archaean and recent times. Inter-element ratios in some back-arc basin basalts (e.g. Saunders and Tarney, 1979) are similar to inter-element ratios of the group II rocks and possibly these lavas were derived from similar source materials.

The group III rocks (Table 7-6) consist of the high-Mg basalts and have significantly different inter-element ratios from the group I and II lavas. The three high-Mg basalts analysed are holocrystalline, relatively coarse grained rocks that contain abundant clinopyroxene and plagioclase. Ratios involving Al, Ca, Sc and V may therefore have been modified if these samples accumulated pyroxene and or plagioclase. As no fine grained aphyric samples of this type of basalt have as yet been analysed from the Onverwacht Group it cannot be stated with any conviction that mantle derived magmas existed with similar inter-element ratios.

The high-Mg basalts from the LUU nevertheless show some striking similarities to boninite rocks recently dredged from the Mariana

trench (Dietrich et al., 1978). Both rock types have high MgO contents (>13% on average), high SiO_2 contents (>52%), very low TiO_2 contents (<0.4%) and relatively high-Cr contents. There are also many similarities in inter-element ratios (Table 7-6) such as higher Al/Ti and lower Ti/Zr and Ti/Sc ratios than either group I or II lavas. In addition boninites and high-Mg basalts have very low $CaO-Al_2O_3$ ratios of 0.54 and 0.58 (on average) respectively. It is suggested from these similarities that the LUU high-Mg basalts and boninites are the same rock type. Cameron et al. (1979) have recently suggested that boninites represent "the closest Phanerozoic equivalent to basaltic komatiite". However from the characteristics outlined for boninites above, it is considered that they have compositions more closely associated with the Onverwacht high-Mg basalt than with mafic komatiites.

All three groups of lavas in Table 7-6 have higher Ti/V and lower V/Zr and Ga/Al ratios than the average chondrite. Nesbitt and Sun (1976), assuming a chondritic earth, proposed that the mantle was depleted in V during the formation of the core, but in a later paper (Sun and Nesbitt, 1977) consider this proposal to be unlikely. Sun and Nesbitt (1977) favour the retention of some V in residual phases during partial melting. The apparent depletion of Ga in the Onverwacht lavas compared to chondrites could also be due to this element being retained in a residual phase, such as chromite, during the melting events that formed the lavas. Alternatively removal of a metal phase from the mantle during core formation could have depleted the mantle in Ga. Willis (1978) has shown that the metal fraction of certain types of meteorite is significantly higher in Ga than the silicate fraction which suggests that Ga preferentially enters the metal phase and therefore could have been depleted in the upper mantle. However,

as $D^{crystal/liq}$ for V and Ga have not been determined at the very high temperatures expected for the generation of komatiites, no attempt has been made to eliminate one of the alternatives proposed above.

From consideration of inter-element ratios of the variety of rock types occurring in the Onverwacht Group the observations have been made that the mantle was heterogeneous in the Archaean on the scale sampled by magmatic processes. This is in agreement with the conclusions of other workers (e.g. Sun and Nesbitt, 1977; Nesbitt et al., 1979) and the discussion in Chapter 5. Erlank et al. (1980) have argued on the basis of $^{87}Sr/^{86}Sr$ initial ratios that heterogeneity existed in the sub-southern Africa mantle 3.0 Ma ago and was well established 2.0 Ma ago. The Onverwacht data indicate that heterogeneity in the mantle had developed on the scale sampled by magmatic events at least 3.5 Ma ago.

The group I low-Ti basalts have similar inter-element ratios to the rare group I ultramafic komatiites (see Table 7-6 and Fig. 7-6) while the group II high-Ti basalts have similar inter-element ratios to the abundant group II ultramafic komatiites. It has been demonstrated in an earlier section that the basalts could not have been derived from the ultramafic komatiites by low pressure crystal fractionation processes. Rather the tholeiitic lavas are considered to have been derived by partial melting of source material similar in composition to that from which the relevant ultramafic komatiites were derived. However, in order to account for the low Ca/Ti and high Ti/Sc ratios of the high-Ti basalts compared to the group II ultramafic komatiites, it is necessary to postulate that either some clinopyroxene remained in the residue (after the melting event that generated the high-Ti basaltic magmas) or that clinopyroxene crystallised and settled as the

basaltic magma rose to the surface. Petrogenesis of the tholeiitic magmas is further discussed in Chapter 8.

RARE EARTH ELEMENT PATTERNS IN ONVERWACHT GROUP LAVAS

In recent years interpretation of rare earth element (REE) patterns in volcanic rocks has become a popular method for elucidating partial melting and crystal fractionation processes in magmas. As yet REE data have not been obtained on a representative suite of Onverwacht Group lavas. However, some REE analyses for Onverwacht Group lavas have been reported in the literature (e.g. Herrmann et al., 1976; Condie et al., 1977; Sun and Nesbitt, 1978; Hamilton et al., 1979). These data have been obtained by different analytical techniques which have varying precision. The samples analysed by the various authors show differing degrees of alteration and consequently the data are of differing reliability when attempting to examine igneous processes that have affected the REE distribution patterns in the lavas. However, Hamilton et al. (1979) have obtained precise Nd and Sm concentrations (by isotope dilution techniques) from a number of selected samples (that are relatively well preserved) from the Onverwacht Group. Several of these samples are considered to be representative of phenocryst free magma compositions. These data will be discussed below in conjunction with REE analyses of two spinifex textured ultramafic komatiites from the LUU obtained by Sun and Nesbitt (1978).

Two samples(HSS-95, an ultramafic komatiite and HSS-56, a low-Ti basalt) analysed for Nd and Sm by Hamilton et al. (1979) and 49-J analysed for REE by Sun and Nesbitt (1978) are group I rocks. These three samples have an average Sm_n/Nd_n (chondrite normalised ratio using the average chondrite REE values given in Evensen et al., 1978) of 0.99.

The chondrite normalized REE pattern of sample 49-J (see Fig. 7-7) is flat. The Sm_n/Nd_n of ∿1 and the flat REE pattern of 49-J indicates that group I lavas have chondritic relative REE abundances which is consistent with the other chondritic ratios (Table 7-6) displayed by this group of rocks from the Onverwacht. The spinifex textured ultramafic komatiites from Munro Township (Arth et al., 1977) and Yakabindie (Sun and Nesbitt, 1978) which are also believed to be group I lavas have light REE depleted patterns ($Sm_n/Nd_n > 1$). $^{143}Nd/^{144}Nd$ isotopic compositions of the Munro Township lavas (Zindler et al., 1978) suggest that this depletion event could have occurred any time from 4550 Ma (assumed age of the earth) to the time the lavas were generated. Nd isotopic constraints will be considered in more detail in the next section.

Hamilton et al. (1979) have also determined the Nd and Sm contents of some group II lavas. These samples include three ultramafic komatiites (average $Sm_n/Nd_n = 0.91$), two evolved low-Mg mafic komatiites, one of which is the pillow sample SC-11 for which other REE's were determined as well (average $Sm_n/Nd_n = 0.92$) and one high-Ti basalt ($Sm_n/Nd_n = 0.95$). Sample 331/78, an ultramafic komatiite from the LUU (Sun and Nesbitt, 1978), belongs to the group II rocks and has a $Sm_n/Nd_n = 0.90$. The Sm_n/Nd_n ratios of the group II rocks (0.90 - 0.95) and the REE pattern of samples 331/78 (Sun and Nesbitt, 1978) and SC-11 (Hamilton et al., 1979) indicate that the group II rocks are light REE enriched relative to chondrites and the group I rocks (see Fig. 7-7). Typical MORB on the other hand generally has light REE depleted patterns (e.g. Gast, 1968) and in this respect differs from the other group II lavas. Green et al. (1975) and Sun and Nesbitt (1978) have argued for the fractionation of garnet from the LUU komatiite magmas

to account for their high CaO/Al_2O_3 ratios and REE patterns. However, if the suggestion made earlier is correct (that all group II lavas have been derived from a similar source composition) then garnet fractionation from the komatiite magmas is considered unlikely as the Al/Ti ratios of komatiites should be lower than the high-Ti basalts. The problem of garnet fractionation in komatiitic magmas is considered in more detail in Chapter 8.

DISCUSSION OF EVIDENCE FOR MANTLE HETEROGENEITY

It is pertinent at this point to summarise the evidence outlined in the preceding discussion for a heterogeneous mantle in the Archaean. Inferences based on similarity of inter-element ratios of ancient and modern basaltic lavas suggest that a similar heterogeneous mantle is still yielding comparable lavas at present. However, anomalies arise when comparing compositions proposed for modern mantle with those proposed for the Archaean. These anomalies are briefly discussed and possible explanations for the differences considered.

On the basis of inter-element ratios displayed by the Onverwacht lavas two main groups have been recognised as summarised below (see also Fig. 7-8):-

	Range of MgO	Al/Ti	Ti/Zr	V/Zr	REE pattern (chond. norm.)
Group I	33 - 7%	∿20	∿110	∿6	Flat
Group II	33 - 6%	∿10	∿ 80	∿4	Light REE enriched

As a very wide range of magma compositions are involved in both groups, these differences are presumed to reflect differences in the source materials from which the lavas were derived.

Sun et al. (1979) have suggested an alternative explanation in that they consider typical MOR basalts with ∿0.7% TiO_2 (with Al/Ti ratios of ∿20 and other ratios that correspond to the group I rocks above) to have been derived by relatively high degrees of partial melting (∿25%) of a pyrolite source composition. They suggest at this degree of melting major Al- and Ca-bearing minerals were eliminated from the residue. Sun et al. (1979) further consider that typical MOR basalts with ∿1.5% TiO_2 (with Al/Ti ratios of ∿10 and other ratios that correspond to the group II rocks above) were derived by lower degrees of melting (∿15%) of a pyrolite source in which major Al- and Ca-bearing phases were not completely eliminated from the residue. This explanation is considered unsatisfactory for accounting for the Al/Ti variations between the group I and II rocks in Table 7-6 as magmas with wide ranges of compositions (presumably reflecting differing degrees of single or multi-stage melting) within each group maintain consistent Al/Ti ratios. As stated previously the preferred explanation is that the two groups of lavas were derived from separate source compositions.

However, the Ca/Ti and Ti/Sc ratios in the lower MgO (<14%) group II lavas show regular changes with decreasing MgO content. In order to account for these variations it has been proposed that clinopyroxene crystallised and settled from the magmas (mafic komatiites and possibly the tholeiites) before extrusion or that clinopyroxene was not eliminated from the residue (tholeiites) during partial melting. In contrast the Ca/Ti and Ti/Sc ratios of the group I lavas show no systematic changes with MgO content suggesting that clinopyroxene was eliminated from the residue and did not fractionate from the magmas before they were extruded.

A further significant difference between the Onverwacht group I and II rocks is that absolute abundances for elements such as Ti, Zr, Y and light REE are higher by a factor of 2 or more in group II lavas compared to group I lavas with similar MgO contents. These differences are illustrated for Ti and Zr in Fig. 7-9. From these differences it is inferred that the group II source material had significantly higher Ti, Zr, Y and light REE contents than the group I source material.

Within the range of typical MOR basalt compositions, lavas occur that correspond to either group I or group II rocks based on inter-element ratios and similar absolute abundances of certain elements (see Tables 7-6 and 7-7 and the data in Sun et al., 1979). Inferences drawn from these similarities are that the source compositions proposed for the Archaean group I and II rocks have modern day analogues from which ocean floor magmas are being derived.

Typical MORB is generally considered to have been derived from source material that had previously been depleted in large ion lithophile (LIL) elements (Tatsumoto et al., 1965; Gast, 1968; Hart, 1971; Hart et al., 1972; Erlank and Kable, 1976; Sun et al., 1979). The evidence for the depleted nature of MORB was initially based on arguments concerning Sr, Pb and more recently Nd isotopic compositions of the lavas, as well as alkali element contents and inter-element ratios (e.g. K/Rb, K/Cs, Rb/Sr) and REE patterns. Erlank and Kable (1976) suggest that MORB with Zr/Nb ratios ranging from 30-110 have been derived from source material that experienced previous episodes of depletion. The group II Onverwacht tholeiites have an average Zr/Nb ratio of 27 suggesting that they (and by inference all the Onverwacht group II lavas) were derived from source compositions depleted in Nb relative to Zr.

The alkali elements and Sr isotopic compositions of the Onverwacht lavas are believed to have been seriously perturbed by later alteration and or metamorphic events. However, the REE contents and Nd isotopic compositions of selected lavas are believed to have remained relatively unaffected by these events (Hamilton et al., 1979) and as such can be used for comparison purposes with MORB. The REE patterns constitute a major difference between the Archaean source compositions (group I and II) and those proposed for typical MORB. The lavas from the Onverwacht have either flat REE patterns (group I) relative to chondrites or light REE enriched patterns (group II) while typical MORB have light REE depleted patterns regardless of whether they have other inter-element ratios characteristic of group I or II lavas. In addition the group II rocks (both modern and Archaean) contain significantly higher abundances of light REE (and Ti, Zr and Y) than the corresponding group I rocks and cannot therefore be considered as depleted in these elements relative to group I lavas or chondrites. Several possible explanations for these features are listed below:-

(a) The similarity of ratios (Table 7-6 and 7-7) between Onverwacht lavas and MORB are fortuitous and inferences drawn from these data about mantle compositions are invalid.

(b) The differences between the ancient and modern lavas within each group reflect the natural range of group I and II source compositions.

(c) Relatively small scale events (such as very low degrees of partial melting) have affected the group I and II source compositions since the Archaean and the concentrations of

incompatible elements (such as the light REE) were significantly decreased without changing other parameters of the source such as Al/Ti.

Possibility (a) is considered unlikely as not only are many inter-element ratios similar but also absolute abundances (e.g. see Table 7-7 and Sun et al., 1979) of elements. There is some evidence for possibility (b) in that other Archaean lavas show a range of REE patterns from slightly light REE enriched to depleted (Sun and Nesbitt, 1979). This suggests that the REE patterns are not unique to each source while Al/Ti, Ti/Zr etc. are. Although this possible explanation for the different REE patterns cannot be discounted it is not favoured, as the source compositions could have been depleted in light REE as suggested below. Possibility (c) is favoured as small degrees of partial melting (and extraction of the melt) can deplete the source of most of its light REE content without changing the Al/Ti and Zr/Ti ratios of the residue to any marked extent. In Fig. 7-10 the percentage depletion of the elements K, Rb, Sr, Nb, Ti, Zr, Y, Y, La, Nd, Sm and Yb have been calculated for the residue after melts, representing small degrees of partial melting, have been extracted using the following equation:-

$$\% \text{ Depletion} = 100 \, (C_o - C_{residue})/C_o$$

The equation $C_1/C_o = 1/(D+F\,(1-D))$ from Arth (1976) was used to calculate the concentrations of the elements in the extracted liquids. Distribution coefficients were selected from the literature and mainly from Philpotts and Schnetzler (1970), Hart and Brooks (1974), Frey et al. (1978), Pearce and Norry (1979) and Le Roex (1980). The initial mantle was assumed to consist of 60% olivine, 20% orthopyroxene, 10%

clinopyroxene and 10% garnet. The bulk distribution coefficients calculated for these mineral proportions are given in Fig. 7-10. The range of melting considered is assumed to have been sufficiently small not to have significantly affected the mineral proportions given above.

As can be seen in Fig. 10, if a 0.5% melt of the initial source was completely removed, the residue would be depleted of $\geq$50% of its La, K and Rb content, $\sim$30% of its Nd and Sr content, $\sim$18% of its Nb and Sm content but <8% of its Ti, Zr, Y, V and Yb content compared to the initial source composition. The individual amounts of major elements (Si, Al, Fe, Mg and Ca) removed by such a melt would be <<0.5 wt.% and the residue would have essentially the same major element composition as the initial mantle. Inter-element ratios in the residue such as Al/Ti, Ca/Ti would increase by $\sim$8% (e.g. from 20 to 21.6), while there would be no change of the Ti/Zr ratio between the initial source and residue as, for the distribution coefficients selected, both elements are equally partitioned into the liquid, and the Zr/Y ratio would be slightly lower in the residue than in the initial source. However, the REE pattern of the residue relative to the initial source would be significantly light REE depleted. This is illustrated, considering La and Yb, in Fig. 7-10 as La is depleted in the residue by $\sim$70% while the Yb content of the residue is only depleted by $\sim$1% compared to the initial source composition. If at some later stage this residual mantle melted ($\sim$20 - 25% so that no major Al- or Ca-bearing phases were left in the residue) and gave rise to tholeiitic magmas, these magmas would have essentially identical Al/Ti, Ca/Ti, Ti/Zr and Zr/Y ratios and similar Ti and Zr contents as tholeiitic magmas derived in the same way from the initial source before it was depleted. However, the tholeiitic magmas derived from the residual mantle would have light REE

depleted patterns, lower K and Rb contents and depending on the timing of the second but major melting event,the magmas may also be depleted in radiogenic Sr and enriched in radiogenic Nd compared to the tholeiites derived from the initial mantle.

In this simple model the effects of minor mantle phases on the distribution of the elements have not been considered, neither is it known whether 0.5% melts could be extracted from the source material. More complex models can be constructed that qualitatively have the same effect on the residual composition such as small but higher degress of melting (say ∿5%) and extraction of 1 - 2% of this melt. Nevertheless, the model does illustrate the importance of small scale melting and melt removal events on residual mantle compositions.

For simplicity if only group II tholeiites are considered (the same arguments can be applied to the group I tholeiites) it is proposed that Onverwacht lavas were derived from an initial source material while the MORB equivalents were derived from the same initial source material that had experienced a previous small scale depletion event as outlined above. This model does account for the known similarities (e.g. Al/Ti and Ti/Zr) and differences (e.g. REE patterns) between the two sets of lavas. However, it should be noted that implicit in this proposal is that the precursor source material (before depletion) for group II MORB did not have chondritic ratios for the elements considered. On the other hand the precursor source material (before depletion) for a group I MORB had chondritic ratios for many of the elements. Further data and in particular reliable K, Rb, Sr, Ba and REE contents on the Onverwacht tholeiites are required to develop and test these ideas.

7-5 $^{143}Nd/^{144}Nd$ ISOTOPIC COMPOSITIONS

Onverwacht Group samples from both group I and II rocks have been used to define a Sm-Nd isochron (Hamilton et al., 1979) which gave an initial $^{143}Nd/^{144}Nd$ of 0.50809±4 (2σ) and an age of 3540±30 Ma. Archaean lavas from other greenstone belts have also been dated by the Sm-Nd method (see review by O'Nions et al., 1979) and in Fig. 7-11 the initial $^{143}Nd/^{144}Nd$ obtained from the isochrons have been plotted against the determined age of the lavas. Also included on Fig. 7-11 are initial $^{143}Nd/^{144}Nd$ isotopic compositions for kimberlites (Basu and Tatsumoto, 1979) of various ages. All these points plot on, or very close to the chondrite evolution line drawn from the initial $^{143}Nd/^{144}Nd$ ratio (0.50682±5) of the achondrite Angra dos Reis (Lugmair and Marti, 1977) at 4550 Ma to the present estimated bulk earth (0.51262, O'Nions et al., 1979). The latter was calculated (O'Nions et al., 1979) assuming the earth's initial $^{143}Nd/^{144}Nd$ was the same as Angra dos Reis and that the bulk earth has a chondritic Sm/Nd ratio. The correspondence of initial $^{143}Nd/^{144}Nd$ isotopic ratios of rocks from a wide range of ages with the chondrite evolution line, suggests that the source material of these lavas had existed since the formation of the earth with chondritic Sm/Nd ratios. This is in agreement with the conclusions of de Paolo and Wasserberg (1976a,b) that the Archaean mantle evolved with approximately chondritic Sm/Nd.

Two source compositions have been proposed for the lavas of the Onverwacht Group and the question arises as to how long these sources could have existed prior to magma generation. Lavas from group I have chondritic ratios for certain elements (Table 7-6) and chondrite relative REE abundances (Fig. 7-7). As the Sm_n/Nd_n ratios of these lavas appear to be ~1 and believed to be the same as the source material,

the $^{143}Nd/^{144}Nd$ of the source would evolve along the chondrite evolution line (Fig. 7-11). The group I source composition therefore could have existed since the formation of the earth (model age of 4550 Ma).

Group II lavas (Table 7-6) have non-chondritic ratios for many elements and Sm_n/Nd_n ratios less than the chondritic value. If the assumption is made that the source material for these lavas had the same $^{147}Sm/^{144}Nd$ composition (0.1784) as the ultramafic rocks in group II (data for HSS-88A, HSS-92 and HSS-523 in Hamilton et al., 1979) and an initial $^{143}Nd/^{144}Nd$ composition as for Angra dos Reis 4550 Ma ago, the expected initial $^{143}Nd/^{144}Nd$ at 3540 Ma would be 0.50803±5. This value is within analytical error of the initial $^{143}Nd/^{144}Nd$ (0.50809±4) obtained from the isochron for the Onverwacht Group lavas (Hamilton et al., 1979). This means that no time constraints can be placed on when the group II source was formed within the present precision of the data. The group II source for the Onverwacht lavas may therefore have existed since the formation of the earth or have formed (possibly from the chondritic source) at any time up to 3540 Ma ago before undergoing melting to yield group II lavas.

7-6 TECTONIC SETTING OF THE BARBERTON GREENSTONE BELT

In recent years detailed work has been undertaken in attempting to establish relationships between major and trace element abundance and the tectonic environments in which lavas were erupted (e.g. Cann, 1970; Pearce and Cann, 1971,1973; Miyashiro, 1974;Miyashiro and Shido, 1975; Pearce, 1976; Gill, 1977). Many assumptions must be made when attempting to infer tectonic settings for Archaean lavas using geochemical parameters established for modern tectonic environments. These assumptions include (inter alia):-

(a) Similar mineralogical and chemical source compositions were involved in the production of Archaean and modern lavas.

(b) Conditions of melting were similar.

(c) Only one process exists that can generate a particular geochemical feature and that it is diagnostic of a specific tectonic environment.

(d) Plate tectonic processes were in operation in the Archaean.

Comparison of geochemical features in Archaean and modern lavas should be restricted to basalts as komatiites are rare in post Archaean volcanic rocks. Evaluation of the major element data (Table 7-1a and 7-2a) from the Onverwacht basalts using AFM (Irvine and Baragar, 1971) and FeO-FeO/MgO plots (Miyashiro, 1974), indicate that these rocks belong to the tholeiitic suite. In previous sections of this Chapter it has been argued, on the basis of selected inter-element ratios, that for the low-Ti and high-Ti basalts modern day analogues can be found in Mid-Ocean Ridge basalts. Ti-Zr (Fig. 7-12)and Ti-Zr-Y (Fig. 7-13) relationships from Pearce and Cann (1973) indicate an association of the Onverwacht Group basalts with Ocean floor basalts and low-K tholeiites. Conventional interpretation of this data suggests that the Onverwacht Group lavas could have been extruded in a similar tectonic environment to Mid-Ocean Ridge basalts. However, the REE patterns of both the group I and II Onverwacht Group lavas indicate that the mantle from which they were derived was not depleted in light REE as has been proposed for typical MORB sources (Gast, 1968; Richard et al., 1976; O'Nions et al., 1977).

An analogy can also be drawn between Onverwacht Group and back-arc basin lavas both geologically (Tarney et al., 1976; Weaver and Tarney, 1979) and geochemically. Recent work on lavas from back-arc basins has shown them to be similar to MORB except for higher K, Rb, Cs, Sr, Ba, REE's and $^{87}Sr/^{86}Sr$ (Hart et al., 1972; Gill, 1976; Hawkesworth et al., 1977; Saunders and Tarney, 1979). The REE patterns in back-arc basin basalts vary from being flat to light REE enriched relative to chondrites. It has been proposed that some of the differences in geochemistry between back-arc basin basalts and MORB (e.g. higher $^{87}Sr/^{86}Sr$ ratios in back-arc basalts) could be related to contamination by sea water (Hawkesworth et al., 1977).

In Table 7-7 inter-element ratios and absolute element abundances (TiO_2, Zr and Y) for the high-Ti and low-Ti basalts from the Onverwacht Group are compared to samples from the Scotia Sea Rise for which major element (Saunders and Tarney, 1979) and REE data (Hawkesworth et al., 1977) are available. From Table 7-7 it can be seen that both basalt types found in the Onverwacht Group have analogues in the lavas from the Scotia Sea Rise. This comparison can be extended to the ophiolites of southern Chile (Saunders et al., 1979) where gabbros, dolerites and basalts have inter-element ratios and element abundances (as shown by TiO_2, Zr and Y) that correspond to group I and II rocks from the Onverwacht Group. The southern Chile 'rocas verdes' have been proposed as possible modern day analogues of Archaean greenstone belts (Tarney et al., 1976; Weaver and Tarney, 1979). Aspects of the geochemical data, as outlined above, show that the Onverwacht Group lavas have many features similar to back-arc basin lavas and to the 'rocas verdes' and as such is consistent with the model of Tarney et al. (1976) and Weaver and Tarney (1979). However, whether this

geochemical similarity means that greenstone belts originate in tectonic environments similar to back-arc basins is still open to debate. The most obvious problem with this analogy is that no ultramafic or mafic komatiites have as yet been recorded from back-arc basins, although this could be related to the changing geothermal regimes in the mantle since the early Archaean.

The similar inter-element ratios and absolute abundance of elements in back-arc basin basalts and the Onverwacht group basalts suggest that similar source compositions and conditions of melting were involved in the production of the two suites of lavas. This provides support for assumptions (a) and (b) listed earlier in this section. However, whether these geochemical features can only be produced in one specific tectonic environment (assumption (c)) remains an open question. In conclusion it is noted that while it cannot be convincingly demonstrated that the Onverwacht Group lavas were generated in a back-arc tectonic environment, the indications that this is possible are encouraging and further detailed comparisons of both geological and geochemical features of the two volcanic associations are warranted.

7-7 CONCLUSIONS

While tholeiitic basalts from the Onverwacht Group have only formed a relatively minor part of this project several important conclusions have been made from interpretation of the available data. These are summarised below:-

(1) The Al_2O_3 content can be used in lavas from the Onverwacht Group and probably in most other greenstone belts to distinguish between tholeiitic basalts and komatiites.

(2) Three types of basalt have been identified (low-Ti, high-Ti and high-Mg basalt) from the Onverwacht Group on the basis of major and trace element geochemistry.

(3) None of the basalt types are related to each other or to low-Mg mafic komatiites by low pressure crystal fractionation or accumulation processes.

(4) On the basis of inter-element ratios (both komatiites and basalts) three groups of lavas have been identified from the Onverwacht Group. Group I consists of relatively low-Ti lavas (the rare low-Ti ultramafic komatiites, aphyric high-Mg mafic komatiites and the common low-Ti tholeiitic basalts), group II rocks consist of the relatively high-Ti lavas (the abundant ultramafic komatiites, porphyritic high-Mg mafic komatiites, aphyric and porphyritic low-Mg mafic komatiites and the common high-Ti tholeiitic basalts) and the group III lavas consisting of the high-Mg basalts which are believed to be similar in composition to boninites.

(5) Inter-element ratios and REE patterns of the group I rocks are similar to chondrites, some Mid-Atlantic Ridge basalts (including those from 39°N - the FAMOUS project area), and certain back-arc basin lavas, suggesting that similar source compositions that gave rise to some of the Onverwacht volcanics are yielding basaltic magmas today.

(6) Group II rocks have inter-element ratios similar to those of typical MORB and similar inter-element ratios and REE patterns to many back-arc basin lavas. It is also suggested

that the source mantle to group II lavas may also have existed or has been periodically generated since early Archaean times.

(7) Inter-element ratios and absolute abundances of certain elements in the basalts are consistent with the model that the Onverwacht Group volcanics were extruded in a back-arc basin tectonic environment. However, serious problems must be investigated before a back-arc basin origin for greenstone belts can be unambiguously proven.

CHAPTER 8.

PETROGENESIS OF BARBERTON GREENSTONE BELT LAVAS

8-1 INTRODUCTION

In recent years detailed investigations of mafic lavas have led to the development of numerous models for processes of magma generation in the mantle. Komatiites have also come in for their share of imaginative thought and many models have been proposed for their origin. Viljoen and Viljoen (1969j) consider that ultramafic komatiite magmas represent liquids derived by high degrees of partial melting of their source material. The expected high temperatures required for the generation of these magmas have been confirmed experimentally (Green et al., 1975). Green (1972a,b) has suggested that major meteorite impacts on the earth's surface $3\text{-}4\times10^9$ years ago were directly related to the generation of ultramafic magmas (impact triggered diapirism) and the formation of greenstone belts. Most recent models proposed for the generation of ultramafic komatiites are less catastrophic in nature, but in general do postulate the upwelling of mantle diapirs from depths as great as 400 kms. (Green, 1975; Nesbitt and Sun, 1976; Arndt, 1977c; Arth et al., 1977; Bickle et al., 1977; Naldrett and Turner, 1977; Nisbet et al., 1977; Nesbitt et al., 1979; Weaver and Tarney, 1979). Such depths for the initiation of upwelling have been proposed to account for the high extrusion temperatures of the ultramafic komatiite lavas.

Clearly energy sources are important when considering the generation of ultramafic magmas and although it is not intended here to investigate possibilities, many sources have been suggested (see discussion in Wetherill, 1972). Heat produced by the radioactive decay

of K, U and Th was higher in the Archaean than at present. Other processes such as the radioactive decay of short lived isotopes, hot accretion of the earth, capture of the moon or core formation, are more speculative and no consensus of opinion exists as to the possible contribution of each of these processes to the early Archaean energy budget. However, it appears to be a generally accepted assumption that geothermal gradients in the Archaean were steeper than those deduced for the earth today (Birch, 1952; Green, 1975; Burke and Kidd, 1978). Nevertheless it has still been found necessary to postulate upwelling of mantle material from great depths to satisfy thermal constraints for the production of ultramafic magmas.

In previous Chapters (5, 6 and 7) fractional and equilibrium crystallisation models have been considered for obtaining the various lava compositions. This Chapter deals with high pressure melting and/or crystallisation processes that may have been involved in the generation of Onverwacht Group lavas, and the following lava types are considered:-

(a) ultramafic komatiites (group I and II)

(b) mafic komatiites (group I and II)

(c) tholeiitic basalts (group I and II).

It should be noted that only lavas that are believed to represent essentially phenocryst-free magma compositions are considered in the ensuing discussions.

8-2 ULTRAMAFIC KOMATIITES

A number of models have been proposed in the literature for the generation of ultramafic komatiite magmas. Some of these postulate

high degrees of partial melting of undepleted mantle material to obtain the high MgO contents of the magmas (Viljoen and Viljoen, 1969j; McIver and Lenthall, 1973; Cawthorn and Strong, 1975; Cawthorn, 1975; Bickle et al., 1976; Nesbitt and Sun, 1976; Nesbitt et al., 1979). Arndt (1977c) however considers that it would be unlikely that all of the liquid would remain in contact with the residual crystals until the required degree of melting had been reached. He proposes that a more likely model would be the melting of a residuum of refractory minerals with some trapped liquid left from a previous melting episode. The advantage of Arndt's sequential melting is that lower degrees of partial melting of depleted material could produce ultramafic magmas.

On the other hand, it has been suggested that ultramafic komatiites could have evolved by polybaric assimilation and melting of overlying mantle material during diapiric uprise (Bickle et al., 1977) or by mechanical mixing of initial melt and entrained residual crystals, the latter dissolving as the crystal mush ascended (Cox, 1978).

The above models will be assessed from the point of view of which possible mineral phases could have controlled the ultramafic magma compositions. Geochemical trends illustrated in Chapter 5, available high pressure experimental data (Arndt, 1976; Bickle et al., 1977) and projections from mineral compositions onto convenient planes in CMAS space (O'Hara, 1968; Cox et al., 1979) are used to constrain the models.

DUNITE RESIDUE

The suggestion that ultramafic magmas originate by high degrees of partial melting of mantle material with olivine as the only residual phase (Viljoen and Viljoen, 1969j; Nesbitt and Sun, 1976; Sun and

Nesbitt, 1977), is supported by the available experimental data. Arndt (1976) has shown that olivine is the only liquidus phase of an ultramafic komatiite (SA-3091 with ∿25% MgO) up to 40 kb. The second liquidus phase that crystallised with olivine was calcium-poor pyroxene (20 - 40 kb.) or spinel (<20 kb.). Olivine dominated as the sole liquidus phase for a temperature range of 100 - 200°C below the liquidus at all pressures. Bickle et al. (1977) have also shown that for lavas that are here considered to be ultramafic (i.e. >24% MgO) olivine was the only liquidus phase at 30 and 40 kb. One sample however (NG-7638 containing 24.4% MgO) had orthopyroxene as the only liquidus phase at 40 kb. Another sample (NG-7621 containing 20.5% MgO) had orthopyroxene as the liquidus phase from 15 - 30 kbs. Olivine and orthopyroxene never crystallised together at pressures up to 40 kb. (Bickle et al., 1977).

It would therefore be expected that olivine would be the only phase in equilibrium with the higher MgO (>25%) ultramafic komatiite magmas at 30 - 40 kb. pressure. This implies that if these magmas originated at 100 - 120 kms. depth by partial melting with or without subsequent crystal fractionation, olivine should be the only mineral controlling this differentiation. Geochemical trends developed in the 25 - 33% MgO lavas should reflect olivine-only control if the magmas originated by equilibrium partial melting. Nesbitt and Sun (1976) and Sun and Nesbitt (1977) consider olivine as the only mineral phase involved in the differentiation of Archaean ultramafic magmas. Duke and Naldrett (1978) also consider that olivine crystallisation from a parental magma (32% MgO) could give rise to the suite of komatiite lavas.

However, examination of the geochemical trends developed in the Barberton group II ultramafic komatiites (e.g. CaO-MgO, Fig. 5-8) shows that olivine differentiation cannot account for the range of ultramafic

lava compositions. This lack of pure olivine control is also reflected in the projections from olivine and enstatite onto planes indicated in Fig. 8-1 and 8-2. Geochemical trends (e.g. TiO_2-MgO) and inter-element variations (Al/Ti) developed in the Belingwe greenstone belt ultramafic lavas (non-cumulate rocks), also demonstrate that fractionation of olivine alone could not have controlled the range of lava compositions (Bickle et al., 1976, 1977). The variations of TiO_2 and CaO with MgO from the Munro Township ultramafic rocks (Fig. 5-19) also eliminate the possibility of fractionation of only olivine in these lavas.

On the other hand trends developed in the group I ultramafic komatiites (Fig. 5-15) from Barberton are consistent with olivine control. In the olivine projection (Fig. 8-1) the two samples that are believed to represent liquid compositions plot very close together and on the enstatite (Fig. 8-2) and diopside (Fig. 8-3) projections these two samples lie on trends that would appear to project through olivine. Olivine control on these trends, either as the only mineral melting as the MgO content of the melt increased from ∿25 to 33% MgO, or by crystallisation from a high MgO parent (∿33%) magma, is compatible with the limited data available for the Barberton group I ultramafic komatiites. These lavas also have similar inter-element ratios to some of the Western Australian ultramafic komatiites (Table 7-6) for which Nesbitt and Sun (1976) and Sun and Nesbitt (1977) have proposed an origin by olivine-only differentiation. The data from the Barberton ultramafic komatiites suggest that the ultramafic lavas could be generated by two different processes. Alternatively the group I and II ultramafic komatiites may have originated from two different source compositions by the same process. This latter possibility is favoured, as discussed in Chapter 7.

HARZBURGITE RESIDUE

The suggestion has been made that ultramafic komatiites could have been generated by the melting of a residue of olivine and orthopyroxene (Cawthorn and Strong, 1975; Nesbitt and Sun, 1976; Bickle et al., 1977; Nesbitt et al., 1979). For this model to be valid the magmas must have originated at well above 40 kbs. pressure, as at lower pressures olivine is the only liquidus phase (Arndt, 1976; Bickle et al., 1977).

The CaO-MgO trend (Fig. 5-8) developed in the group II ultramafic komatiites intercepts the MgO axis at 40.66%. Assuming that orthopyroxene has 33% MgO and olivine 50% MgO then a mixture of olivine and orthopyroxene in the ratio 45 : 55 dissolving into an initial melt with ∿24% MgO could generate the CaO-MgO trend. As orthopyroxene in equilibrium with ultramafic liquids contains some CaO (e.g. Bickle et al., 1977) the proportion of orthopyroxene given above must be considered a minimum value. Assuming that the orthopyroxene contained 1.5% CaO, the MgO content at which the projected ultramafic trend and the orthopyroxene-olivine join, intercept can be found by solving the equations below:-

$$CaO_{(\text{ultramafic komatiite trend})} = -0.5759\ MgO + 23.417 \text{ (from Table 5-7)}$$

$$CaO_{(\text{orthopyroxene-olivine join})} = -0.0882\ MgO + 4.142$$

The value of 38.97% MgO obtained indicates that olivine and orthopyroxene entered the melt in the ratio 35 : 65. From this MgO value and the regression equations in Table 5-7 the composition of the olivine-orthopyroxene mixture entering the melt can be estimated (given in Table 8-1). Assuming all the TiO_2, Al_2O_3, CaO, Zr, Y, V, Ga and Sc in this composition were contributed by orthopyroxene, the contents of these

elements in the orthopyroxene (olivine : orthopyroxene = 35 : 65) are given in Table 8-1. The Al_2O_3 content (3.3%) is significantly higher than the orthopyroxene in equilibrium with komatiitic liquids at 40 kb. (Bickle et al., 1977).

Distribution coefficients calculated between the postulated orthopyroxene and two extreme liquid compositions (24 and 33% MgO) are also given in Table 8-1. The values for Ti, Zr, Y, V and Ga must vary from <1 to >1 as the MgO content of the melt increases from 24 to 33%. $D_{Ti}^{opx/liq}$ calculated from the data given in Bickle et al. (1977) is 0.09, significantly less than the required value. Le Roex (1980) has found $D_{Zr}^{opx/liq}$ in basaltic systems to be <0.01, much lower than the required value for this model. $D^{opx/liq}$ values for other elements (Y, V, Ga and Sc) are similar to those reported in the literature.

The relatively high $D^{opx/liq}$ values required for Ti and Zr suggest the olivine-orthopyroxene melting model is unreasonable for the generation of the Barberton ultramafic komatiites. Bickle et al. (1977) have rejected the harzburgite residue model, as their calculations show that 5% Al_2O_3 must occur in the residual orthopyroxene while the liquidus orthopyroxenes (up to 40 kbs.) in their experiments only contained 2% Al_2O_3. They also found that olivine and orthopyroxene did not occur together as liquidus phases at pressures up to 40 kbs. It therefore appears that at pressures of 40 kbs. or less the harzburgite melting model can be rejected. However, the question of whether ultramafic komatiites could have been generated by this mechanism at higher pressures and temperatures must remain open, as mineral compositions and distribution coefficients between melt and the residual minerals in such a model are unknown at present.

GARNET REMOVAL

The high CaO/Al_2O_3 ratios of the group II ultramafic komatiites from Barberton, particularly when compared to similar rocks from other greenstone belts, has led to speculation about their Al_2O_3 depleted nature. Models of garnet settling (McIver and Lenthal, 1973; Green, 1975; Nesbitt and Sun, 1976; Sun and Nesbitt, 1978; Nesbitt et al., 1979) have been proposed. Cawthorn and Strong (1975) have suggested shallow-level melting of a layered mantle in which the clinopyroxene/garnet ratio decreases with depth. As both group I and II ultramafic komatiites (Table 5-3b) have high CaO/Al_2O_3 ratios, if garnet loss processes were operative, then it must apply to both groups of lavas.

In considering possible garnet removal processes in the generation of Barberton ultramafic komatiites, only the more abundant group II data will be discussed. Variations of the CaO/Al_2O_3 ratio of the ultramafic komatiites (Table 5-3b) show that this ratio decreases regularly from 2.5 at 24% MgO to 1.2 at 33% MgO. This indicates that if garnet loss was operative the low MgO lavas lost proportionately more garnet than the high MgO lavas. If garnet loss was able to produce a two-fold variation in the CaO/Al_2O_3 ratio by selective removal of Al_2O_3, it would also be expected to produce equally large variations in the Al/Ti ratio. This however is not the case as all the group II ultramafic komatiites have Al/Ti ratios in the restricted range of 9.1 to 9.9 (Table 5-3b). Garnet loss, either by removal during melting or by later crystallisation, is therefore considered improbable for developing the high CaO/Al_2O_3 ratios in ultramafic komatiite magmas.

As the Ca/Ti ratio in the same samples (Table 5-3b) shows a range from 31 to 16 it is considered more likely that a process was operative that decreased the relative abundance of CaO with increasing MgO content

of the magmas. An analogous process of relatively decreasing clinopyroxene contribution with increasing MgO content could account for the CaO/Al_2O_3, Al/Ti and Ca/Ti trends. Cawthorn and Strong (1974) have suggested that the clinopyroxene/garnet ratio in the mantle decreased with depth. This would imply that the higher MgO ultramafic magmas originated from greater depths than the lower MgO magmas. Other processes can also be envisaged, such as sequential melting in an upwelling diapir and removal of the first melts before clinopyroxene had been exhausted in the source, later melts having lower CaO contents relative to Al_2O_3. However, none of these processes is favoured as they would be expected to produce a more scattered CaO-MgO trend than observed (Fig. 5-8) or to yield lavas with different inter-element ratios (such as Al/Ti or Ti/Zr) and REE patterns.

MIXING HYPOTHESIS

The intuitively reasonable models of olivine or olivine plus orthopyroxene control on the ultramafic magma compositions, do not account for the observed trends adequately. Cox (1978) has discussed possible processes that may be involved in the generation of high MgO liquids and has proposed mechanical mixing of melt and residual crystals to account for ultramafic magmas. He suggests that magmas initially form as crystal-rich mushes and become partly but unselectively separated from their refractory residue. The incorporated residual crystals dissolve as the magmas rise to the surface. The geochemical trends developed in the lavas are mixing lines between the composition of the initial melt and the residual crystals.

Bickle et al. (1977) have proposed a similar model for the origin of the Belingwe ultramafic komatiites. Their model differs from that

suggested by Cox (1978) in that they consider the initial melt incorporated overlying mantle material (garnet lherzolite) which was assimilated before extrusion. However, in either model the trends developed in the ultramafic komatiite magmas represent mixing lines between initial melt and incorporated solid. If the overlying mantle was of the same composition as that from which the initial melt was derived then the initial melt, range of magma compositions, initial mantle and residue must all have compositions that lie on the same mixing lines.

It has been argued in Chapter 5 that the trends developed in the group II ultramafic lavas (Fig. 5-8) are linear and although the scatter in some of the plots results in some uncertainty, the more coherent trends fall on reasonably straight trends (e.g. CaO-MgO). These trends may therefore represent mixing lines between initial melt and assimilated solid material. If this is so, several processes could have given rise to the initial melt composition such as single or multi-stage melting events. Similarly there are several possible compositions for the assimilated solid such as: residuum (Cox, 1978), garnet lherzolite (Bickle et al., 1977) or material of different composition to that which gave rise to the initial melt.

As the phase relationships, depths and temperatures at which the ultramafic komatiite magmas originated are unknown, the possibilities listed above will be examined to illustrate which are more probable. It should be noted that the following interpretation is applicable to dry systems only. In Fig. 8-4 the approximate fields occupied by group II ultramafic komatiites have been sketched onto the CS-MS-A (olivine projection) and C_2S_3-M_2S-A_2S_3 (enstatite projection) planes.

Assuming that the ultramafic magmas originated from a depth where garnet and clinopyroxene were stable, diagrammatic phase boundaries have

been indicated as well. Assuming further that mixing of initial melt, with its source mantle composition (assumed to be garnet lherzolite) or residuum, gave rise to the range of ultramafic magma compositions, this mantle composition must lie on the extension of the ultramafic komatiite trend (point M chosen to illustrate the argument). With the ultramafic komatiite compositional fields and phase boundaries indicated in Fig. 8-4, it would be expected that the mantle composition should lie on the clinopyroxene rich side of the join between the pseudo-invariant point to olivine (or orthopyroxene). Under these conditions garnet would be exhausted in the residue, before clinopyroxene and melts with high CaO/Al_2O_3 ratios could be developed. Initial melting of this mantle composition would result in liquid L and a residue somewhere along M-N.

Once the melting of garnet was complete (residue at point N) continued melting would result in liquids along line L-P (and residue along N-D). It should be noted that the precise location of point N and the residues along N-O will depend on the extent of solid solution in the pyroxenes. Although the extent of solid solution would be expected to decrease with increased degree of melting, for simplicity, only one set of pyroxene compositions has been considered.

Once clinopyroxene was exhausted in the residue (residue at O) further melting would produce liquids along line P-M. Mixing of any liquids along L-P-M with corresponding residues along M-N-O or initial mantle M, would give rise to lavas that plot in the shaded area. These possible lava compositions are quite distinct from the observed ultramafic komatiite fields and imply that:-

(a) phase relationships were significantly different from those given in Fig. 8-4 (approximately the 30 kb. phase relationships from O'Hara, 1968),

or (b) the initial melt was derived from a different mantle composition to that which was assimilated,

or (c) that single stage melting and mixing models are not applicable.

In considering possibility (a) above, if the pseudo-invariant point was significantly displaced towards the clinopyroxene rich side than that given in Fig. 8-4, it would be possible to produce ultramafic komatiite magmas in a single stage melting and mixing event. For this to be possible the pseudo-invariant point, range of ultramafic komatiite compositions and initial mantle must all lie on the same line. Under these conditions, melts produced at the invariant point would give rise to residues that plotted along the back projection of the line discussed above. Mixing of initial melt with overlying mantle or residual crystals (as suggested by Cox, 1978) would give rise to magmas that lay along the same line and therefore in principle could give rise to the range of ultramafic komatiite lavas observed. However, this model does imply phase boundaries significantly different from those given in O'Hara (1968) and until detailed high pressure phase relationships are at hand for group II ultramafic komatiite compositions, this option cannot be discounted.

The possibility (b) above, that the initial liquid was generated from a mantle composition different to that which was later assimilated, is considered unlikely. This is because different mantle compositions would also be expected to have different Al/Ti, Ti/Zr etc. ratios and a systematic change of these ratios with increasing MgO content of the lavas would be expected. No systematic changes of the Al/Ti ratio with increasing MgO content are observed for the group I or II ultramafic komatiites.

Assuming that the phase relationships are approximately correct and that only one mantle composition was involved in the production of each komatiite type, then a two stage melting-mixing model can be developed (Fig. 8-5) that can in principle account for the group II ultramafic komatiite compositions. Here mantle of composition M melts to produce liquid L and a residue somewhere along M-N. If at this point some of the liquid was removed, the bulk composition of the remaining liquid plus residue would lie along M-N (point Q chosen to illustrate the argument further). Additional melting of the residue would produce liquids at L (if garnet had not been exhausted in the residue) or along L-P, if the residue from the previous melting episode lay at point N. Once clinopyroxene had been exhausted in the source further melting would produce liquids along line P-Q.

If liquids close to point P along the melting path were extracted and assimilated variable amounts of overlying mantle (of composition M), lava compositions could be produced along P-M, i.e. along the observed ultramafic komatiite trend. In principle any compositions in the fields L-P-Q and Q-N-O could be produced by mixing the range of liquids with initial mantle M or corresponding residue.

This two stage melting-mixing model can explain many of the features of the group II Onverwacht lavas. Provided some of the liquid L (Fig. 8-5) was removed at the stage when garnet had just completely entered the melt, this liquid could be a precursor to the group II tholeiitic basalts. It would have similar REE patterns to the initial source material and the bulk of the other incompatible elements such as Ti and Zr as well as Al_2O_3 (as garnet was exhausted in the source), would be expected to be in this melt. Further melting would give rise to liquids along L-P which would have similar REE patterns to the

extracted magma L as $D_{REE}^{mineral/liq}$ for olivine, orthopyroxene, and clinopyroxene are all significantly less than 1 (e.g. see compilation of REE distribution coefficients in Frey et al., 1978). The melts along L-P (and at point P) would have similar Al/Ti and Ti/Zr ratios as the extracted melt L and the initial source M, but higher Ca/Ti ratios than either L or M. Mixing of melt P with mantle M would not alter the Al/Ti ratio but would decrease the Ca/Ti (and CaO/Al_2O_3) ratio with increasing MgO content of the mixture. These features are consistent with the observed element variations in the group II ultramafic lavas.

The success of this model hinges around whether melts close to point P could be extracted and mixed with overlying mantle as melts on either side of P (i.e. along L-P or P-Q) would tend to blur the mixing trends. Mysen and Kushiro (1977) have shown that melting curves (T^oC versus % melt) of the ultramafic nodule PHN-1611 show a sharp increase in temperature for increased degree of melting (at 20 and 35 kb) once clinopyroxene had been exhausted and olivine and orthopyroxene remained as residual phases. Once liquids had evolved to point P (Fig. 8-5) clinopyroxene would have been exhausted in the residue and from the melting curves given in Mysen and Kushiro (1977), there would have to be a marked increase in temperature (or decrease in pressure) before additional melting could form liquids along P-Q. Therefore liquids around point P could possibly exist over a relatively large range of temperatures (or pressure changes) without undergoing much change of composition, and consequently the better the chances of these liquids being extracted from the residuum.

In summary the mixing model that is preferred for the generation of the group II ultramafic komatiites, was a two stage melting and

mixing process. The first stage was melting of initial mantle until garnet was exhausted in the source. The composition of this melt is located at the invariant point relevant to the pressure of melting. At this point a portion of the liquid was removed (possibly being the precursor magma to the group II tholeiitic basalts) and a second stage of melting commenced until clinopyroxene had been exhausted in the residue. This melt was partly or completely extracted from the refractory residuum (olivine and orthopyroxene) and assimilated varying amounts of overlying mantle as it rose to the surface. Two assumptions have been incorporated into this model, the first being that the source composition was a garnet lherzolite at the depth and temperature of melting. Although not considered here, models can be constructed for garnet free systems. The second assumption is that high pressure phase boundaries given in O'Hara (1968) for dry systems are approximately correct (i.e. the pseudo-invariant point lies on the garnet rich side of the ultramafic komatiite fields given in Fig. 8-5). Similar models can be constructed for the group I ultramafic komatiites, but because of the limited data for this rock type from the Onverwacht Group, they are not considered here.

In Table 8-2 estimates have been made of the amount of material that must be assimilated by the initial melt in order to generate the observed range of ultramafic komatiite compositions. MgO contents of the initial melt (18 and 24% MgO) and the assimilated solid (38 and 42% MgO) have been assumed. Under these conditions the amount of material that must be assimilated varies between 0 and 75 wt.%. Bickle et al. (1977) have considered the energy budget for assimilation of garnet lherzolite (42% MgO) into a liquid (24% MgO). According to their calculations 45 wt.% garnet lherzolite incorporated into the

liquid at a depth of 100 kms. would have dissolved before the magma reached the surface, provided that the heat lost to the surroundings was no more than heating due to release of gravitational energy into the melt. Whether assimilation of up to 75 wt.% solid is possible will depend in part, on how hot the material was, the depth at which it was incorporated and how quickly it was decompressed. Combinations in Table 8-2 that suggest between 0 and 50 wt.% assimilation, would seem to be reasonable from the data given by Bickle et al. (1977).

MANTLE COMPOSITIONS

It has been noted earlier that if the mixing hypothesis is correct then the initial melt, range of magma compositions and assimilated solid must lie on the same mixing lines. These mixing lines are defined by the trends developed in the ultramafic magmas. In Fig. 8-6 the trends developed in group I and II ultramafic komatiites have been projected to higher MgO values. Assuming the mixing hypothesis to be correct, the composition of the assimilated solid must lie on these projections.

The MgO content of the assimilated solid for the group II ultramafic magmas must lie between 33% (highest MgO ultramafic komatiite lava that represents a liquid composition) and 40.67% MgO (CaO-MgO trend intercepts the MgO axis at this value) while the assimilated solid forming the group I ultramafic magmas could lie between 33% and ∿50% MgO. As the composition of this assimilated solid could be overlying mantle, some proposed mantle compositions or ranges of compositions have been plotted on these diagrams.

Maaløe and Aoki (1977) have computed regression equations for lherzolites from different areas using many hundreds of analyses in

some cases. The trends of their three categories of lherzolite are also shown on Fig. 8-6. In general the projections from the ultramafic komatiites trends pass through or close to the suggested mantle compositions. However, in detail most mantle estimates are too low in Fe to be suitable source material for the group II ultramafic komatiites, except for some garnet lherzolites, such as the well known sheared nodule PHN-1611 (Nixon and Boyd, 1973) from Thaba Putsoa, Lesotho. In Table 8-3 possible mantle compositions have been calculated for the group II ultramafic komatiites (using the regression equations given in Table 5-7) assuming MgO contents of 38 and 39% MgO. The major element composition of PHN-1611 and the postulated 38% MgO mantle (group II) agree well, except for higher CaO in the former. Meyer (1977) has suggested that sheared xenoliths are representative of primary undifferentiated mantle. Mantle material of similar major element composition to the sheared nodule PHN-1611 would be suitable source material for the group II ultramafic komatiites, at least for major elements.

The projected trends of the group I ultramafic komatiites are based on two analyses and, as noted previously, any conclusions regarding source compositions for these lavas, are poorly constrained. However, several suggested mantle compositions appear to be suitable source material for the group I ultramafic komatiites. The trends obtained by Maaløe and Aoki (1977) for continental spinel lherzolites (CSL) closely parallel the projected trends of the group I ultramafic komatiites. The FeO-MgO CSL trend intersects the group I ultramafic komatiite trend at 38% MgO and possible mantle compositions have been calculated for both CSL and group I ultramafic komatiites (Table 8-3) with MgO contents of 38%. These compositions agree well,

except for higher Al_2O_3 and lower Ni in the postulated CSL mantle.

More refractory nodule compositions plot close to the projected group I trends, such as the average Premier coarse garnet lherzolite (APCGL) obtained by Danchin (1979). In addition the regression trends for African garnet lherzolites (AGL) of Maaløe and Aoki (1977) intersect the projected group I ultramafic komatiite trends at higher MgO values. Another possible mantle composition for the group I ultramafic komatiites has been calculated with 42% MgO and compared with APCGL (Danchin, 1979) and 42% MgO AGL calculated from regression equations given in Maaløe and Aoki (1977). The three compositions all have very similar major element contents except for the higher Ni in the calculated source from the group I ultramafic komatiite data.

It therefore appears that if the ultramafic komatiites were generated by a mixing process as previously outlined, suitable source compositions, at least for the major elements, for both group I and II lavas, occur in the mantle. However, it should be noted that of the vast suite of mantle derived nodule compositions available in the literature, the majority are not suitable source material because of their depleted nature, particularly for crucial elements such as Ca, Na, Al and K. Source compositions for the group I ultramafic komatiites could either be relatively fertile (38% MgO) or depleted (42% MgO). If the group I ultramafic komatiites had been derived from depleted source compositions, they might be expected to show light REE depleted patterns which is not the case for the Onverwacht lavas. (See Chapter 7). On the other hand, komatiites such as those from Munro Township (Arth et al., 1977) do have light REE depleted patterns and therefore could possibly be derived from depleted source compositions as suggested by Arndt (1977c).

Before leaving this section it is of interest to compare the two mantle source compositions proposed for the ultramafic komatiites from the Onverwacht. In Table 8-3 possible source compositions for the group I and II lavas have been calculated at 38% MgO. The differences in major element abundances are small except for lower (roughly half) CaO and Cr_2O_3 contents in the group II source. In the trace elements the main differences are higher Zr, Y and Co in the group II source. In general, however, the compositional differences are small and would require careful analysis of potentially suitable mantle-derived material if the two postulated source compositions are to be identified.

8-3 MAFIC KOMATIITES

In Chapter 6 the origin of mafic komatiites has been considered in detail and variations in composition for the group II mafic komatiites were accounted for by the fractionation of olivine, followed by olivine plus clinopyroxene from an ultramafic parent liquid. Partial melting processes, however, have not been discounted as a possible alternative mechanism for obtaining the range of mafic komatiite compositions. In the discussion below, processes that may have generated the composition of the group II mafic komatiites, are considered with reference to projections from mineral compositions in CMAS space (O'Hara, 1968).

The projection from enstatite onto the M_2S-A_2S_3-C_2S_3 plane (Fig. 8-8) and from olivine onto the MS-A-CS plane (Fig. 8-7) show that the high-Mg mafic komatiites could be derived from the range of ultramafic komatiites by fractionation of olivine. Although from these diagrams fractionation of minor amounts of orthopyroxene cannot

be discounted, major and trace element modelling (Chapter 6) of the high-Mg mafic komatiites indicates that it was not an essential fractionating phase from these lavas. The low-Mg mafic komatiites plot around or close to the one atmosphere phase boundaries between orthopyroxene-clinopyroxene (Fig. 8-7) and olivine-clinopyroxene (Fig. 8-8). This indicates a low pressure origin for the low-Mg mafic komatiites. Fractionation of olivine followed by olivine plus clinopyroxene to obtain the low-Mg mafic komatiites from an ultramafic parent liquid, is consistent with the major and trace element data (see Chapter 6) and the projection from orthopyroxene (Fig. 8-8). The projection from olivine (Fig. 8-7) however indicates that some orthopyroxene was also involved although no orthopyroxene has been observed in the low-Mg mafic komatiite lavas. However, small amounts of orthopyroxene cannot be discounted by the geochemical data of the low-Mg mafic komatiites due to the scatter in the plots (see Chapter 6).

Provided the phase boundaries delineated at different pressures by O'Hara (1968) are valid for komatiite compositions, partial melting processes for the origin of mafic komatiites up to 30 kbs. are considered unlikely. It is therefore concluded that the composition of the low-Mg mafic komatiites was controlled by low pressure olivine and clinopyroxene (small amounts of orthopyroxene fractionation cannot be discounted) fractionation, ultimately from an ultramafic magma. It is further suggested that olivine fractionation from an ultramafic magma at low pressures generated the high-Mg mafic komatiites as an intermediate step in the above fractionation process. Polybaric fractionation of olivine plus orthopyroxene does not appear to have been a major process in the development of the mafic komatiite

compositions, as no mafic komatiites plot on the 20 to 30 kb. side of the one atmosphere phase boundaries in Fig. 8-7.

High-Mg mafic komatiites that have been assigned to the group I rock type (Chapter 7) show no obvious crystal fractionation relationships to the group I ultramafic komatiites (see Fig. 6-5). Lower MgO komatiite lavas belonging to group I appear to be lacking except for one porphyritic low-Mg mafic komatiite HSS-6 (Table 6-5a and 6-5b). The fractionation of olivine plus clinopyroxene has been suggested (Chapter 6) as a possible mechanism for obtaining the range of compositions of the group I high-Mg mafic komatiites. However, the projection from diopside into the C_3A-M-S plane suggests that orthopyroxene (plus clinopyroxene) may have played a significant role in the derivation of these lavas. Further speculation on the origin of the group I high-Mg mafic komatiites is not warranted at present and the development of more detailed models for their petrogenesis must await the acquisition of additional data.

8-4 THOLEIITIC BASALTS

In Chapter 7 it has been argued on the basis of certain inter-element ratios, that the high-Ti tholeiites and the group II ultramafic komatiites have been derived from the same source composition. Similar conclusions were reached for the low-Ti tholeiites and the group I ultramafic komatiites. Assuming this to be correct the mantle compositions calculated for the group I and II ultramafic komatiites should be the same as that for the relevant tholeiite type. If the group I and II tholeiites represent unmodified partial melts of their source material, then F (the degree of melting) can be estimated from the elements that are considered incompatible

($C_o/C_1 \sim F$, Arth, 1976) in likely residual phases. In Table 8-4 estimates of F have been calculated from the Ti, Al, Zr, Y and Ga data in group I and II tholeiites and their respective source compositions (assumed to be the 38% MgO compositions given in Table 8-3). The degree of melting is consistent for all the elements and indicates that the tholeiites represent 17 - 18% partial melts of their source material.

Basalts that have been derived directly from the mantle where they were in equilibrium with residual olivine should have Mg numbers (atomic 100 $Mg/(Mg+Fe^{2+})$ in the range 68 -72 (e.g. Duncan and Erlank, 1979). The Mg numbers calculated for the group I tholeiites (average = 64) and group II thoeliites (average = 51) indicate that these lavas are not primary melt compositions but have been compositionally modified before extrusion. The degrees of melting calculated above must therefore be considered minimum values. The projections from olivine (Fig. 8-7), orthopyroxene (Fig. 8-8) and clinopyroxene (Fig. 8-9) indicate that fractionation of olivine, orthopyroxene and clinopyroxene may have modified the composition of the primary tholeiitic magmas before they were extruded.

In Chapter 7 it was suggested that the lower Ca/Ti ratios found in the group II tholeiites compared to the ultramafic komatiites were due to clinopyroxene remaining as a residual phase after the melting event that generated the tholeiitic magmas. The low Ca/Ti ratios in the tholeiitic lavas could also be due to fractionation of clinopyroxene from the primary magma as it rose to the surface.

The crystallisation and settling of olivine, orthopyroxene and clinopyroxene from primary tholeiitic magmas would have little

effect on inter-element ratios such as Al/Ti or Ti/Zr as these elements are preferentially partitioned into the liquid. Therefore, although it appears likely that the tholeiitic lavas represent modified mantle-derived melts, the similarity of inter-element ratios between group I ultramafic komatiites and tholeiites and between group II ultramafic komatiites and tholeiites, nevertheless indicates that the tholeiites were derived from similar source compositions as their respective type of ultramafic komatiite. It was suggested earlier in this Chapter that the first melt extracted in the two stage melting-mixing model, could possibly represent the precursor magma of the tholeiitic basalts. Although speculative, this model would account for the similarity of the inter-element relationships between the ultramafic komatiites and the relevant tholeiitic basalt types.

8-5 CONCLUSIONS

In considering possible processes of petrogenesis of the komatiitic lavas from the Onverwacht, many problems have been pinpointed and clearly further work on the freshest available material is required. However, some of the models investigated could possibly account for the range of magma compositions and these are summarised below, although some mechanisms are considered more likely than others.

ULTRAMAFIC KOMATIITES

(a) Olivine control developed, either by melting processes (into lower MgO liquids) or subsequent crystallisation from the magmas (from higher MgO liquids), is compatible with range of group I ultramafic komatiite lava compositions, but not the group II ultramafic komatiites. It is

however considered probable that both group I and II ultramafic komatiite magmas were generated by the same process but that different source compositions were involved.

(b) Olivine plus orthopyroxene fractionation is considered an unlikely mechanism for generating the group II ultramafic komatiite magmas at pressures <40 kb. This possibility cannot be discounted for higher pressures (>40 kb.), as information on mineral compositions or distribution coefficients are not yet available.

(c) Mixing processes are considered the most probable method for generating the ultramafic komatiite magmas. The preferred mechanism involves an initial stage of melting (until garnet was exhausted in the residue) and extraction of a portion of this first melt, followed by a second stage of melting (until clinopyroxene was exhausted in the residue). Some of this second melt was extracted and assimilated variable amounts of overlying mantle, giving rise to a range of ultramafic lava compositions. A model, involving a single stage of melting and mixing this melt, either with residuum (Cox, 1978) or overlying mantle (Bickle et al., 1977), cannot be discounted if phase boundaries for ultramafic komatiites are significantly different to those given by O'Hara (1968).

MAFIC KOMATIITES

(a) The group II high-Mg mafic komatiites and basic low-Mg

mafic komatiites were derived by olivine fractionation from an ultramafic magma.

(b) Further olivine plus clinopyroxene fractionation generated the evolved low-Mg mafic komatiites. Projections in CMAS space indicate that minor amounts of orthopyroxene were also involved and that the low-Mg mafic komatiites were derived at low pressures.

(c) The origin of the group I high-Mg mafic komatiites remains a problem and additional data on this group of rocks are required before reasonable petrogenetic models can be developed.

THOLEIITIC BASALTS

(a) The group II tholeiitic basalts were derived from source material similar in composition to that from which the group II ultramafic komatiites were derived. Olivine and pyroxene fractionated from these primary tholeiitic magmas before the lavas were extruded.

(b) Group I tholeiitic basalts were derived from similar source material to the group I ultramafic komatiites. These primary tholeiitic magmas however may have crystallised olivine before being extruded.

MANTLE COMPOSITIONS

(a) Suitable source material can be identified from available mantle-derived nodule compositions for

both group I and II lavas, although it is noted that suitable compositions are rare.

The reasons why volumes of mantle material undergo partial melting to produce magmas is an intriguing problem. It has been suggested (e.g. Green, 1975; Nesbitt and Sun, 1976; Arndt, 1977c; Bickle et al., 1977; Weaver and Tarney, 1979) that mantle diapirs from which ultramafic komatiites were derived rose from depths as great as 300 - 400 kms. The depth at which melting began and the depths at which separation of magma from residuum took place could conceivably be at much greater pressures than the currently available experimental data. The models outlined above for the generation of the Barberton greenstone belt lavas must be considered as tentative, until phase relationships, mineral compositions, melting relationships and distribution coefficients have been studied at much higher pressures.

FINAL COMMENT

Slightly more than a decade has elapsed since Richard and Morris Viljoen published their exciting results of work carried out on the Barberton greenstone belt. Their work has proved to be a milestone (greenschist facies!)in the development of Archaean geology and geochemistry. The vast quantities of research stimulated by the discovery of komatiites is evidenced by the rapidly growing piles of reprints and preprints, dealing with all aspects of the existence of komatiites, accumulating on the desks of Archaean geologists and geochemists. Clearly research into the Barberton greenstone belt, the birthplace of komatiites, should be kept abreast of advancements

being made in the knowledge of greenstone belt evolution in other countries. Of course, additional data are always required to further develop and test models proposed for the generation and evolution of lavas and in this respect Barberton is no exception. However, in the case of the ultramafic lavas from the LUU there is some urgency attached to clearly establishing the effects of alteration and metamorphism on the geochemical trends in these lavas. REE data on a suite of selected lavas, of all types, from the LUU would also be a valuable addition to the geochemical data set. The LUU only forms half of the Onverwacht volcanic pile and detailed investigations into the geochemistry of the Upper Mafic to Felsic Unit are also required. Apart from the volcanic rocks, the Barberton greenstone belt consists of sedimentary sequences and surrounding granitic intrusions. Numerous aspects of this interesting terrain are currently being investigated or still remain to be researched in the years to come.

ACKNOWLEDGEMENTS

My sincere appreciation is expressed for the efforts of the many people who have, in one way or another, provided me with invaluable support during the course of this work. Professors A.J. Erlank and L.H. Ahrens are thanked for providing the opportunity and making available the facilities of the Geochemistry Department, not only to carry out work directly related to the project, but also to gain experience in related fields.

The unfailing support and enthusiasm, expressed on many different levels, of my supervisor, Professor A.J. Erlank, was of immeasurable assistance in initiating, carrying out and concluding this project. His insight and sure guidance helped me over the difficult periods when my enthusiasm was either too low or too high. The efforts of Professor K.G. Cox were also greatly appreciated, as he spent many hours battling with my handwriting, English and science, and provided me with graphic explanations of the workings of CMAS space. His valuable criticisms and suggestions have significantly improved the quality of this thesis. All other shortcomings in the work are my responsibility.

The vast experiences of Dr. J.P. Willis and Dr. A.R. Duncan in the field of X-ray fluorescence have greatly improved the quality of the analytical data reported in this work. In particular, the many hours of instruction I have received from them both in the theory and practice of XRF, form a valuable part of my education. Dr. J.P. Willis also provided some of the Ga data disclosed in this thesis. In the related field of programming computers to carry out the tedious work of reduction and manipulation of data, Dr. A.R. Duncan has provided me with an equally valuable education. His lucid explanations of high and low level computer languages have enabled me to develop programs that, at least, were intelligible to the computers for

which they were written! Dr. Duncan is sincerely thanked for the many hours of formal and informal instruction in these matters. The help obtained from Mr. R.S. Rickard in performing the electron microprobe analyses and the associated computer reduction of the data was also greatly appreciated.

Many of the ideas incorporated in this thesis were born during discussions with my colleagues including the aforementioned people as well as Dr. D.L. Reid, Mr. A.P. le Roex, Mr. A.B. Simpson, Dr. J.J. Gurney, Dr. B.W. Flemming, Dr. A.E. Moore, Dr. A.C. Moore, Dr. H.L. Allsopp, Dr. C.R. Anhaeusser and Dr. C.J. Hawkesworth. In addition I have had numerous interesting 'komatiitic' discussions with earth scientists who have visited the University of Cape Town, from all parts of the world.

The financial contributions received from the University of Cape Town and the CSIR throughout the course of this work are greatfully acknowledged. Dr. J.J. Gurney and Mr. S. Moir are also thanked for providing opportunities to apply much of what I have learned to industrial problems, and for the financial renumerations, without which it would have been impossible to continue with these studies. In 1978 I was able to attend the 'Archaean Geochemistry Field Conference' held in Ontario, Canada and the U.S.A. The generous financial contributions, which made this valuable experience possible, from the Bremner Travel Fund, the Geochemistry Department and from Professor A.M. Goodwin are greatfully acknowledged.

Mrs. G. Verblun, Miss L. Yeats and Mrs. E.H. Tydeman kindly assisted in the typing of early drafts of the manuscript. Mrs. G. Verblun is thanked for her conscientious efforts and patience in typing the final manuscript. Miss L. Yeats tackled the arduous task of proof-reading with enthusiasm and skill. Miss L. Willows-Munroe and Miss L. Yeats assisted with the draughting and the varied assistance obtained from Mrs. M. Brauteseth, Henri Fortuin and other members of the geochemistry and geology staff, is greatly appreciated.

Mrs. A. Smith is thanked for her loyal support under extremely difficult conditions, in the earlier years of this work. Finally a warm thanks is extended to Lynette, Mikki, Wendy, Margie, Andy and Julian for their friendship and unique 'Monte Vista' type distractions which helped me over some trying times.

REFERENCES

Abbey, S., 1975. Studies in 'Standard Samples' of silicate rocks and minerals. Part 4 : 1974 edition of 'usable' values. Geol. Surv. Paper, 74-41 : 1-23.

Allsopp, H.L., Davies, R.D., de Gasparis, A.A.A. and Nicolaysen, L.O., 1969. Review of Rb-Sr age measurements from the early Precambrian terrain in the south-eastern Transvaal and Swaziland. Geol. Soc. S. Africa, Spec. Publ., 2 : 433-444.

Allsopp, H.L., Roberts, H.R., Schreiner, G.D.L. and Hunter, D.R., 1962. Rb-Sr age measurements on various Swaziland granites. J. Geophys. Res., 67 : 5307-5313.

Allsopp, H.L., Ulrych, T.J. and Nicolaysen, L.O., 1968. Dating some significant events in the history of the Swaziland System by the Rb-Sr isochron method. Can. J. Earth Sci., 5 : 605-619.

Allsopp, H.L., Viljoen, M.J. and Viljoen, R.P., 1973. Strontium isotopic studies of the mafic and felsic rocks of the Onverwacht Group of the Swaziland Sequence. Geol. Rundsch., 62 : 902-917.

Anhaeusser, C.R., 1971a. Cyclic volcanicity and sedimentation in the evolutionary development of Archaean greenstone belts of shield areas. Geol. Soc. Aust., Spec. Publ., 3 : 57-70.

Anhaeusser, C.R., 1971b. The Barberton Mountain Land, South Africa - A guide to the understanding of the Archaean geology of Western Australia. Gecl. Soc. Aust., Spec. Publ., 3 : 103-120.

Anhaeusser, C.R., 1972. The geology of the Jamestown Hills area of the Barberton Mountain Land, South Africa. Trans. Geol. Soc. S. Africa, 75 : 225-263.

Anhaeusser, C.R., 1973. The evolution of the early Precambrian crust of southern Africa. Phil. Trans. R. Soc. Lond. A., 273 : 359-388.

Anhaeusser, C.R., 1975. The geological evolution of the primitive earth: evidence from the Barberton Mountain Land. Econ. Geol. Res. Unit, Inf. Circ., 98 : 1-22.

Anhaeusser, C.R., 1976a. Archean metallogeny in southern Africa. Econ. Geol., 71 : 16-43.

Anhaeusser, C.R., 1976b. A bibliography of the geology relating to the Barberton Mountain Land and surrounding granitic terrane. Econ. Geol. Res. Unit, Inf. Circ., 102 : 1-51.

Anhaeusser, C.R., 1976c. The geology of the Sheba Hills area of the Barberton Mountain Land, South Africa, with particular reference to the Eureka syncline. Trans. Geol. Soc. S. Africa, 79 : 253-280.

Anhaeusser, C.R., 1976d. The nature of chrysotile asbestos occurrences in southern Africa : A review. Econ. Geol., 71 : 96-116.

Anhaeusser, C.R., 1977. Geological and geochemical investigations of the Roodekrans ultramafic complex and surrounding Archaean volcanic rocks, Krugersdorp District. Trans. Geol. Soc. S. Africa, 80 : 17-28.

Anhaeusser, C.R., Mason, R., Viljoen, M.J. and Viljoen, R.P., 1969. A reappraisal of some aspects of Precambrian shield geology. Bull. Geol. Soc. Amer., 80 : 2175-2200.

Anhaeusser, C.R. and Robb, L.R., 1978. Regional and detailed field and geochemical studies of Archaean tronohjemitic gneisses, migmatites and greenstone xenoliths in the southern part of the Barberton Mountain Land, South Africa. Econ. Geol. Res. Unit. Inf. Circ., 125 : 1-14.

Anhaeusser, C.R., Roering, C., Viljoen, M.J. and Viljoen, R.P., 1968. The Barberton Mountain Land. A model of the elements and evolution of an Archaean fold belt. Geol. Soc. S. Africa, 71 (annex.) : 225-253.

Arndt, N.T., 1975. Ultramafic rocks of Munro Township and their volcanic setting. Ph.D. thesis, University of Toronto.

Arndt, N.T., 1976. Melting relations of ultramafic lavas (komatiites) at 1 atm. and high pressure. Carnegie Inst. Wash. Yearbook, 75 : 555-562.

Arndt, N.T., 1977a. Thick, layered peridotite-gabbro lava flows in Munro Township, Ontario. Can. J. Earth Sci., 14 : 2620-2637.

Arndt, N.T., 1977b. Partitioning of nickel between olivine and ultrabasic and basic komatiite liquids. Carnegie Inst. Wash. Yearbook, 76 : 553-557.

Arndt, N.T., 1977c. Ultrabasic magmas and high-degree melting of mantle. Contrib. Mineral. Petrol., 64 : 205-221.

Arndt, N.T., 1980. What is a komatiite? (in press).

Arndt, N.T., Francis, D. and Hynes, A.J., 1979. The field characteristics and petrology of Archean and proterozoic komatiites. Can. Mineral., 17 : 147-163.

Arndt, N.T., Naldrett, A.J. and Pyke, D.R., 1977. Komatiitic and iron-rich tholeiitic lavas of Munro Township, northeast Ontario. J. Petrol., 18 : 319-369.

Arriens, P.A., 1971. The Archaean geochronology of Australia. Geol. Soc. Aust., Spec. Publ., 3 : 11-23.

Arth, J.G., 1976. Behavior of trace elements during magmatic processes - A summary of theoretical models and their applications. J. Res. U.S. Geol. Surv., 4 : 41-47.

Arth, J.G., Arndt, N.T. and Naldrett, A.J., 1977. Genesis of Archean komatiites from Munro Township, Ontario : Trace-element evidence. Geology, 5 : 590-594.

Atkins, F.B., 1969. Pyroxenes of the Bushveld Intrusion, South Africa. J. Petrol., 10 : 222-249.

Aumento, F., Mitchell, W.S. and Fratta, M., 1976. Interaction between sea water and oceanic layer two as a function of time and depth - 1. Field evidence. Can. Mineral., 14 : 269-290.

Baird, A.K., 1961. A pressed specimen die for the Norelco vacuum-path X-ray spectrograph. Norelco Rep. : 8, 108-109.

Baragar, W.R.A., 1966. Geochemistry of the Yellowknife volcanic rocks. Can. J. Earth Sci., 3 : 9-30.

Baragar, W.R.A., 1968. Major-element geochemistry of the Noranda volcanic belt, Quebec-Ontario. Can. J. Earth Sci., 5 : 773-790.

Baragar, W.R.A. and Goodwin, A.M., 1969. Andesite and Archean volcanism of the Canadian shield. Proc. Andesite Conf. Bull., 65 : 121-142.

Baragar, W.R.A. and McGlynn, J.C., 1976. Early Archean basement in the Canadian shield : A review of the evidence. Geol. Surv. Canada Paper, 76-14 : 1-20.

Barghoorn, E.S. and Schopf, J.W., 1966. Microorganisms three billion years old from the Precambrian of South Africa. Science, 152 : 758-763.

Barton, J.M., 1975. Rb-Sr isotopic characteristics of the 3.6 b.y. Hebron Gneiss, Labrador. Earth Planet. Sci. Lett., 27 : 427-435.

Barton, J.M. (Jr), Fripp, R.E.P. and Ryan, B., 1977. Rb/Sr ages and geological setting of ancient dykes in the Sand River area, Limpopo Mobile Belt, South Africa. Nature, 267 : 487-490.

Barton, J.M. (Jr) and Ryan, B., 1977. A review of the geochronologic framework of the Limpopo Mobile Belt. Botswana Geol. Surv. Bull., 12 : 183-200.

Basu, A.R. and Tatsumoto, M., 1979. Kimberlites and mantle evolution : Nd-isotopic evidence (in press).

Bhattacharji, S., 1967. Scale model experiments on flowage differentiation in sills. In: P.J. Wyllie (Ed), Ultramafic and related rocks. Wiley and Sons, Inc., New York : 69-70.

Bhattacharji, S. and Smith, C.H., 1964. Flowage differentiation. Science, 145 : 150-153.

Bickle, M.J., Ford, C.E. and Nisbet, E.G., 1977. The petrogenesis of peridotitic komatiites : evidence from high-pressure melting experiments. Earth Planet. Sci. Lett., 37 : 97-106.

Bickle, M.J., Hawkesworth, C.J., Martin, A., Nisbet, E.G. and O'Nions, R.K.. 1976. Mantle composition derived from the chemistry of ultramafic lavas. Nature, 263 : 577-580.

Bickle, M.J., Martin, A. and Nisbet, E.G., 1975. Basaltic and peridotitic komatiites and stromatolites above a basal unconformity in the Belingwe greenstone belt, Rhodesia. Earth Planet. Sci. Lett., 27 : 155-162.

Birch, F., 1952. Elasticity and constitution of the Earth's interior. J. Geophys. Res., 51 : 227-286.

Bliss, N.W. and Stidolph, P.A., 1969. The Rhodesian basement complex : A review. Geol. Soc. S. Africa, Spec. Publ., 3 : 304-331.

Bottinga, Y. and Weill, D.F., 1970. Densitites of liquid silicate systems calculated from partial molar volumes of oxide components. Amer. J. Science, 269 : 169-182.

Bottinga, Y. and Weill, D.F., 1972. The viscosity of magmatic silicate liquids : A model for calculation. Amer. J. Science, 272 : 438-475.

Boyd, F.R. and Schairer, J.F., 1964. The system $MgSiO_3$-$CaMgSi_2O6$. J. Petrol., 5 : 545-560.

Brooks, C. and Hart, S.R., 1974. On the significance of komatiite. Geology, 2 : 107-110.

Bryan, W., Finger, L. and Chayes, F., 1969. Estimating proportions in petrogenic mixing equations by least squares approximation. Science, 163 : 926-927.

Burke, K. and Kidd, W.S.F., 1978. Were Archean continental geothermal gradients much steeper than those of today? Nature, 272 : 240-241.

Cameron, W.E., Nisbet, E.G. and Dietrich, V.J., 1979. Boninites, komatiites and ophiolitic basalts. Nature, 280 : 550-553.

Cann, J.R., 1969. Spilites from the Carlsberg Ridge, Indian Ocean. J. Petrol., 10 : 1-19.

Cann, J.R., 1970. Rb, Sr, Y, Zr and Nb in some ocean floor basaltic rocks. Earth Planet. Sci. Lett., 10 : 7-11.

Cawthorn, R.G., 1975. Degrees of melting in mantle diapirs and the origin of ultrabasic liquids. Earth Planet. Sci. Lett., 23 : 113-120.

Cawthorn, R.G. and McIver, J.R., 1977. Nickel in komatiites. Nature, 266 : 716-718.

Cawthorn, R.G. and Strong, D.F., 1975. The petrogenesis of komatiites and related rocks as evidence for a layered upper mantle. Earth Planet. Sci. Lett., 27 : 369-375.

Clark, S.P., 1966. Handbook of physical constants. Geol. Soc. Amer., Memoir, 97 : 86.

Clarke, D.B., 1970. Tertiary basalts of Baffin Bay : possible primary magma from the mantle. Contrib. Mineral. Petrol., 25 : 203-224.

Coish, R.A., 1977. Ocean floor metamorphism in the Betts Cove Ophiolite, Newfoundland. Contrib. Mineral. Petrol., 60 : 255-270.

Condie, K.C. and Harrison, N.M., 1976. Geochemistry of Archean Bulawayan group, Midlands greenstone belt, Rhodesia. Precam. Res., 3 : 253-271.

Condie, K.C., Macke, J.E. and Reimer, T.O., 1970. Petrology and geochemistry of early Precambrian graywackes from the Fig Tree Group, South Africa. Bull. Geol. Soc. Amer., 81 : 2759-2776.

Condie, K.C. and Madison, J.A., 1969. Compositional and volume changes accompanying progressive serpentinisation of dunites from the Webster-Addie ultramafic body, N. Carolina. Amer. Mineral., 54 : 1173-1179.

Condie, K.C., Viljoen, M.J. and Kable, E.J.D., 1977. Effects of alteration on element distributions in Archean tholeiites from the Barberton greenstone belt, South Africa. Contrib. Mineral. Petrol., 64 : 75-89.

Cox, K.G., 1978. Komatiites and other high-magnesia lavas : Some problems. Phil. Trans. R. Soc. Lond. A, 288 : 599-609.

Cox, K.G. and Bell, J.D., 1972. A crystal fractionation model for the basaltic rocks of the New Georgia group, British Solomon Islands. Contrib. Mineral. Petrol., 37 : 1-13.

Cox, K.G., Bell, J.D. and Pankhurst, R.J., 1979. The interpretation of igneous rocks. George Allen and Unwin, London : 450 pp.

Dale, I.M. and Henderson, P., 1972. The partition of transition elements in phenocryst-bearing basalts and the implications about melt structure. 24th Internat. Geol. Cong. Sec., 10 : 105-111.

Danchin, R.V., 1967. Chromium and nickel in the Fig Tree shale from South Africa. Science, 158 : 261-262.

Danchin, R.V., 1979. Mineral and bulk chemistry of garnet lherzolite and garnet harzburgite xenoliths from the Premier Mine, South Africa (in press).

Davies, R.D. and Allsopp, H.L., 1976. Strontium isotopic evidence relating to the evolution of the Lower Precambrian granitic crust in Swaziland. Geology, 4 : 553-556.

Davis, B.T.C. and Boyd, F.R., 1966. The join $Mg_2Si_2O_6$-$CaMgSi_2O_6$ at 30 kb. pressure and its application to pyroxenes from kimberlite. J. Geophys. Res., 71 : 3567-3576.

Deer, W.A., Howie, R.A. and Zussman, J., 1962. Rock-forming minerals. Vol. 1, Ortho- and ring silicates. Longmans, London : 333 pp.

De Gasparis, A.A.A., 1967. Rb-Sr isotopic studies relating to problems of geochronology on the Nelspruit and Mpageni granites. M.Sc. thesis, Univ. Witwatersrand, Johannesburg.

De Paolo, D.J. and Wasserburg, G.J., 1976a. Nd isotopic variations and petrogenetic models. Geophys. Res. Lett., 3 : 249-252.

De Paolo, D.J. and Wasserburg, G.J., 1976b. Inferences about magma sources and mantle structure from variations of $^{143}Nd/^{144}Nd$. Geophys. Res. Lett., 3 : 743-746.

Dèvigne, J.P., 1975. Fossil bacteria populations in an Archaean rock of the Swaziland System (Transvaal, South Africa). Ann. Rep. Res. Hist. Afr. Geol. Univ. Leeds, 19 : 73-74.

Dickey, J.S. (Jnr), 1972. A primary peridotite magma-revisited : Olivine quench crystals in a peridotite lava. Geol. Soc. Amer., Memoir, 132 : 280-297.

Dietrich, V., Emmermann, R., Oberhansli, R. and Puchelt, H., 1978. Geochemistry of basaltic and gabbroic rocks from the west Mariana basin and the Mariana trench. Earth Planet. Sci. Lett., 39 : 127-144.

Donaldson, C.H., 1974. Olivine crystal types in harrisitic rocks of the Rhum pluton and in Archean spinifex rocks. Bull. Geol. Soc. Amer., 85 : 1721-1726.

Donaldson, C.H., 1976. An experimental investigation of olivine morphology. Contrib. Mineral. Petrol., 57 : 187-213.

Donaldson, C.H., 1977. Laboratory duplication of comb layering in the Rhum pluton. Min. Mag., 41 : 323-336.

Drever, H.I. and Johnston, R., 1957. Crystal growth of forsteritic olivine in magmas and melts. Trans. Roy. Soc. Edin., 63 : 289-317.

Drever, H.I. and Johnston, R., 1966. A natural high-lime silicate liquid more basic than basalt. J. Petrol., 7 : 414-420.

Drever, H.I. and Johnston, R., 1972. Metastable growth patterns in some terrestrial and lunar rocks. Meteoritics, 7 : 327-340.

Drever, H.I., Johnston, R. and Thomas, C.M., 1961. Ultrabasic liquids. Nature, 192 : 157-158.

Duke, J.M., 1976. Distribution of the period four transition elements among olivine, calcic clinopyroxene and mafic silicate liquids : experimental results. J. Petrol., 17 : 499-521.

Duke, J.M. and Naldrett, A.J., 1978. A numerical model of the fractionation of olivine and molten sulfide from komatiite magma. Earth Planet. Sci. Lett., 39 : 255-266.

Duncan, A.R. and Erlank, A.J., 1979. Regional geochemistry of the Karoo volcanics. Abstracts Vol. Geokongres 79, Port Elizabeth, South Africa.

Durney, D.W., 1972. A major unconformity in the Archaean, Jones Creek, Western Australia. J. Geol. Soc. Aust., 19 : 251-259.

Engel, A.E.J., Engel, C.G. and Havens, R.G., 1965. Chemical characteristics of oceanic basalts and the upper mantle. Bull. Geol. Soc. Amer., 76 : 719-734.

Eriksson, K.A., 1978. Alluvial and destructive beach facies in the Archaean Moodies Group of the Barberton Mountain Land. Econ. Geol. Res. Unit. Inf. Circ., 115 : 1-18.

Erlank, A.J., Allsopp, H.L., Duncan, A.R. and Bristow, J.W., 1980. Mantle heterogeneity beneath southern Africa : Evidence from the volcanic record. Phil. Trans. Roy. Soc. Lond. A., (in press).

Erlank, A.J. and Kable, E.J.D., 1976. The significance of incompatible elements in Mid-Atlantic ridge basalts from 45°N with particular reference to Zr/Nb. Contrib. Mineral. Petrol., 54 : 281-291.

Erlank, A.J. and Reid, D.L., 1974. Geochemistry, mineralogy, and petrology of basalts, leg 25, Deep sea drilling project. In: Simpson et al. 'Initial Reports of the Deep Sea Drilling Project, 25 : 543-551.

Evensen, N.M., Hamilton, P.J. and O'Nions, R.K., 1978. Rare-earth abundances in chondritic meteorites. Geochim. Cosmochim. Acta, 42 : 1199-1212.

Ferguson, J. and Currie, K.L., 1972. Silicate immiscibility in the ancient "basalts" of the Barberton Mountain Land, Transvaal. Nature, 235 : 86-89.

Fleet, M.E. and MacRae, N.D., 1975. A spinifex rock from Munro Township, Ontario. Can. J. Earth Sci., 12 : 928-939.

Flemming, B.W., 1978. Some effects of particle shape on settling velocity, sediment texture, and depositional processes. Marine Geol. Programme, Technical Report No. 9, Dept. of Geology, Univ. Cape Town : 94-107.

Floyd, P.A. and Winchester, J.A., 1975. Magma type and tectonic setting discrimination using immobile elements. Earth Planet. Sci. Lett., 27 : 211-218.

French, W.J., 1971. The correlation between 'anhydrous' crystallisation temperatures and rock composition. Contr. Mineral. Petrol., 31 : 154-158.

Frey, F.A., Green, D.H. and Roy, S.D., 1978. Integrated models of basalt petrogenesis : a study of quartz tholeiites to olivine melilitites from south eastern Australia utilizing geochemical and experimental petrological data. J. Petrol., 19 : 463-513.

Gass, I.G., 1958. Ultrabasic pillow lavas from Cyprus. Geol. Mag., 95 : 241-251.

Gast, P.W., 1968. Trace element fractionation and the origin of tholeiitic and alkaline magma types. Geochim. Cosmochim. Acta, 32 : 1057-1086.

Gélinas, L., Brooks, C. and Trzcienski, Jr., W.E., 1976. Archean variolites - quenced immiscible liquids. Can. J. Earth Sci., 13 : 210-230.

Gill, J.B., 1976. Composition and age of Lau Basin and Ridge volcanic rocks. Implications for evolution of an interarc basin and remnant arc. Bull. Geol. Soc. Amer., 87 : 1384-1395.

Gill, R.C.O., 1977. Comparative petrogenesis of Archaean and modern low-K tholeiites. A critical review of some geochemical aspects. In: L.H. Ahrens (Ed.). Physics and Chemistry of the Earth, 11 : 431-447. Pergamon Press, New York.

Glikson, A.Y., 1971. Primitive Archaean element distribution patterns : Chemical evidence and geotectonic significance. Earth Planet. Sci. Lett., 12 : 309-320.

Glikson, A.Y. and Lambert, I.B., 1976. Vertical zonation and petrogenesis of the early Precambrian crust in Western Australia. Tectonophysics, 30 : 55-89.

Goldich, S.S. and Hedge, C.E., 1975. Reply to Farhat, J.S. and Wetherill, G.W., 1975 : Interpretation of apparent ages in Minnesota. Nature, 257 : 722.

Goldich, S.S. and Wooden, J.L., 1978. Geochemistry of the Archean rocks in the Morton and Granite Falls areas, south-western Minnesota. In: I.E.M. Smith and J.G. Williams (Eds.). Proc. 1978 Archean Geochem. Conf., Univ. Toronto Press, : 285-318.

Goodman, R.J., 1972. The distribution of Ga and Rb in coexisting groundmass and phenocryst phases of some basic volcanic rocks. Geochim. Cosmochim. Acta, 36 : 303-317.

Goodwin, A.M., 1971. Metallogenic patterns and evolution of the Canadian Shield. Geol. Soc. Aust., Spec. Publ., 3 : 157-174.

Goodwin, A.M., 1972. The Superior Province. In: R.A. Price and R.J.W. Douglas (Eds.). Variations in tectonic styles in Canada. Geol. Assoc. Can., Spec. Paper, 11 : 528-623.

Goodwin, A.M., 1978. Archean crust in the Superior geotraverse area : Geological overview. In: I.E.M. Smith and J.E. Williams (Eds.). Proc. 1978 Archean Geochem. Conf., Univ. Toronto Press, : 73-106.

Graf, W.H. and Acaroglu, E.R., 1966. Settling velocities of natural grains. Bull. Int. Assoc. Sci. Hydrol., 11 : 27-43.

Green, D.C., Baadsgaard, H. and Cumming, G.L., 1968. Geochronology of the Yellowknife area, Northwest Territories, Canada. Can. J. Earth Sci., 5 : 725-735.

Green, D.H., 1972a. Magmatic activity as the major process in the chemical evolution of the Earth's crust and mantle. Tectonophysics, 13 : 47-71.

Green, D.H., 1972b. Archaean greenstone belts may include terrestrial equivalents of Lunar Maria? Earth Planet. Sci. Lett., 15 : 263-270.

Green, D.H., 1975. Genesis of Archean peridotitic magmas and constraints on Archean geothermal gradients and tectonics. Geology, 3 : 15-18.

Green, D.H., Nicholls, I.A., Viljoen, M. and Viljoen, R., 1975. Experimental demonstration of the existence of peridotitic liquids in earliest Archean magmatism. Geology, 3 : 11-14.

Gresens, R.L., 1967. Composition-volume relationships of metasomatism. Chem. Geol., 2 : 47-65.

Gülaçar, O.F. and Delaloye, M., 1976. Geochemistry of nickel, cobalt and copper in alpine-type ultramafic rocks. Chem. Geol., 17 : 269-280.

Gunn, B.M., 1971. Trace element partition during olivine fractionation of Hawaiian basalts. Chem. Geol., 8 : 1-13.

Hallberg, J.A., 1972. Geochemistry of Archaean volcanic belts in the Eastern Goldfields Region of Western Australia. J. Petrol., 13 : 45-56.

Hallberg, J.A., Carter, D.H. and West, K.N., 1976b. Archaean volcanism and sedimentation near Meekatharra, Western Australia. Precam. Res., 3 : 577-595.

Hallberg, J.A., Johnston, C. and Bye, S.M., 1976a. The Archaean Madra igneous complex, Western Australia. Precam. Res., 3 : 111-136.

Hallberg, J.A. and Williams, D.A.C., 1972. Archaean mafic and ultramafic rock associations in the Eastern Goldfields region, Western Australia. Earth Planet. Sci. Lett., 15 : 191-200.

Hamilton, P.J., Evensen, N.M., O'Nions, R.K., Smith, H.S. and Erlank, A.J., 1979. Sm-Nd dating of Onverwacht group volcanics, southern Africa. Nature, 279 : 298-300.

Hamilton, P.J., O'Nions, R.K. and Evensen, N.M., 1977. Sm-Nd dating of Archaean basic and ultrabasic volcanics. Earth Planet. Sci. Lett., 36 : 263-268.

Hamilton, P.J., O'Nions, R.K., Evensen, N.M., Bridgewater, D. and Allaart, J.H., 1978. Sm-Nd isotopic investigations of Isua supracrustals and implications for mantle evolution. Nature, 272 : 41-43.

Harrigan, D.B. and MacLean, W.H., 1976. Petrography and geochemistry of epidote alteration patches in Gabbro Dykes at Matagami, Quebec. Can. J. Earth Sci., 13 : 500-511.

Hart, R., 1970. Chemical exchange between sea water and deep ocean basalts. Earth Planet. Sci. Lett., 9 : 269-279.

Hart, S.R., 1969. K, Rb, Cs contents and K/Rb, K/Cs ratios of fresh and altered submarine basalts. Earth Planet. Sci. Lett., 6 : 295-303.

Hart, S.R., 1971. K, Rb, Cs, Sr and Ba contents and Sr isotope ratios of ocean floor basalts. Phil. Trans. Roy. Soc. Lond. A., 268 : 573-588.

Hart, S.R. and Brooks, C., 1974. Clinopyroxene-matrix partitioning of K, Rb, Cs, Sr, Ba. Geochim. Cosmochim. Acta, 38 : 1799-1806.

Hart, S.R. and Brooks, C., 1977. The geochemistry and evolution of early Precambrian mantle. Contrib. Mineral. Petrol., 61 : 109-128.

Hart, S.R., Brooks, C., Krogh, T.E., Davis, G.L. and Nava, D., 1970. Ancient and modern volcanic rocks : A trace element model. Earth Planet. Sci. Lett., 10 : 17-28.

Hart, S.R. and Davis, K.E., 1978. Nickel partitioning between olivine and silicate melt. Earth Planet. Sci. Lett., 40 : 203-219.

Hart, S.R., Erlank, A.J.E. and Kable, E.J.D., 1974. Sea floor basalt alteration : Some chemical and Sr isotopic effects. Contrib. Mineral. Petrol., 44 : 219-230.

Hart, S.R., Glassley, W.E. and Karig, D.E., 1972. Basalts and sea floor spreading behind the Mariana Island arc. Earth Planet. Sci. Lett., 15 : 12-18.

Hart, S.R. and Nalwalk, A.J., 1970. K, Rb, Cs and Sr relationships in submarine basalts from the Puerto Rico trench. Geochim. Cosmochim. Acta, 34 : 145-155.

Hawkesworth, C.J. and Bickle, M.J., 1976. Rhodesian Rb/Sr geochronology from 3.6-2.0 b.a. : a brief review. Ann. Res. Report, Dept. Earth Sci., Univ. Leeds, 20 : 22-27.

Hawkesworth, C.J., Moorbath, S. and O'Nions, R.K., 1975. Age relationships between greenstone belts and 'granites' in the Rhodesian Archaean craton. Earth Planet. Sci. Lett., 25 : 251-262.

Hawkesworth, C.J. and O'Nions, R.K., 1977. The petrogenesis of some Archaean volcanic rocks from southern Africa. J. Petrol., 18 : 487-520.

Hawkesworth, C.J., O'Nions, R.K., Pankhurst, R.J., Hamilton, P.J. and Evensen, N.M., 1977. A geochemical study of island-arc and back-arc tholeiites from the Scotia Sea. Earth Planet. Sci. Lett., 36 : 253-262.

Hekinian, R., Moore, J.G. and Bryan, W.B., 1976. Volcanic rocks and processes of the Mid-Atlantic Ridge Rift Valley near 36°49' N. J. Petrol., 39 : 301-308.

Herrmann, A.G., 1970. Yttrium and lanthanides. In: K.H. Wedepohl (Ed.). Handbook of Geochemistry. Springer-Verlag, Berlin.

Herrmann, A.G., Blanchard, D.P., Haskin, L.A., Jacobs, J.W., Knake, D., Korotev, R.L. and Brannon, J.C., 1976. Major, minor, and trace element compositions of peridotitic and basaltic komatiites from the Precambrian crust of southern Africa. Contrib. Mineral. Petrol., 59 : 1-12.

Humphris, S.E. and Thompson, G., 1978a. Hydrothermal alteration of oceanic basalts by seawater. Geochim. Cosmochim. Acta, 42 : 107-125.

Humphris, S.E. and Thompson, G., 1978b. Trace element mobility during hydrothermal alteration of oceanic basalts. Geochim. Cosmochim. Acta, 42 : 127-136.

Hunter, D.R., 1973. The granitic rocks of the Precambrian in Swaziland. Geol. Soc. S. Africa, Spec. Publ., 3 : 131-145.

Hunter, D.R., 1974. Crustal development in the Kaapvaal Craton, I. The Archaean. Precam. Res., 1 : 295-326.

Hurley, P.M., Pinson, W.H., Nagy, B. and Teska, T.M., 1972. Ancient age of the Middle Marker Horizon, Onverwacht Group, Swaziland Sequence, South Africa. Earth Planet. Sci. Lett., 14 : 360-366.

Irvine, T.N. and Baragar, W.R.A., 1971. A guide to the chemical classification of the common volcanic rocks. Can. J. Earth Sci., 8 : 523-548.

Jahn, B-M., Auvray, B., Blais, S., Capdevila, R., Cornichet, J., Vidal, F. and Hameurt, J., 1978. Trace element geochemistry and petrogenesis of Finnish greenstone belts. J. Petrol. (in press).

Jahn, B-M. and Shih, C-Y., 1974. On the age of the Onverwacht Group, Swaziland Sequence, South Africa. Geochim. Cosmochim. Acta, 38 : 873-885.

Jahn, B., Shih, C. and Murthy, V.R., 1974. Trace element geochemistry of Archean volcanic rocks. Geochim. Cosmochim. Acta, 38 : 611-627.

Jahn, B-M. and Sun, S-S., 1977. Trace element distribution and isotopic composition of Archean greenstones. In: L.H. Ahrens (Ed.). Physics and Chemistry of the Earth, 11 : 597-618. Pergamon Press, New York.

Jolly, W.T., 1975. Subdivision of the Archean lavas of the Abitibi area, Canada, from Fe-Mg-Ni-Cr relations. Earth Planet. Sci. Lett., 27 : 200-210.

Jolly, W.T. and Smith, R.E., 1972. Degradation and metamorphic differentiation of Keweenawan tholeiitic lavas of northern Michigan, U.S.A. J. Petrol., 13 : 273-309.

Krishnamurthy, P. and Cox, K.G., 1977. Picrite basalts and related lavas from the Deccan Traps of Western India. Contrib. Mineral. Petrol., 62 : 53-75.

Krogh, T.E. and Davis, G.L., 1971. Zircon U-Pb ages of Archean metavolcanic rocks in the Canadian Shield. Carnegie Inst. Wash. Yearbook, 70 : 241-242.

Kuenen, Ph.H., 1963. Turbidites in South Africa. Trans. Geol. Soc. S. Africa, 66 : 191-195.

Lajoie, J. and Gélinas, L., 1978. Emplacement of Archean peridotitic komatiites in La Motte Township, Quebec. Can. J. Earth Sci., 15 : 672-677.

Langmuir, C.H., Bender, J.B., Bence, A.E., Hanson, G.N. and Taylor, S.R., 1977. Petrogenesis of basalts from the FAMOUS area : Mid-Atlantic Ridge. Earth Planet. Sci. Lett., 36 : 133-156.

Larimer, J.W., 1971. Composition of the Earth : Chondritic or achondritic. Geochim. Cosmochim. Acta, 35 : 769-786.

Leake, B.E., 1972. The mineralogical modification of the chemistry of metamorphic rocks. Geol. Mag., 109 : 331-337.

Leeman, W.P. and Scheidegger, K.F., 1977. Olivine/liquid distribution coefficients and a test for crystal-liquid equilibrium. Earth Planet. Sci. Lett., 35 : 247-257.

Le Roex, A.P., 1980. Ph.D. thesis. Univ. Cape Town.

Le Roex, A.P. and Reid, D.L., 1978. Geochemistry of Karroo dolerite sills in the Calvinia District, Western Cape Province, South Africa. Contrib. Mineral. Petrol., 66 : 351-360.

Lindstrom D.J., 1976. Experimental study of the partitioning of transition metals between clinopyroxene and coexisting silicate liquids. Ph.D. thesis. Univ. Oregon.

Lugmair, G.W. and Marti, K., 1977. Sm-Nd-Pu timepieces in the Angra Dos Reis meteorite. Earth Planet. Sci. Lett., 35 : 273-284.

Maaløe, S. and Aoki, K., 1977. The major element composition of the upper mantle estimated from the composition of lherzolites. Contrib. Mineral. Petrol., 63 : 161-173.

MacGregor, A.M., 1951. Some milestones in the Precambrian of Southern Rhodesia. Proc. Geol. Soc. S. Africa, 54 : 27-71.

Martin, B., 1971. Some comments on the processes of serpentinisation from experimental and other work. Geol. Soc. Aust., Spec. Publ., 3 : 301-310.

McCall, G.J.H., 1973. Geochemical characteristics of some Archaean greenstone suites of the Yilgarn structural province, Australia. Chem. Geol., 11 : 243-269.

McCarthy, T.S. and Robb, L.J., 1977. On the relationship between cumulus mineralogy and trace and alkali element chemistry in an Archaean granite from the Barberton region, South Africa. Econ. Geol. Res. Unit., Inf. Circ., 112 : 1-15.

McGregor, V.R. and Mason, B., 1977. Petrogenesis and geochemistry of metabasaltic and metasedimentary enclaves in the Amîtsoq gneisses, West Greenland. Amer. Mineral., 62 : 887-904.

McIver, J.R., 1972. Hornblendite from Bon Accord, Pretoria, and a possible komatiite equivalent. Trans. Geol. Soc. S. Africa, 75 :313-315.

McIver, J.R., 1975. Aspects of high magnesia eruptives in Southern Africa. Contrib. Mineral. Petrol., 51 : 99-118.

McIver, J.R. and Lenthall, D.H., 1973. Mafic and ultramafic extrusives of the Onverwacht Group in terms of the system XO-YO-R_2O_3-ZO_2. Econ Geol. Res. Unit., Inf. Circ., 80 : 1-8.

Melson, W.G. and Thompson, G., 1971. Petrology of a transform fault zone and adjacent ridge sediments. Phil. Trans. Roy. Soc. Lond. A., 268 : 423.

Melson, W.G., Thompson, G. and van Andel, T.H., 1968. Volcanism and metamorphism in the Mid-Atlantic Ridge, 22°N latitude. J. Geophys. Res., 73 : 5925-5941.

Melson, W.G. and van Andel, T.H., 1966. Metamorphism in the Mid-Atlantic Ridge, 22°N latitude. Marine Geol., 4 : 165-186.

Meyer, O.A., 1977. Mineralogy of the upper mantle : A review of the minerals in mantle xenoliths from kimberlite. Earth Sci. Rev., 13 : 251-281.

Minnitt, R.C.A., 1975. The geology and geochemistry of the eastern portion of the Murchison Range. Abstract 16th Congr., Geol. Soc. S. Africa, Stellenbosch : 107-108.

Miyashiro, A., 1974. Volcanic rock series in island arcs and active continental margins. Amer. J. Sci., 274 : 321-355.

Miyashiro, A. and Shido, F., 1975. Tholeiitic and calc-alkalic series in relation to the behaviors of titanium, vanadium, chromium, and nickel. Amer. J. Sci., 275 : 265-277.

Moeskops, P.G., 1977. Volume increase serpentinization in Archaean quench texture ultramafic rocks near Kalgoorlie, Western Australia. Geol. Mag., 114 : 41-46.

Moorbath, S., 1977. Ages, isotopes and evolution of Precambrian continental crust. Chem. Geol., 20 : 151-187.

Moorbath, S., 1978. Age and isotopic evidence for the evolution of continental crust. Phil. Trans. R. Soc. Lond. A., 288 : 401-413.

Moorbath, S., O'Nions, R.K., Pankhurst, R.J. and Gale, N.H., 1972. Further rubidium-strontium age determinations on very early Precambrian rocks of the Godthaab District, West Greenland. Nature, 240 : 78-82.

Moorbath, S., Wilson, J.F. and Cotterill, P., 1976. Early Archaean age for the Sebakwian Group at Selukwe, Rhodesia. Nature, 264 : 536-538.

Moore, A.E. and Erlank, A.J., 1979. Unusual olivine zoning-evidence for complex physico-chemical changes during the evolution of olivine melilite and kimberlite magmas. Contrib. Mineral. Petrol., 70 : 391-405.

Moore, J.G., 1975. Mechanism of formation of pillow lava. Amer. Scientist, 63 : 269-276.

Moore, J.G., Cristofolini, R. and Lo Giudice, A., 1971. Development of pillows on the submarine extension of recent lava flows, Mount Etna, Sicily. U.S. Geol. Surv. Prof. Paper, 750-C : C89-C97.

Muir, M.D. and Grant, P.R., 1976. Micropalaeontological evidence from the Onverwacht Group, South Africa. In: F.G. Windley (Ed.) The Early History of the Earth. J. Wiley and Sons, New York : 595-604.

Mysen, B.O. and Kushiro, I., 1977. Compositional variations of coexisting phases with degree of melting of peridotite in the upper mantle. Amer. Mineral., 62 : 843-865.

Naldrett, A.J. and Mason, G.D., 1968. Contrasting Archean ultramafic igneous bodies in Dundonald and Clergue Townships, Ontario. Can. J. Earth Sci., 5 : 111-143.

Naldrett, A.J. and Turner, A.R., 1977. The geology and petrogenesis of greenstone belt and related nickel sulfide mineralization at Yakabindie, Western Australia. Precam. Res., 5 : 43-103.

Nesbitt, R.W., 1971. Skeletal crystal forms in the ultramafic rocks of the Yilgarn Block, Western Australia : Evidence for an Archaean ultramafic liquid. Geol. Soc. Aust., Spec. Publ., 3 : 331-347.

Nesbitt, R.W. and Sun, S-S., 1976. Geochemistry of Archaean spinifex-textured peridotites and magnesian and low-magnesian tholeiites. Earth Planet. Sci. Lett., 31 : 433-453.

Nesbitt, R.W., Sun, S-S. and Purvis, A.C., 1979. Komatiites; Geochemistry and Genesis. Can. Mineral., 17 : 165-186.

Nisbet, E.G., Bickle, M.J. and Martin, A., 1977. The mafic and ultramafic lavas of the Belingwe greenstone belt, Rhodesia. J. Petrol., 18 : 521-566.

Nixon, P.H. and Boyd, F.R., 1973. Petrogenesis of the granular and sheared ultrabasic nodule suite in kimberlites. In: P.H. Nixon (Ed.) Lesotho Kimberlites. Lesotho. Nat. Devel. Co. : 48-56.

Norrish, K. and Hutton, J.T., 1969. An accurate X-ray spectrographic method for the analysis of a wide range of geological samples. Geochim. Cosmochim. Acta, 33 : 431-453.

Numes, P.D. and Thurston, P.C., 1978. Evolution of a single greenstone belt over 220 million years ... A zircon study of the Uchi Lake area north-western Ontario. (Abstr.) 4th Int. Conf. Geochron. Cosmochron. and Isotope Geology. (Ed. R.E. Zartman) U.S.G.S. Open File Report, 78-701 : 313-315.

O'Hara, M.J., 1968. The bearing of phase equilibria studies on synthetic and natural systems on the origin and evolution of basic and ultra-basic rocks. Earth Sci. Rev., 4 : 69-133.

O'Nions, R.K., Carter, S.R., Evensen, N.M. and Hamilton, P.J., 1979. Geochemical and cosmochemical application of Nd-isotopes. An. Rev. Earth Plan. Sci., 7 : 11-38.

O'Nions, R.K., Hamilton, P.J. and Evensen, N.M., 1977. Variations in $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ ratios in oceanic basalts. Earth Planet. Sci. Lett., 34 : 13-22.

Oosthuyzen, E.J., 1970. The geochronology of a suite of rocks from the granitic terrain surrounding the Barberton Mountain Land. Ph.D. thesis, Univ. Witwatersrand, Johannesburg.

Page, N.J., 1967. Serpentinization considered as a constant volume metasomatic process : A discussion. Amer. Mineral., 52 : 545.

Page, N.J., 1976. Serpentinization and alteration in an olivine cumulate from the Stillwater Complex, Southwestern Montana. Contrib. Mineral. Petrol., 54 : 127-137.

Pearce, J.A., 1976. Statistical analysis of major element patterns in basalts. J. Petrol., 17 : 15-42.

Pearce, J.A. and Cann, J.R., 1971. Ophiolite origin investigated by discriminant analysis using Ti, Zr and Y. Earth Planet. Sci. Lett., 12 : 339-349.

Pearce, J.A. and Cann, J.R., 1973. Tectonic setting of basic volcanic rocks determined using trace element analyses. Earth Planet. Sci. Lett., 19 : 290-300.

Pearce, J.A. and Norry, M.J., 1979. Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks. Contrib. Mineral. Petrol., 69 : 33-47.

Pearce, T.H., 1978. Olivine fractionation equations for basaltic and ultrabasic liquids. Nature, 276 : 771-774.

Pearce, T.H. and Birkett, T.C., 1974. Archean metavolcanic rocks from Thackeray Township, Ontario. Can. Mineral., 12 : 509-519.

Pearce, T.H., Gorman, B.E. and Birkett, T.C., 1975. The TiO_2-K_2O-P_2O_5 diagram : A method of discriminating between oceanic and non-oceanic basalts. Earth Planet. Sci. Lett., 24 : 419-426.

Philpotts, J.A. and Schnetzler, C.C., 1970. Phenocryst-matrix partition coefficients for K, Rb, Sr, and Ba, with application to anorthosite and basalt genesis. Geochim. Cosmochim. Acta, 34 : 307-322.

Pyke, D.R., Naldrett, A.J. and Eckstrand, O.R., 1973. Archean ultramafic flows in Munro Township, Ontario. Bull. Geol. Soc. Amer., 84 : 955-978.

Reimer, T.O., 1967. Die Geologie der Stolzburg Synklinale, Barberton Bergland (Transvaal-Südafrika): Diplom-Arbeit, Goethe-Univ., Frankfurt, Germany.

Richard, P., Shimizu, N. and Allegre, C.J., 1976. $^{143}Nd/^{146}Nd$, A natural tracer : An application to oceanic basalts. Earth Planet. Sci. Lett., 31 : 269-278.

Robb, L.J., 1977. A general geological description of the Archaean granitic terrane between Nelspruit and Bush Buckridge, Eastern Transvaal. Econ. Geol. Res. Unit, Inf. Circ., 111 : 1-12.

Roeder, P.L. and Emslie, R.F., 1970. Olivine-liquid equilibrium. Contr. Mineral. Petrol., 29 : 275-289.

Saager, R. and Köppel, V., 1976. Lead isotopes and trace elements from sulfides of Archean greenstone belts in South Africa - A contribution to the knowledge of the oldest known mineralizations. Econ. Geol., 71 : 44-57.

Saunders, A. and Tarney, J., 1979. The geochemistry of basalts from a back-arc spreading centre in the East Scotia Sea. Geochim. Cosmochim. Acta, 43 : 555-572.

Saunders, A.D., Tarney, J., Stern, C.R. and Dalziel, I.W.D., 1979. Geochemistry of mesozoic marginal basin floor igneous rocks from southern Chile. Bull. Geol. Soc. Amer., 90 : 237-258.

Scott, R.B. and Hajash, Jr., A., 1976. Initial submarine alteration of basaltic pillow lavas : A microprobe study. Amer. J. Sci., 276 : 480-501.

Shido, F., Miyashiro, A. and Ewing, M., 1974. Compositional variations in pillow lavas from the Mid-Atlantic Ridge. Marine Geol., 16 : 177-190.

Simkin, T. and Smith, J.V., 1970. Minor-element distribution in olivine. J. Geol., 78 : 304-325.

Sinha, A.K., 1972. U-Th-Pb systematics and the age of the Onverwacht series, South Africa. Earth Planet. Sci. Lett., 16 : 219-227.

Smith, R.E., 1968. Redistribution of major elements in the alteration of some basic lavas during burial metamorphism. J. Petrol., 9 : 191-219.

Smith, R.E. and Smith, S.E., 1976. Comments on the use of Ti, Zr, Y, Sr, K, P and Nb in classification of basaltic magmas. Earth Planet. Sci. Lett., 32 : 114-120.

Sobolev, A.V., 1978. Phase composition of meimechites of north Siberia and some problems of their genesis. (In Russian). In: Problems of Petrology of the Earth's Crust and Upper Mantle : 330-347.

Stowe, C.W., 1971. Summary of the tectonic development of the Rhodesian Archaean craton. Geol. Soc. Aust., Spec. Publ., 3 : 377-383.

Sun, S-S. and Nesbitt, R.W., 1977. Chemical heterogeneity of the Archaean mantle, composition of the Earth and mantle evolution. Earth Planet. Sci. Lett., 35 : 429-448.

Sun, S-S. and Nesbitt, R.W., 1978. Petrogenesis of Archaean ultrabasic and basic volcanics : Evidence from rare earth elements. Contrib. Mineral. Petrol., 65 : 301-325.

Sun, S-S., Nesbitt, R.W. and Sharaskin, A., 1979. Geochemical characteristics of Mid-Ocean Ridge basalts. Earth Planet. Sci. Lett., 44 : 119-138.

Tarney, J., Dalziel, I.W.D. and de Wit, M.J., 1976. Marginal basin "Rocas Verdes" complex from S. Chile : A model for Archaean greenstone belt formation. In: B.F. Windley (Ed.) The Early History of the Earth. J. Wiley and Sons, London : 131-146.

Tatsumoto, M., Hedge, C.E. and Engel, A.E.J., 1965. Potassium, rubidium, strontium, thorium, uranium and the ratio of Sr-87 to Sr-86 in oceanic tholeiitic basalt. Science, 150 : 886-888.

Taylor, S.R., 1964. Trace element abundances and chondritic Earth model. Geochim. Cosmochim. Acta, 28 : 1989-1998.

Thayer, T.P., 1966. Serpentinisation considered as a constant volume metasomatic process. Amer. Mineral., 51 : 685-710.

Thayer, T.P., 1967. Serpentinisation considered as a constant volume metasomatic process : A reply. Amer. Mineral., 52 : 549-553.

Thompson, R.N., 1973. One-atmosphere melting behaviour and nomenclature of terrestrial lavas. Contr. Mineral. Petrol., 41 : 197-204.

Turek, A. and Compston, W., 1971. Rubidium-strontium geochronology in the Kalgoorlie region. Geol. Soc. Aust., Spec. Publ., 3 : 72.

Ulrych, T.J., Burger, A. and Nicolaysen, L.O., 1967. Least radiogenic terrestrial leads. Earth Planet. Sci. Lett., 2 : 179-184.

Upton, B.G.J. and Thomas, J.E., 1973. Precambrian potassic ultramafic rocks : South Greenland. J. Petrol., 14 : 509-534.

Vallance, T.G., 1965. On the chemistry of pillow lavas and the origin of spilites. Min. Mag., 34 : 471-481.

Van Niekerk, C.B. and Burger, A.J., 1969. A note on the minimum age of the acid lava of the Onverwacht series of the Swaziland System. Trans. Geol. Soc. S. Africa, 72 : 9-21.

Van Niekerk, C.B. and Burger, A.J., 1975. The Moodies conglomerate boulders. Abstr. 16th Congr. Geol. Soc. S. Africa, Stellenbosch : 150-153.

Van Schmus, W.R. and Hayes, J.M., 1974. Chemical and petrographic correlations among carbonaceous chondrites. Geochim. Cosmochim. Acta, 38 : 47.

Viljoen, R.P., Saager, R. and Viljoen, M.J., 1970. Some thoughts on the origin and processes responsible for the concentration of gold in the early Precambrian of Southern-Africa. Mineral. Deposita, 5 : 164-180.

Viljoen, M.J. and Viljoen, R.P., 1969a. An introduction to the geology of the Barberton granite - greenstone terrain. Geol. Soc. S. Africa, Spec. Publ., 2 : 9-28.

Viljoen, R.P. and Viljoen, M.J., 1969b. The effects of metamorphism and serpentinization on the volcanic and associated rocks of the Barberton region. Geol. Soc. S. Africa, Spec. Publ., 2 : 29-54.

Viljoen, M.J. and Viljoen, R.P., 1969c. The geology and geochemistry of the Lower Ultramafic Unit of the Onverwacht Group and a proposed new class of igneous rock. Geol. Soc. S. Africa, Spec. Publ., 2 : 55-86.

Viljoen, M.J. and Viljoen, R.P., 1969d. Evidence for the existence of a mobile extrusive peridotitic magma from the Komati Formation of the Onverwacht Group. Geol. Soc. S. Africa, Spec. Publ., 2 : 87-112.

Viljoen, R.P. and Viljoen, M.J., 1969e. The geological and geochemical significance of the upper formations of the Onverwacht group. Geol. Soc. S. Africa, Spec. Publ., 2 : 113-152.

Viljoen, M.J. and Viljoen, R.P., 1969f. A proposed new classification of the granitic rocks of the Barberton region. Geol. Soc. S. Africa, Spec. Publ., 2 : 153-188.

Viljoen, M.J. and Viljoen, R.P., 1969g. The geochemical evolution of the granitic rocks of the Barberton region. Geol. Soc. S. Africa, Spec. Publ., 2 :189-220.

Viljoen, R.P. and Viljoen, M.J., 1969h. The relationship between mafic and ultramafic material derived from the upper mantle and the ore deposits of the Barberton region. Geol. Soc. S. Africa, Spec. Publ., 2 : 221-244.

Viljoen, M.J. and Viljoen, R.P., 1969i. A reappraisal of the granite-greenstone terrains of shield areas based on the Barberton model. Geol. Soc. S. Africa, Spec. Publ., 2 : 245-274.

Viljoen, R.P. and Viljoen, M.J., 1969j. Evidence for the composition of the primitive mantle and its products of partial melting from a study of the rocks of the Barberton Mountain Land. Geol. Soc. S. Africa, Spec. Publ., 2 : 275-296.

Viljoen, M.J. and Viljoen, R.P., 1970a. Archaean vulcanicity and continental evolution in the Barberton region, Transvaal. In: T.N. Clifford and I.G. Gass (Eds.) African Magmatism and Tectonics. Oliver and Boyd, Edinburgh : 27-49.

Viljoen, R.P. and Viljoen, M.J., 1970b. The geology and geochemistry of the layered ultramafic bodies of the Kaapmuiden area, Barberton Mountain Land. Geol. Soc. S. Africa, Spec. Publ., 1 : 661-688.

Viljoen, R.P. and Viljoen, M.J., 1971. The geological and geochemical evolution of the Onverwacht volcanic group of the Barberton Mountain Land, South Africa. Geol. Soc. Aust., Spec. Publ., 3 : 133-150.

Wänke, H., Baddenhausen, H., Palme, H. and Spettel, B., 1974. On the chemistry of Allende inclusions and their origin as high-temperature condensates. Earth Planet. Sci. Lett., 23 : 1-7.

Watson, J., 1976. Mineralization in Archaean provinces. In: B.F. Windley (Ed.) The Early History of the Earth. John Wiley and Sons, London :

Weaver, B.L. and Tarney, J., 1979. Thermal aspects of komatiite generation and greenstone belt models. Nature, 299 : 689-692.

Wetherill, G.W., 1972. The beginning of continental evolution. Tectonophysics, 13 : 31-45.

White, A.J.R., Jakes, P. and Christie, D.M., 1971. Composition of greenstones and the hypothesis of sea-floor spreading in the Archaean. Geol. Soc. Aust., Spec. Publ., 3 : 47-56.

Willett, G., Eshuys, E. and Guy, B., 1978. Ultramafic rocks of the Widgiemooltha-Norseman area, Western Australia : Petrological diversity, geochemistry and mineralisation. Precam. Res., 6 : 133-156.

Williams, D.A.C., 1971. Determination of primary mineralogy and textures in ultramafic rocks from Mt. Monger, Western Australia. Geol. Soc. Aust., Spec. Publ., 3 : 259-268.

Williams, D.A.C., 1972. Archaean ultramafic, mafic and associated rocks, Mt. Monger, Western Australia. J. Geol. Soc. Aust., 19 : 163-188.

Williams, D.A.C. and Furnell, R.G., 1979. Reassessment of part of the Barberton type area, South Africa. Precam. Res., 9 : 325-347.

Williams, D.A.C. and Hallberg, J.A., 1973. Archaean layered intrusions of the eastern Goldfields region, Western Australia. Contrib. Mineral. Petrol., 38 : 45-70.

Willis, J.P., 1978. Some aspects of the geochemistry of gallium in silicate rocks and stony meteorites. Ph.D. thesis. Univ. Cape Town.

Willis, J.P., Ahrens, L.H., Danchin, R.V., Erlank, A.J., Gurney, J.J., Hofmeyr, P.K., McCarthy, T.S. and Orren, M.J., 1971. Some inter-element relationships between lunar rocks and fines, and stony meteorites. Proc. Second Lunar Sci. Conf. Levinson, A.A. (Ed.), 2 : 1123-1138. Cambridge, Mass. : M.I.T. Press.

Willis, J.P. Erlank, A.J., Gurney, J.J., Theil, R.H. and Ahrens, L.H., 1972. Major, minor and trace element data for some Apollo 11, 12, 14 and 15 samples. Proc. Third Lunar Sci. Conf., Heymann, L.D., (Ed.), 2 : 1269-1273. Cambridge, Mass. : M.I.T. Press.

Wilson, J.F., 1973. The Rhodesian Archaean craton - An essay in cratonic evolution. Phil. Trans. R. Soc. Lond. A., 273 : 389-411.

Wilson, J.F., Bickle, M.J., Hawkesworth, C.J., Nisbet, E.G., Martin, A. and Orpen, J.L., 1978. Granite - greenstone terrains of the Rhodesian Archaean Craton. Nature, 271 : 23-27.

Winchester, J.A. and Floyd, P.A., 1977. Geochemical discrimination of different magma series and their differentiation products using immobile elements. Chem. Geol., 20 : 325-343.

Wright, T.L., 1971. Chemistry of Kilauea and Mauna Loa lava in space and time. U.S. Geol. Surv. Prof. Pap., 735 : 1-40.

Zindler, A., Brooks, C., Arndt, N.T. and Hart, S., 1978. Nd and Sr isotope data from komatiitic and tholeiitic rocks of Munro Township, Ontario. (Abstr.) 4th Int. Conf. Geochron. Cosmochron. and Isotope Geology (Ed. R.E. Zartman) U.S.G.S. Open File Report, 78-701 : 469-471.