Geochemical and petrological investigations of the deeper portions of the Ontong Java Plateau: Malaita, Solomon Islands

Thesis submitted for the degree of Doctor of Philosophy at the University of Leicester

by

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Abstract

Geochemical and petrological investigations of the deeper portion of the Ontong Java Plateau: Malaita, Solomon Islands

Tanya Louise Babbs

The island of Malaita in the Solomons Islands represents a unique opportunity to study the deeper portions of the largest oceanic plateau discovered to date - the Ontong Java Plateau (OJP). Malaita exposes several kilometres of Cretaceous OJP lavas, stratigraphically named the Malaita Volcanic Group (MVG), within en-echelon inliers. Most of the lavas have suffered lowgrade (brownstone facies) metamorphism. The MVG basalts are low-K tholeiites, displaying a limited range of major and trace element concentrations, consistent with the lavas being well mixed prior to eruption, probably in shallow level magma chambers. However, two distinct groups, Type A and Type C-G, can be identified from their rare earth elements (REEs), platinum group elements, Sr and Nd radiogenic isotopes, indicating that mixing has not effectively wiped out all source-related and melting signatures.

The MVG basalts were formed by extensive melting, mostly within the spinel lherzolite field as indicted by their chondritic REE patterns, suggesting a ridge-centred eruption. However, melting in the MVG's source started within the garnet lherzolite field, consistent with a thermal anomaly such as a decompressing mantle plume. Paradoxically for a ridge centred plume, the MVG erupted in a deep submarine environment, when the elevation caused by a surfacing plume would surely imply a subaerial eruption environment, such as seen in Iceland. In addition, the MVG's overlying sedimentary sequence implies a history of uplift, rather than that of subsidence which would be expected as the plume wanes or crust is transported away.

The OJP plume was probably heterogeneous, consisting of enriched (HIMU) streaks in a depleted matrix. This depleted component is thought not to be entrained upper MORB material, but a depleted component intrinsic to the plume itself, possibly entrained as the plume ascended through the lower mantle.

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Chapter 1

The Ontong Java Plateau - rationale for study

1.1 Introduction to LIPs, plumes and plateaus

Large Igneous Provinces

Large Igneous Provinces (LIPs) are vast magmatic provinces, consisting of predominately mafic extrusive and intrusive rocks, and have been the subject of much recent speculation and debate. They include continental flood basalts (CFBs), volcanic margins, aseismic ridges, seamount groups and oceanic plateaus such as the Ontong Java Plateau (Coffin and Eldholm, 1991; Figure 1.1). The voluminous magmatism required in LIP formation has caused them to be recognised as major global events, whose genesis and evolution are directly linked to mantle dynamics. While our modern day theories of plate tectonics account for the slow, steady magmatic activity at mid ocean ridges and subduction zones (active volcanic arcs), it can't explain the vigorous voluminous lava outpourings of LIPs, sometimes erupted through the middle of normally inactive lithospheric plates. Elevated mantle temperature are required which has been attributed to mantle plumes or hotspots (Wilson, 1963; Morgan, 1971; White and McKenzie, 1989).

The increase in magmatic activity associated with LIPs, whether submarine or subaerial, may have had a global effect on past climates due to the increase on the Earth's production of heat, magma and volatiles. Such activity would doubtless have changed the chemistry and circulation of the hydrosphere and atmosphere, with direct and devastating consequences to life on earth. Thus several LIPs, e.g. Kerguelen, Deccan and Columbia River, have been linked to mass extinctions (Coffin and Eldholm, 1993). This includes the K-T boundary demise of the dinosaurs which occurred simultaneously with the Deccan Trap and Mascarene Plateau formation.

Plumes

Mantle plumes are widely accepted to provide the anomalously-high energy required to explain the voluminous magmatism required in LIPs generation. Plume models generally invoke



Chapter 1: The Ontong Java Plateau - rationale for study

Figure 1.1

the presence of a hot 'jet' of buoyant material ascending through the mantle originating from a boundary layer, the core-mantle boundary (D") or the 670 km discontinuity. The plume ascends until it impinges on the base of the lithosphere causing dynamic uplift, or doming, spreading out laterally to form a zone of high ambient temperature mantle, achieving a diameter of approximately 1000 km (Courtney and White, 1986) or even as large as 2000 km (White and McKenzie, 1989) beneath a stationary plate. Voluminous magmatism results because of decompressional melting of the plume material, either the result of extension within the over-lying plate, where the plume has a passive role responding to structural changes in the overlying lithosphere (e.g. White et al., 1987; McKenzie and Bickle, 1988; White and McKenzie, 1989; Figure 1.2a) or because the plume plays a more active role, driving plate separation. Material within the ascending plume head may already be above its solidus prior to lithosphere impact, resulting in the production of extensive melting without the need for any substantial lithospheric extension (e.g. Griffiths and Campbell, 1990; Richards et al., 1989; Figure 1.2b). Morgan (1981) suggest that such active plume roles in LIP production may even promote continental break-up. The debate about whether plumelithosphere interactions is an active or passive process (e.g. Saunders et al., 1992) will doubtless continue.

Oceanic plateaus versus continental flood basalts

Oceanic plateaus are greatly over thickened portions of oceanic crust (≈ 15 to 40 km) present in some of the world's oceans. They cover areas of more than 1,500,000km² and approximately 3% of the ocean floor (Sandwell and MacKenzie, 1989). They form broad, isostatically compensated, shallow regions of the sea floor (elevated some 2 to 3 km above normal), moving with the oceanic plates in which they are embedded (e.g. Coffin and Eldholm, 1994). The Pacific Ocean contains a number of the larger oceanic plateaus, the Manihiki Plateau, Nauru Basin, Magellan, Hess and Shatsky Rise (Figure 1.1), including the largest discovered to date - the Ontong Java Plateau. Although the importance of these oceanic plateaus has now been realised, little concise sampling and data have actually been obtained. Several Deep Sea Drilling Project (DSDP) and Ocean Drill Program (ODP) boreholes have been drilled into these Pacific plateaus over the last 20 years (Figure 1.3). But the only drill hole to penetrate the basement lavas of an oceanic plateau (with the exception of the Nauru Basin, DSDP Leg 61 Site 462) by more than one hundred metres was at Site 807 during ODP Leg 130 on the Ontong



Figure 1.2



Figure 1.3. ODP and DSDP sites and depths of plateaus drilled to basement in the Pacific (Berger et al., 1991).

Java Plateau (Mahoney et al., 1993).

In contrast, continental flood basalt (CFB) provinces have been studied in detail for many years, and consequently much knowledge has been gained about their geochemistry and accretion processes (e.g. Macdougall, 1988; Upton, 1988; Reidel and Hooper, 1989). However a lack of knowledge and understanding of oceanic plateaus still remains, not least because of the obvious submarine environment of these plateaus and their thick sedimentary covers. One of the major advantages of studying oceanic plateaus compared to CFBs is that they have not erupted through the continental crust and therefore we are able to ignore the complication of the role of the continental crust and old lithospheric mantle that occurs in the generation of CFBs (Mahoney et al, 1993).

1.2 The Ontong Java Plateau

The largest oceanic plateau discovered to date is the Ontong Java Plateau (OJP), covering an area greater than $1.5 \times 10^6 \text{ km}^2$ with an approximate volume of 5-6 x 10^7 km^3 (Mahoney, 1987; Coffin and Eldholm, 1991). The plateau has been drilled, with three sites by DSDP Leg 30 and ODP Leg 130 penetrating through to the basement volcanic rocks. A unique opportunity to study part of the OJP exists in the Solomon Islands as portions of the plateau have been obducted on the islands of Malaita and Santa Isabel (and possibly Makira).

It was originally speculated that the OJP erupted in less than 3 m.y., between 121 and 124 m.y., at the beginning of the Cretaceous Normal Polarity Superchron (Tarduno et al., 1991). However, more recent discoveries of a later stage \approx 90 m.y. magmatic episode (Mahoney et al., 1993) has caused such theories to be revised. A summary of ⁴⁰Ar-³⁹Ar ages determined to date on samples recovered from the main body of the OJP and portions of the OJP obducted in the Solomon Islands are summarised in Table 1.1, and their locations shown in Figure 1.4.

The timing and duration of the OJP magmatism appears to be dominated by two widespread, major plateau building events dated at 122 m.y. and ≈90 m.y., with no evidence as yet of any obvious age progression across the plateau. Lavas erupted during the 122 m.y. event have been recovered from ODP Site 807 and DSDP Site 289 (Mahoney et al., 1993), aswell as being a dominant age of magmatism in portions of the SW margin of the OJP obducted in several of the Solomon Islands (Mahoney et al., 1993; Tejada et al., 1996), which lie some

1500 km south of the drill sites (Figure 1.4). The second phase of magmatism which occurred around 90 m.y., has been dated in samples drilled by the ODP at Site 803 and also form part of the OJP basement exposed on the island of Santa Isabel. To date no 90 m.y. plateau lavas have been dated from Malaita; all seem to group in the 122 m.y. phase of activity, with seven new dates indicating age of eruption to be between 121 and 125 m.y. (Duncan, Mahoney, Neal and Tejada, unpublished data).

ODP	ODP	DSDP	Malaita	Santa Isabel	Ramos	Makira
Site 807	Site 803	Site 289			Island	
(Units A and			(Older series	(Sigana		
C-G)			or MVG)	Basalts)		
[1]	[1]	[1]	[2,3]	[2]	[2]	[4]
121.0±4.5		121.7±2.7	121.8±2.4	122.9±1.5	119 <i>.</i> 6±1.6	
121.4±1.9		122.8±2.4	120.8±2.4			
119.9 ± 2.6			plus / ages			
123.4 ± 2.5			101 105 m v			
124.7 ± 2.2			(unpub)			
122.112.2			(unpub.)			
	88.2±1.1			92.0±1.6		
	86.1±4.6†*			94.5±3.1		
	83.7±3.1†			89.8±1.8		
	93.9±1.4†					
an manage of the second sec						
	1					21 (A C) / M
-	4			60 0+1 6		63 0+0 5
				00.911.0		03.0±0.5
				1		A44
						33.9±0.7

Shaded bars indicate apparent age gaps in the plateau lavas, possibly representing true gaps in volcanic activity but may be the result of sampling bias. Ages are in millions of years \pm 1 standard deviation. †Age determined from feldspar separates. 86.1 \pm 4.6 \dagger * was from a more altered mineral fraction.

[References: 1, Mahoney et al., 1993; 2, Tejada et al., 1996; 3, Duncan et al., unpublished data; Petterson et al., 1995c].

Table 1.1. ⁴⁰Ar - ³⁹Ar ages for the Ontong Java Plateau tholeiites.

This apparent bimodal age distribution for the OJP lavas has been complicated by recent dates from OJP-type lavas exposed on the island of Makira (Section 2.8). Whether these Makira OJP-type lavas are truly related to the same plume as that responsible for the main body of the OJP, or are the results of melting of a completely separate, younger plume source



Mahoney, C.R. Neal, R.A. Duncan and M.L.G. Tejada, unpublished data). The OJP is outlined by the 4000 m contour.

Figure 1.4

region is a matter of debate. Many questions on the age of the OJP eruption remain unanswered and will remain until further drilling and dating on the main body of the OJP is carried out.

The melting and thermal buoyancy resulting from mantle plumes means that basaltic sequences of some oceanic plateaus, e.g. the Kerguelen Plateau (Schlich and Wise, 1992), indicate the lavas were emplaced subaerially. The best modern day analogue is perhaps lceland, the elevation being caused by the Iceland plume, which has captured a segment of the northern Mid-Atlantic Ridge. The OJP appears to have erupted in a definite submarine environment, indicated by borehole core samples and more recently the Malaitan basement sequences. Malaita exposes several kilometres of OJP with *en-echelon* inliers (Figure 1.5), enabling us to sample to far greater depths of the OJP than previous drilling has so far permitted. It seems the OJP was not only erupted in a submarine environment, but the portions of the OJP obducted on Malaita appear to have been erupted in a deep marine environment, below the CCD.

Superficially, the OJP shows major and trace element abundances which are similar to those in MORB. Although trace element ratios of highly to moderately incompatible elements and Pb, Nd and Sr isotopic data imply they are not normal MORB lavas but similar to enriched or transitional type MORB (e.g. Floyd, 1989; Saunders, 1986; Mahoney, 1987). REE inversion modelling implies they are results of more extensive degrees of melting, at a greater average depth than normal MORB (Mahoney et al., 1993). It is therefore suggested that the OJP erupted as the result of a mantle plume, probably interacting with a spreading ridge (e.g. Mahoney, 1987).

Mantle plumes generally, following the main episode of volcanism, continue to produce smaller amounts of melting typically in the form of aseismic ridges or seamount chains on the moving plate (Morgan, 1972). However, in the Pacific, where some of the largest oceanic plateaus can be found, there is an absence of obvious hotspot trails leading away from various plateaus. The Ontong Java Plateau is one of the plateaus which lacks a present day active plume source. The situation in the Pacific is very different to the Indian and Atlantic oceans were clearly defined hotspot trails can be directly traced to their major episodes of magmatism, e.g. Chagos-Laccadive Ridge and the Deccan Traps; Rajmahal-Ninetyeast Ridge-Kerguelen Plateau





and the Kerguelen Islands; the North Atlantic Tertiary Province/Greenland/Faeroes Ridge and Iceland.

1.3 Aims of this study

The aims of this study are straightforward. To carry out comprehensive field sampling of exposed portions of the OJP obducted on the island of Malaita, followed by complementary petrographic (optical and electron microprobe), geochemical (XRF, INAA, ICP-MS) and isotopic (Pb-Sr-Nd-Hf) investigations.

By sampling an obducted tract of plateau basement to deeper levels than previously drilled, I hope to offer a better insight into the eruption environment of the OJP, magmatic plumbing and mantle melting processes. Finally, the ultimate aim of this project is to shed some light on the nature of the plume responsible for eruption of the Ontong Java Plateau, i.e. its composition and mantle boundary origin.

Chapter 2

The Solomon Islands and their tectonic evolution

2.1 Introduction

The Solomon Islands are located in the south-west Pacific Ocean some 800 km east of mainland Papua New Guinea (Figure 2.1). They form a double chain of islands flanked on either side by parallel trenches, which stretch for more than 900 km and trend in a WNW-ESE orientation. Little was known about the geology of the Solomon Islands until the 1950s, and in 1965 the first 1:1,000,000 geological map of the Solomon Islands was produced by Coleman (1965). By 1975, most of the major islands which make up the Solomon Islands had been mapped to a 1:50,000 scale.

Structurally, the Solomon Island chain also includes the islands of Bougainville and Buka to the NW, although politically they are part of Papua New Guinea. Whilst the Santa Cruz islands to the SE may nationally belong to the Solomon Islands, they are considered to be a northerly extension of the New Hebrides Group, separated from the main body of islands by the 6000 m-deep Torres Trench.

Recently, much interest has been focused on the islands of Malaita, Small Malaita (Maramasike), Ulawa and NE Santa Isabel. Coleman (1965) first noted that these islands were distinctive in their structure and sequence of lithologies and microfossils, distinctly different from the volcanically young New Georgia Group and the metamorphosed basement of Choiseul and Guadalcanal (Kroenke, 1972; Coleman and Packham, 1976; Coleman and Kroenke, 1981; Kroenke et al., 1986). These geologically distinct islands shape Coleman's Pacific Province, one of the three provinces (Figure 2.2) that make up the Solomon Island group (Coleman, 1965). Geophysical data (Kroenke, 1972; Resig et al., 1986) showed these islands to be an obducted portion of the SW margin of the Ontong Java Plateau (OJP). The OJP represents a portion of over-thickened oceanic crust (Hussong et al., 1979) and plays a vital role in the tectonic history of the Solomon Islands. It was the collision of the OJP with the North Solomon Trench in the Lower



Chapter 2: The Solomon Islands and their tectonic evolution

Figure 2.1

Chapter 2: The Solomon Islands and their tectonic evolution



Figure 2.2

Miocene (Yan and Kroenke, 1993) which caused a reversal in subduction direction. Subsequently, uplift of the SW margin of the plateau caused it to be exposed in several (Malaita, Small Malaita, NE Santa Isabel, Ulawa and possibly Makira) of the Solomon Islands. This chapter outlines the tectonic evolution of the Solomon Islands, introduces the important role of the Ontong Java Plateau in the history, and summarises the unusual present-day seismic and volcanic activity in the region.

2.2 A troublesome history of geological expeditions in the Solomon Islands

Previous Solomon geological expeditions have proved to be difficult, even to the extent of being fatal. Hostile tribes, cannibals and head hunters (!) as well as difficult access and terrain helped to deter geologists. It wasn't until the 1950s when, after the Second World War, the first real progress was made. Several surveys were attempted and a brief history follows:

In 1881 Dr H.B. Guppy on the H.M.S. *Lark* visited the Solomon islands of Choiseul, Florida, San Cristobal, Savo and the Shortlands, subsequently publishing "The Solomon Islands - Their Geology, General Features, and Suitability for Colonisation" in 1887. This remained the only published source of information on the Solomon Islands until 1955. It was the ill-fated Austrian "*Albatross*" expedition in 1896 which caused 'geological discouragement' until the 1950s.

The "*Albatross*" expedition to Guadalcanal, sponsored by the Geographical Society of Vienna, consisted of 18 armed soldiers and scientists and four beach natives as guides. Five days after crossing the plains of Guadalcanal the group reached the steep slopes of the sacred mountain Tatuve. The locals told the expedition members of their belief that if anyone climbed the mountain of their great spirit, then all their people would die. The Austrians chose to ignore their warnings, stating they had come a long way to climb the mountain and would not return without doing so. On the morning of the attempt the party was surrounded while breakfasting, but as "great pity was felt for the white men about to die" the natives decided the least they could do was wait and let them finish their meal so they could fight on a full stomach*. They waited for the party to split in two before overwhelming numbers of naked painted warriors armed with battle axes attacked. Five of the Europeans lost their lives in the attack, and one died later from his wounds (Grover et al., 1965).

[* as told to J.C. Grover by the last surviving member of the attackers].

A small amount of gold prospecting occurred in the 1930s but was halted by the Japanese invasion in 1942, and after the war new mining rules caused the withdrawal of large scale American interest. It was during the 1950s that the British government appointed a single geologist, J.C. Grover, who was presented with the 'rare' opportunity to research several large islands, 1500 to 2000 km² of unexplored jungle-covered mountains stretching over a 1600 km long region of the Pacific. As a result of the work of several geologists, P.J. Coleman produced stratigraphic and structural notes in 1965 to accompany the first geological maps of the Solomons (1959 to 1962).

2.3 Coleman's Provinces within the Solomon Islands

Grover and Coleman's early mapping and subdivision of the Solomon Islands into three provinces (Figure 2.2) was of fundamental importance towards early interpretations of the geological evolution of the Solomon Islands. Each province and its unique sequence of lithologies suggests an independent formation history from the rest (Coleman, 1965). Conclusions from the early studies show the complex geology and tectonic history of the Solomon Islands to be far removed from that of a simple island arc system. It includes subduction flip, obduction of the Ontong Java Plateau (OJP), and is the site of subduction of an active spreading ridge.

Coleman (1965) originally divided the Solomon Islands into (i) the Pacific Province (an obducted portion of the OJP); (ii) the Central Province (the arc remnants of an ancient SW dipping subduction zone, otherwise referred to as Stage 1 or the Vitiaz Arc); and (iii) the Volcanic Province (above the currently active NE dipping subduction zone, Stage 2 or South Solomon Arc) (Figure 2.2). Coleman changed his province boundaries several times; the island of Makira (San Cristobal), originally in the Central Province (Figure 2.2), was later partly included in the Pacific Province (Coleman and Hackman, 1974). Several islands record a complicated geological history and therefore cannot be isolated into a single province. For example, NE Santa Isabel is an obducted portion of the OJP and belongs in the Pacific Province while to the SW of the Kapipito-Korighole Fault (KKF) lies an assemblage of ophiolites and arc type volcanic rocks (Hawkins and Barron, 1991) that belongs to the Central Province. Table 2.1 presents a summary of the main islands in the Solomon Islands, their individual basement and/or volcanic arc

	Malaita*	Santa Isabel*	Makira	Ulawa*	Guadalcanal	Choiseul	Florida Islands	Shortland Islands	New Georgia Group	Russell Islands
	Malaita Volcs.	Sigana Basalts	Warahito	Oroa Basalts						
Ontong Java	Group (MVG)	121.3±0.9 m.y.	Lavas	≈120 m.y. (?)						
Plateau	124-121 m.y.	92.0±1.6 m.y.	35 & 63 m.y.†	(correlate			-	-	-	-
	(Ar-Ar)	(Ar-Ar)	(Ar-Ar)	to MVG)						
	[1,2]	[2]	[3]	[6,10]						
				,	Itina Ultrabasics	Choiseul Schists				
N-type		62-46 m.y.	35 m.y.		& Mbirao Volcs.	(age of amph.				
MORB	-	(Ar-Ar)	(Ar-Ar)	-	92±20 m.y.	metamorphism)	-	-	-	-
Basement		San Jorge			(K-Ar)	44±18 m.y.				
		Volcanics	[3]		[5,9]	(K-Ar) [4,7]				
		(Back arc basin					Hanuvaine Belt	Masamasa		
Vitiaz		and arc type			Poha Diorites	Maetambe	36.7±0.4 m.y.	Volcanics		
(Stage 1) Arc		volcanics)			24.4±0.3 m.y.	Volcanics	38.4±0.7 m.y.	& Alu		
(45 (?) -	-		-		(K-Ar)	(mid Miocene)	Sota Belt	basalts/andesites	-	-
15 m.y.)							35.2±1.4 m.y.	(Oligocene-		
							44.7±2.1 m.y.	lower Miocene)		
		[2]			[8,9]	[7]	(K-Ar) [5]	[5,6,7]		
									2.3±1 m.y.	
South Solomon			Maihada		Gallego volcs.	Komboro		Togha	(K-Ar)	Pavuvu
(Stage 2) Arc			Volcanics		6.4±1.9 m.y.	Volcanics		Pyroclastics	plus	Breccias &
(5 m.y	-	-	(Miocene)		Kolula Diorite	(Pliocene)	?	(Pliocene-	presently active	Banika Lavas
Recent)			2222222222		2.4±1.5 m.y.		1	Holocene)	volcanic centres	(late Miocene
					(K-Ar)				ie Simbo	Holocene)
					[5,9]	[7]	1	[5,6,7]	[5,6]	[6

* The alkali volcanics from Malaita, Santa Isabel and Ulawa are omitted as are the Malaitan Alnoites. † The 63 m.y. and 35 m.y. Makira basement is included as a member of the OJP because they are geochemically similar to the 122 m.y. and 90 m.y. OJP basalts.

Table 2.1:Major basement and/or volcanic units in their associated tectonic framework for the main islands within the Solomon Island
archipelago. [References: 1, Mahoney et al., 1993; 2, Tejada et al., 1996; 3, Petterson et al, 1995c; 4, Richards et al., 1966;
5, Pound, 1986; 6, Coulson and Vedder, 1986; 7, Ridgeway et al., 1987; 8, Chivas, 1981; 9, Hackman, 1980; 10, Danitofea, 1981].

lithologies and the associated tectonic origins. For a more detailed account of the geology within individual major Solomon Islands, readers are referred to Pound (1986) and Coulson and Vedder (1986).

2.4 Age constraints of tectonic events effecting the Solomon Islands

The sequence of tectonic events which affected the Solomon Islands has been attributed to plate motion changes, docking events, subduction flip and oceanic plateau obduction. These events and their timing largely rely on a scant collection of geochronological ages. Therefore, whilst this study has no desire to directly criticise previous work, possible weak points need to be highlighted.

Apart from recent ⁴⁰Ar-³⁹Ar dating carried out on the islands of Malaita, Makira and Santa Isabel, geochronological dating in the Solomon Islands is restricted to a small number of K-Ar dates, often with large errors. The ages of the arc volcanic rocks, assisted by stratigraphical and palaeontological constraints, fall into a bimodal distribution and an age gap of approximately 10-20 m.y. is apparent. However, these few, poorly dated samples have been interpreted to give a highly complex tectonic story. For example, the docking of the OJP with the North Solomon Trench (Section 2.5.4) and resulting subduction flip (Section 2.5.5) is dated by the apparent gap in volcanic activity between \approx 20-25 m.y. and \approx 5 m.y.. Though there is no reason why this time gap is not real, it is more likely to be simply a sampling fault and more widespread, detailed sampling may eventually fill the age gap. If this time gap is purely a result of limited sampling, then the timing of the subduction flip would become harder to pinpoint, for the composition of the volcanic rocks do not give a signature allowing us to distinguish the polarity of the subduction zone from which they are generated. Therefore tectonic event timing would rely more on the plate motion reconstructions (e.g. Yan, 1988; Harbet and Cox, 1989; Yan and Kroenke, 1993) calculated via palaeomagnetic data and hotspot tracks.

Drill cores recovered by DSDP Leg 30 and ODP Leg 130 show sedimentary hiatuses and volcanic ash layers in the OJP's sedimentary cover sequence (Kroenke et al., 1993). These ash layers and sedimentary hiatuses have been used to help constrain the timing of plate motion changes and arc activity affecting the OJP. Sedimentary hiatuses are interpreted as records of catastrophic mass transport or non-deposition/erosion caused by current activity (Kennett, 1982; Kroenke et al., 1993) which coincide with intensified bottom water flow related to the opening and closing of oceanic gateways thought to be caused by major plate motion changes (e.g. Kennett et al., 1975; Kennett, 1982; Keller et al., 1987; Theyer et al., 1989; Berger et al., 1991). While volcanic ash layers reveal the presence of regional volcanic activity (Kroenke et al., 1993), whether plate boundary island arc or intra-plate activity (e.g. the Haruta Limestone Formation; Section 3.6.4). Therefore, such ash layers and sedimentary hiatuses within the OJP's cover sedimentary sequence have been related to, and apparently confirm, the complex tectonic events which affected the Solomon Islands. However, while there continues to be a lack of accurate geochronological ages on the extensive arc volcanic rocks exposed within the Solomon Islands, our understanding of the sequence of tectonic events and their timing remains somewhat tenuous.

2.5 The tectonic evolution of the Solomon Islands

2.5.1 Introduction to the Ontong Java Plateau and the tectonic history of the Solomon Islands

Two major plateau-building events for the Ontong Java Plateau (OJP) have been dated using ⁴⁰Ar-³⁹Ar, the first occurring at approximately 122 Ma and the second at 90 Ma (Mahoney et al., 1993, Tejada et al., 1996) (more details on the age of formation of the OJP are given in Chapter 1). Due to their buoyant, overthickened nature (the OJP may have a crustal thickness as much as 32 km (Gladczenko et al., in press), or 35 to 42 km (Furumoto et al., 1976; Hussong et al., 1979)), oceanic plateaus have the potential for preservation if, at some point in their history, they collide with a subduction zone (e.g. Kroenke, 1974; Nur and Ben-Avraham, 1982; Cloos, 1993; Saunders et al., 1996). The Solomon Islands has experienced such a collision, the result being subduction flip and the eventual obduction of the OJP forming the basement within Malaita, NE Santa Isabel and Ulawa, plus possibly Makira. The other major islands are a combination of a Cretaceous/Paleogene MORB basement and Eocene to present-day arc volcanic rocks resulting from opposite polarity subduction zones (Table 2.1). With the current subduction of the active Woodlark spreading ridge, the Solomon Islands make up a unique and complex collage of past and present volcanic terrains.

2.5.2 Extension of the Ontong Java Plateau and younger volcanism

The SW region of the OJP underwent extension during the Eocene, thought to be caused

by the passing of the OJP over the Samoan (or possibly the Rarotongan) hotspot (Yan and Kroenke, 1993). Extension resulted in the formation of the Faufaumela Basin (Petterson, 1995b; Section 3.8.3), intrusion of younger alkali basalt suites such as the Maramasike Volcanic Formation at 44 Ma (Tejada et al., 1996; Section 3.6.3) and the emplacement of the 34 Ma ultramafic alnoite pipes (Davis, 1977; Section 3.7), all of which are present on Malaita. The Faufaumela Basin is situated in north-central Malaita and is possibly the best example of this extension period preserved in the Solomon Islands, with thickened units of Eocene (Haruta Limestone Formation, Section 3.6.4) through to Pliocene (Suafa Limestone Formation, Section 3.6.5) sediments dating its formation (Petterson, 1995b; Petterson et al., in press). The Faufaumela Basin is the only region on Malaita and Small Malaita where the unusual ultramafic alnoite pipes crop out (Nixon, 1980; Neal and Davidson, 1989), and is one of only four main structural areas where the younger alkali basalt intrusives (Hughes and Turner, 1976, 1977; Tejada et al., 1996) are known to occur. Predominately an extensional feature, possibly a graben (Petterson, 1995b; Petterson et al., in press), the Faufaumela Basin is thought to have formed a convenient pathway allowing the intrusion of the alkali suites (44 m.y.) and later the alnoite pipes (34 m.y.).

2.5.3 Vitiaz (Stage 1) Arc development

Ash layers and hiatuses recovered from cores drilled into the sedimentary cover from the main body of the OJP (DSDP Site 289 and ODP Site 803) correspond with a major plate motion change for both the Indo-Australia and Pacific Plates around 43 Ma/Chron 19 (Kroenke et al., 1993). This is also in accordance with the time of obduction and ophiolite emplacement in Eastern Papua New Guinea, as implied by K-Ar dates of 52 ± 1 m.y. and 42 ± 4 m.y. on metamorphic amphiboles from the basal region of the overthrust oceanic crust (Davies, 1980). It was during this mid-Eocene plate reorganisation that the Pacific Plate began to subduct in a SW direction beneath the north-easterly moving Australia-India Plate (Figure 2.3a) resulting in the initial formation of the Vitiaz Arc (Stage 1) volcanism. Arc-derived turbidite sedimentation in Malaita's cover sequence, the Haruta Limestone Formation (Section 3.6.4), indicate the OJP drifted within reach of the Vitiaz arc fallout during the Eocene. Volcanic ash layers deposited in the mid Eocene to Oligocene/Miocene sedimentary sequence have also been recorded in DSDP Sites 288 and 289 and ODP Site 807 (Kroenke et al., 1993). This SW directed-subduction resulted in the


formation of the North Solomon/Vitiaz Trench system and erupted the Vitiaz Arc (Stage 1) volcanic rock sequences which occur on the islands of Guadalcanal, Choiseul and south Santa Isabel. The volcanism also caused the initial formation of a number of the smaller groups of islands in the Solomon Islands, the Shortland Islands and possibly the Florida Islands (Richards et al., 1966; Pound, 1986; Ridgeway et al., 1987; Chivas, 1981) as summarised in Table 2.1. These islands (with the exception of the Shortland Islands) make up Coleman's Central Province (Figure 2.2) which essentially is the remnants of the old SW facing arc (Coleman and Kroenke, 1981) or Vitiaz (Stage 1) Arc (Ridgeway et al., 1987). The Poha Diorites from Western Guadalcanal have been K-Ar dated at 24±0.3 m.y. (Chivas, 1981; Table 2.1). The Poha Diorites are believed to be the youngest igneous rocks in the Solomon Islands related to the SW dipping subduction zone. Therefore cessation of the Vitiaz Arc volcanism, constrained admittedly somewhat poorly by the Poha Diorites, has been bracketed around 25-20 m.y. (Coleman and Kroenke, 1981; Kroenke, 1984; Yan and Kroenke, 1993).

2.5.4 The Ontong Java Plateaus docking with the North Solomon/Vitiaz Trench

The interruption of the Vitiaz volcanism is believed to have been caused by the OJP, embedded in the Pacific plate, docking with the North Solomon/Vitiaz Trench. This docking event successfully choked the subduction zone, when the SW margin of the OJP, being too thick and buoyant to subduct, collided with the forearc of the Vitiaz Arc (Figure 2.3b). The initial collision of the OJP with the subduction zone caused no major compressive deformation in the lower/mid Miocene sedimentary cover overlying the basaltic basement and is therefore considered to be a 'soft docking' event (Petterson et al., in press).

Volcanic activity appears to have ceased during the Miocene, except for a brief burst around 15 m.y. with the eruption of the Maetambe Volcanics on Choiseul possibly from resumed SW-directed subduction (Kroenke et al., 1993) (Figure 2.3c). This reactivation of the SW dipping slab has been suggested to be the result of a slight change in plate motion for both the Pacific and Australia-India plates (Yan, 1988) indicated by sedimentary hiatuses of Lower Miocene age found at DSDP Site 288 and ODP Sites 803 and 804 (Kroenke et al., 1993). Indeed, the SWdipping subduction zone seems to have been active intermittently from the lower/mid Miocene to the present day, though is no longer considered to be the major cause of volcanism in the Solomon Islands (because from around the upper Miocene/Pliocene onwards, this has been attributed to the onset of NE directed subduction; Section 2.5.5). The evidence for the continued North Solomon/Vitiaz trench activity includes the emplacement of the mid Miocene to Pliocene Maetambe Volcanics and Pliocene Kömboro Volcanics on Choiseul (Ridgeway et al., 1987). These Choiseul volcanic rocks are situated at a more typical arc-trench distance relative to the North Solomon/Vitiaz Trench, especially as the present day volcanism related to the NE dipping subduction zone is situated anomalously close to the New Britain/San Cristobal Trench (Section 2.7; Johnson et al., 1987). Evidence that the OJP may have subducted recently along the North Solomon/Vitiaz Trench comes from the presence of a seismically active SW dipping hanging slab (Figure 2.4) under the region of Santa Isabel (Section 2.6; Cooper and Taylor, 1987; Taylor and Exon, 1987). In addition, evidence that the Vitiaz/Cape Johnson Trench to the east of the Solomon Islands has been active very recently comes from SOPAC maps which clearly show a linear array of volcanic centres, the rejuvenated Vitiaz Magmatic Zone (Kroenke, 1995), due south of the trench (Figure 4.11; Section 4.7).

2.5.5 Stage 2 Arc development

At about 12 m.y. another change in Indo-Australia Plate motion occurred (Yan, 1988; Epp, 1984; Kroenke et al., 1993) and subduction of the Indo-Australia Plate beneath the Pacific Plate commenced (Figure 2.3d), triggering the formation of the New Hebrides and San Cristobal/New Britain trench system (South Solomons trench system). Although the NE directed (Stage 2) subduction is thought to have begun at approximately 12 Ma, the oldest volcanic rocks related to this phase in the Solomon Islands are dated at 6.4±1.9 Ma in East Guadalcanal (Table 2.1; Pound, 1986). Petterson et al. (in press) suggest that subduction along the South Solomon Trench system began in the east near Fiji at about 12 Ma and did not reach the Solomon Islands until ≈8-7 m.y.. This active NE-dipping subduction zone is responsible for the formation of the upper Miocene to Recent Volcanic Province (Figure 2.2) as defined by Coleman (1965). Included in Coleman's Volcanic Province are more than thirty well-preserved Recent volcanoes (Richards et al., 1966; Dunkley, 1983) though only a few are presently active. The majority of these are situated among the New Georgia Group, but Stage 2 volcanism can be found in Guadalcanal, Makira, Savo, the Russell Islands and the Shortland Islands, Choiseul, and the Florida Islands (Richards et al., 1966; Pound 1986; Coulson and Vedder, 1986; Ridgeway et al., 1987; Chivas, 1981; Dunkley, 1983; Petterson, 1995b).

2.5.6 Obduction of the Malaita Anticlinorium

Around 5-6 m.y. the deposition of shallow marine facies (Suafa Limestone Formation, Section 3.6.4) occurred on the SW margin of the OJP, indicating the onset of emergence of the Malaita Anticlinorium (Figure 2.3e). The Malaita Anticlinorium was first demonstrated by Kroenke (1972) to be a seismic continuation of the SW margin of the OJP, and later geochemically confirmed to indeed be an obducted portion of the OJP (Mahoney and Spencer, 1991) despite conflicting palaeomagnetic data (Musgrave, 1990). The formation and eventual subaerial emergence of the Malaita Anticlinorium in several of the Solomon Islands was the result of an increase in the coupling between the OJP and Solomon Block bracketed at 2-4 m.y., a 'hard docking event' (Petterson et al., in press). This maximum deformation phase on Malaita occurred during the mid Pliocene, causing ~30% shortening. This deformation is stratigraphically constrained by the Suafa Limestone Formation, the youngest unit to be affected by this deformation event, which is dated around upper Miocene to lower/middle Pliocene (Section 3.6.5). In turn the deformed Suafa Limestone Formation is unconformably overlain by localised upper Pliocene to Pleistocene units which remain undeformed (Petterson, 1995b; Petterson et al., in press). Proposed obduction mechanisms for the Malaita Anticlinorium (e.g. Kroenke, 1974, 1986; Petterson et al., in press) are developed in Chapter 4. Although the OJP makes up the basement in several Solomon Islands, the best exposure, thickest basement pile and clearest insight into the obduction of the anticlinorium occurs on the island of Malaita.

2.6 Seismic studies in the Solomon Islands

Today the Solomon Islands are a highly active seismic area, including both deep and shallow activity. Cooper and Taylor (1987) were the first to show with the use of seismic cross sections conclusive proof of the current NE subduction (Figure 2.4). This subducting slab is dipping almost vertically to depths of 300 km both to the north of Bougainville and just to the south of Malaita (Figure 2.4) (Cooper and Taylor, 1987; Taylor and Exon, 1987). Deeper activity is also picked up below 400 km indicating possible detached slab relics from an old SW dipping plate. However, cross section D (Figure 2.4) from directly below Santa Isabel appears to show a seismically active SW dipping slab down to depths of about 200 km. Whether this is a rejuvenation of a portion of the original (Vitiaz Arc) SW-dipping slab that didn't detach, or a newly



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Figure 2.4

developed subduction zone, possibly even a subducting portion of the OJP, is unclear, but it does indicate the North Solomon Trench is not, as originally thought, an inactive, relict trench.

A shallow region located beneath the New Georgia Group shows a small degree of seismic activity to depths of <80 km. In contrast to the dense seismic activity to the north of Choiseul and the south of Malaita, the earthquake activity in the New Georgia Group region is sparse (Figure 2.4). The absence of deep quakes and the shallowing of the two trench systems to less then 3000 m (Figure 2.2) has been attributed to the subduction of the thermally buoyant Woodlark Ridge (Cooper and Taylor, 1987; Taylor and Exon, 1987). The general paucity of shallow seismic activity may be because the hot, thin lithosphere of the active spreading ridge deforms (plastically?) without generating many earthquakes (Taylor and Exon, 1987). Cooper and Taylor (1987) believe that the NE subducting young lithosphere may be responsible for pushing the old SW-dipping slab to greater depths.

2.7 Present day anomalous volcanic activity in the Solomon Islands

Present day tectono-magmatic activity in the Solomon Islands is unusual in many respects, not least in the absence of widespread and extensive volcanism expected at typical oceanic island arcs. Recent volcanic activity seen in the Solomon Islands is situated mainly within the New Georgia Group and differs from magmatism typically associated with island arcs. The geochemistry of the New Georgia Group forearc ranges from basalts through to andesites (Stanton and Bell, 1969; Perfit et al., 1987) with an unusually large percentage of picrites which is a rare feature of most arc lava sequences. These picrites, first noted by Stanton and Bell (1969), have affinities more with oceanic island type basalts such as Hawaii rather than island arcs. Several comprehensive geochemical and petrological studies for the currently active volcanic sites around the Woodlark Basin and New Georgia Group have been published by Johnson et al. (1987), Perfit et al. (1987) and Staudigel (1987).

Magmatism since the late Miocene/Pliocene and up to the present day, associated with the NE dipping subduction zone, is highly unusual in its close proximity to the San Cristobal Trench. Normally arc magmatism can be expected to be found some 150-200 km from the trench. In the New Georgia Group the activity is generally much closer, averaging 30 to 80 km, with the active submarine volcano of Kavachi situated only 30 km from the trench (Johnson and Tusi, 1987). Volcanic edifices along the Simbo and Ghizo Ridge on the downgoing Australia-India Plate are believed to be unique features (Johnson et al., 1987; Crook and Taylor, 1994). These peculiar volcanic features are attributed to the local subduction of the 5 m.y. old Woodlark Basin spreading ridge (Figure 2.1) (Milsom; 1970; Johnson et al., 1987; Perfit et al., 1987; Taylor and Exon, 1987; Crook and Taylor, 1994).

An actively subducting spreading ridge will affect the thermal structure of the subducting plate and its overlying mantle wedge. This in turn will affect the depth of melting, allowing shallow near-trench volcanism, and the angle of the subduction will be highly acute because of thermal buoyancy (Figure 2.4, cross section C). The volcanic centres on the downgoing NE-dipping Australia-India Plate, i.e. Kana Keoki on the Ghizo Ridge, have been attributed to magma being able to migrate up conduits towards the Woodlark Ridge from under the main island arc in the region of the New Georgia Group to south of the trench (Johnson et al., 1987) because of this shallow melting.

Away from the subducting Woodlark Ridge there is little Recent arc volcanism. The collision of the OJP with the subduction zone may have inhibited usual arc volcanism. The descending Australia-India slab away from the Woodlark Ridge is in contact with the cool, thick lithosphere of the OJP (Figure 3d and e) instead of the asthenospheric wedge necessary to generate arc volcanism (Coleman and Kroenke, 1981). Savo is one of the few active volcanic centres outside the New Georgia Group Islands (Figure 2.2). Ramsay (1981) attributes the volcanic activity of Savo to its position at the eastern end of the Mborukua Lineament. The Mborukua Lineament is a major structural feature on the north side of the trench which runs parallel to the trends of the ridge segments of the Woodlark Basin spreading ridges, along which Kavachi lies at the westernmost end (Ramsay, 1981). However, with Savo lying 80-110 km from the San Cristobal Trench (a perfectly respectable trench-arc distance but unusual in the case of the Solomon Islands with their near-trench activity), the suggestion by Taylor and Exon (1987) that Savo is due to the currently active SW-dipping slab (Figure 2.4, cross section D) cannot be dismissed.

2.8 Development of a new geological framework for the Solomon Islands

New field, geochemical and age data from several of the main Solomon Islands acquired

during the latest 1990-1995 field seasons by the Solomon Islands Ministry of Natural Resources (MNR) has resulted in the development of a new framework model (Petterson et al., 1995a). The basis of the new framework has been developed from Coleman's original three provinces but it has been necessary to extend and subdivide the islands into five groups. For example, islands such as Choiseul and Guadalcanal were grouped with the Florida Islands in Coleman's Central Province. All of these islands were affected by both Stage 1 and 2 arc volcanism, but the absence of any obvious Cretaceous N-type MORB basement, as found in Choiseul and Guadalcanal in the Florida Islands has led them to being classified separately in the new model (Figure 2.5).

Although the new model largely regroups islands on presence/absence of Stage 1/2 arc volcanism or the presence of a Cretaceous MORB basement (Petterson et al., 1995a) it is the new data on the island of Makira (Appendix Div; this study), causing it to be grouped separately, that has caused most interest. Makira, originally thought by Coleman to belong solely to the Central Province, does indeed display Stage 2 arc volcanism. But Makira, its pelagic cover sequence mostly removed, also shows an uplifted basaltic basement (hence it was later interpreted by Coleman as belonging in the Pacific Province). This basaltic basement shows frequent, interbedded limestones and cherts, somewhat different to the Malaitan basement lava pile with its paucity of intercalated sediments, indicating significant hiatuses in the eruption of the Makira lava sequence. Geochemically the Makira basement lavas can be subdivided into N-type MORB and OJP type basalts (Figure 2.6). These OJP-type basalts from Makira are geochemically indistinguishable from those drilled from the main body of the plateau and seen obducted on Malaita and Santa Isabel. Decisions whether to include Makira as a portion of the OJP has been complicated by three new ⁴⁰Ar-³⁹Ar dates of 63.0±0.5 m.y. and 33.9±0.7 m.y. for the OJP type basalts and 35.1±1.1 m.y. for the N-type MORB samples (Petterson et al., 1995c). These ages are very different to the 122 m.y. and 90 m.y. ages discovered to date drilled on the main plateau body (Mahoney et al., 1993) and from Malaita and Santa Isabel (Mahoney and Spencer, 1991, Tejada et al., 1996).

The unexpected youth of the Makira lavas, along with the absence of any samples from the eastern lobe of the OJP, shows that a considerable part of the OJP story still remains untold. In addition the 35.1 ± 1.1 m.y. MORB and 33.9 ± 0.7 m.y. OJP-type basalts appear to be



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interbedded on Makira. This must have required a complex source region or magmatic plumbing system, allowing OJP plume related lavas to intercalate but not eradicate the geochemical signature of the MORB source. Makira therefore remains something of a puzzle.



Figure 2.6

2.9 Proof that Malaita is an obducted fragment of the Ontong Java Plateau

In the division of the Solomon Islands into 3 provinces by Coleman (1965), Malaita was included in the Pacific Province. Malaita is very different from the metamorphosed and faulted volcanic rocks of the Central Province and the more recently active Pliocene to Holocene sequences which Coleman included in the Volcanic Province (Coleman, 1965). It was the Pacific

Province's unmetamorphosed basaltic basement and overlying series of deep water pelagic ooze containing Upper Cretaceous foraminifera (Coleman, 1966) which led to the earliest theories that the Pacific Province represented an obducted portion of the southwestern edge of the Ontong Java Plateau.

There have been suggestions that the Pacific Province may not represent part of the OJP. Musgrave (1990), using palaeomagnetic data, suggested that the Pacific Province was an unrelated relict fragment of the Australia-India plate. We can include the following (along with an ever more discriminating collection of geochemical and isotopic data) to suggest Musgrave is wrong: i) the thickness of the oceanic crust in the region of Malaita may be as much as ~32 km (Hussong et al., 1979); ii) similar overlying sedimentary and fauna sequences seen on Malaita to those drilled on the OJP (Berger et al., 1992); iii) the similarity in composition of the rocks between Leg 130 and Malaita (Mahoney et al., 1973); iv) the seismic continuity between the OJP and the Malaita Anticlinorium (e.g. Kroenke, 1972); v) the presence of alnoite intrusions in Malaita.

The Malaita Alnoites, dated at 34 m.y. (Nixon, 1980; Neal and Davidson, 1989), are an ultramafic, kimberlite-type intrusion of deep origin. The alnoites intrude the Malaitan lava pile, cropping out in the Babaru'u area in the Faufaumela Basin, and the appearance of pipe like intrusions on seismic reflection profiles (Kroenke, 1974) on the main plateau body itself indicates the intrusions are not local only to Malaita. Considering that the OJP was some 2000 km east of its present day position (Nixon and Coleman, 1978) when the alnoites were intruded at 34 m.y., the likelihood of Malaita originating from the Australia-India plate therefore seems unlikely.

In addition to the stratigraphic evidence, the Malaitan Basalts display major and trace element, REE and isotope geochemistry (Mahoney and Spencer, 1991) overlapping with data from samples drilled by DSDP Site 289 (Stoeser, 1975) and ODP Leg 130, Sites 807 and 803 (Mahoney et al., 1993). These samples were collected from the main body of the OJP over 1600 km away from Malaita, and all of the isotopic and incompatible element ratios from the main plateau and from Malaita, fall within a very limited and distinctive range (Chapter 7). Such a large and conclusive data set leaves us in little doubt that Malaita represents an obducted portion of the OJP's southwestern margin and therefore Malaita is an ideal location in which to carry out sampling and fieldwork studies of a major plateau.

Chapter 3 The Geology of Malaita

3.1 Introduction

Malaita offers a unique opportunity to study and sample strategic sections of an oceanic plateau to far greater depths than previously drilled by any ODP/DSDP drilling operation. For these reasons two teams, one from Leicester (T.L. Babbs and A.D. Saunders) and the second from the USA (J.J. Mahoney and C.R. Neal), carried out extensive sampling on Malaita during the summer of 1993. Throughout 1993, mapping of North-Central Malaita was being undertaken by M. Petterson and B. McGrail working for the Solomon Islands Ministry of Natural Resources (MNR). This was part of a major geological survey undertaken in central and northern Malaita between 1990 and 1995 by the MNR which has resulted in the production of one 1:100,000 and nine 1:50,000 geological maps plus two geological/economic memoirs (Petterson, 1995b; Mahoa and Petterson, 1995). This work complements the 1:50,000 mapping carried out in southern and Small Malaita between Oct. 1970 and Apr. 1974 by C.C. Hughes and G.W. Turner (1976).

It requires an astonishingly large amount of paperwork and background work to organise a 3 week field expedition in a tribally segregated, thickly jungled remote island such as Malaita. Knowledge of the tribal areas to be visited is of fundamental importance when setting up meetings with the chiefs to gain essential land access permission, not always easy as old misconceptions of white people stealing from their land are often well ingrained. It was only the co-operation of and collaboration with the MNR, use of its resources, workers and field experience that enabled us to carry out a field season on the island of Malaita.

3.2 Topography, vegetation and climate

Malaita is one of the largest islands in the Solomon Islands, being nearly 200 km long and 40 km wide. The interior of Malaita is dominated by sharp ridges running parallel to the coast. A series of periclines in the underlying rocks (Figure 3.1) is the dominant cause of such topographic





features. The highest point on the island (1303 m) is Mt. Kolovrat. The vegetation of Malaita is dense rainforest often grading to mangrove swamps around coastal areas and coral reefs surround a large percentage of the coastline. The climate, typical of the Solomon Islands, is sub-equatorial with rainfall throughout the year, temperatures high and conditions humid. Highest humidity and rainfall occurs in the summer months of mid December to mid May, while the average daily temperature of 27 °C varies little seasonally (±1°C). Many of the fast flowing rivers acted as a natural barrier to cross-island movement, hence the survival of many different tribal customs and languages.

The dense rainforest coverage caused difficulties not only in the lack of exposure but also in the negotiation of the jungle terrain. Rivers gave best access into the interior with the occasional bush track (climb/scramble/hang on or grim death) around the deep gorges and waterfalls. Rivers had the advantage of being marked accurately on the topographic maps, so outcrop position was established by pace and compassing method. This is easier and more precise than the seemingly impossible task of mapping in a densely vegetated rainforest.

3.3 Field Logistics

The Leicester field trip to Malaita was carried out between the 4th to the 21st of August, 1993, totalling 18 days. The aims of the trip were simple; to study the Malaitan Volcanic Group and the overlying sedimentary sequence, and to sample extensively the exposed lava pile of the obducted plateau to greater depths then previously drilled by DSDP, Leg 30 and ODP, Leg 130.

In total four rivers were sampled, the Kwaleunga, Ala'olo and Singgalo on the western limb of the Kwaio Anticline and the Ngwanaa River on the eastern limb of the Fateleka Anticline (Figure 3.3). The stratigraphic sequences exposed in these riverbeds enabled us to carry out observations and sampling to a significantly greater depth of penetration into the lava sequence than the 149 m reached at ODP, Leg 130, Site 807.

A suite of 91 samples was collected from the four rivers sections in northern and central Malaita. We further supplemented our collection by a similar number of samples from the Ministry of Natural Resources (MNR) rock collection in Guadalcanal. These included samples collected from other rivers sections sampled in north and central Malaita by the MNR during their 1993

mapping season, plus a few highly altered samples collected by Hughes and Turner (1976) in their expeditions to south and Small Malaita. Therefore we returned to the UK with a good wide collection totalling \approx 200 samples.

Since the completion of the MNR's mapping of northern Malaita, work has begun on the islands of Makira (San Cristobal) and Ulawa. It is unclear where the SW margin of the OJP lies in relation to these islands. Coleman (1965) includes Ulawa in the Pacific Province, while Makira's origins are less clear (Section 2.8). Analysis of basalt samples from these islands was undertaken in Leicester to help constrain the boundaries of the Pacific Province and exposure of the OJP throughout the southern most Solomon Islands.

3.4 Nomenclature changes to the stratigraphy of Malaita

The geology of Malaita consists of a Cretaceous "basement" of oceanic origin basalts exposed in a series of *en-echelon* periclines, overlain by an approximately 1-3 km thick sequence of Cretaceous to Recent sediments (Figure 3.2). Malaita has been mapped, in parts, by several independent geologists (e.g. Rickwood, 1957; Hughes and Turner, 1976; Barron, 1993; Petterson, 1995b) and as a result various names have become attached to the same stratigraphic units (as summarised in Table 3.1). In this thesis we use the nomenclature adopted by Petterson (1995b; Petterson et al., in press) and his co-workers from their recent mapping of northern and central Malaita.

Rickwood (1957), after his mapping of Northern Malaita, originally named the exposed basaltic basement the Alite Volcanics, subdividing them into the Fiu Lavas and the Fo'ondo Clastics. Hughes and Turner (1976) found it necessary to use a separate nomenclature when they mapped southern and Small Malaita because central Malaita remained unmapped, therefore the tracing of continued stratigraphic units southwards was not possible. Revising the stratigraphic record, they renamed the Alite Volcanics the Malaita Volcanics (Hughes and Turner, 1976), and renamed the Fiu Lavas and the Fo'ondo Clastics as the 'Older Basalts'. In Southern Malaita, this comprises mostly the Fiu Lavas, because the Fo'ondo Clastics die out southwards. In this account the Cretaceous basement lava sequence is referred to simply as the Malaita Volcanic Group (MVG) after Petterson (1995b; Petterson et al., in press).

Hughes and Turner (1976) were the first to notice the occurrence of a suite of younger

lavas cropping out within the cover sediments in South Malaita and on Small Malaita (Maramasike), and called them the Younger Basalts. Recent mapping has discovered a similar group of younger basalts cropping out in Northern Malaita. Petterson (1995b) groups them, both north and south groups, as the Maramasike Volcanic Formation because of their identical stratigraphic position. Because of their geochemical differences we will further divide them for geochemical purposes into the 'Southern Younger Volcanics' (SYV) and the 'North Malaita Alkali Suite' (NMAS) in accordance with Tejada et al. (1996).

Rickwood 1957	Hughes and Turner 1976	Barron 1993	Petterson 1995 (b)
North Malaita	South and Small Malaita	North Malaita	North and Central Malaita
Fo'ondo Clastics/ Fiu Lavas <i>(Alite Volcanics)</i>	Older Basalts (Malaita Volcanics)	Malaita Volcanic Group	Malaita Volcanic Group (MVG)
Kware Mudstone (Malaita Group)	Kwara'ae Mudstone (Malaita Sediments)	Kwara'ae Mudstone	Kwara'ae Mudstone Formation
Alite Limestone (Malaita Group)	Apuloto Limestone/ Are'are Limestone (Malaita Sediments)	Alite Limestone	Alite Limestone Formation
Not observed	Younger Basalts (Malaita Volcanics)	Malaita Volcanic Group	Maramasike Volcanic Formation *
Suaba Chalk (<i>Malaita Group)</i>	Haruta Calcisiltites/ Are'are Limestone (Malaita Sediments)	Suafa Calcisiltites	Haruta Limestone Formation
Suaba Chalk (Malaita Group)	Hada Calcisiltites/ Are'are Limestone (Malaita Sediments)	Suafa Calcisiltites	Suafa Limestone Formation

* The Maramasike Volcanic Formation of Petterson (1995b) includes both north and south alkali basalt suites, for geochemical reasons these are further divided into the Southern Younger Volcanics and the North Malaita Alkali Suite (Tejada et al.,1996). (Any geological divisions made by the authors are shown in brackets).

 Table 3.1. Nomenclature changes in stratigraphic units from Malaita

The older lavas of north of Malaita are conformably overlain by the Kwara'ae Mudstone

Formation. This formation dies out southwards, and is not found on Maramasike (Hughes and

Turner, 1976). The Kwara'ae Mudstone Formation is conformably overlain by a 3-unit limestone sequence. The lower unit is the Alite Limestone Formation (Rickwood, 1957), the middle is the Haruta Limestone Formation, and the topmost is the Suafa Limestone Formation. The dividing line between the Haruta and Suafa Limestone units varies widely between different authors. Large portions of the upper Haruta Limestone Formation were classified as Suaba Chalk (Rickwood, 1957; later corrected to Suafa Chalk by Hackman, 1968), and Barron's (1993) Suafa Calcisiltite Formation also includes large portions of the uppermost Haruta Limestone Formation (as defined by Petterson, 1995b) only really includes the uppermost parts of Rickwood's (1957) original Suaba/Suafa Chalk formation, as most belongs in the Haruta Limestone; and Hughes and Turner (1976) called the equivalent of the Suafa Limestone the Hada Calcisilitie unit. Unconformably overlying the Suafa Limestone Formation is a sequence of thin localised mid Pliocene to Recent limestone/conglomerate formations; Lau Limestone, Tomba Limestone, Hauhui Conglomerate and the Rokera Limestone.

3.5 The Malaita Volcanic Group (MVG)

3.5.1 Stratigraphic relationships, exposure and age

The Malaita Volcanic Group (MVG) makes up the basement of the island. Consisting of a basaltic lava sequence that is oceanic in origin, the MVG is exposed within the cores of *enechelon* periclines (Figures 3.1 and 3.3) and is overlain by an approximately 2-3 km thick sedimentary sequence which is often chaotically deformed near its contact with the basement lavas. In north Malaita the MVG is conformably overlain by the Kwara'ae Mudstone Formation but southwards this unit pinches out and the MVG is directly overlain by the Alite Limestone Formation (Hughes and Turner, 1976; 1977). The thickness of the MVG is unclear as the lower contact is not exposed but estimates from river sections (Section 4.5.2) and calculated detachment depths (Petterson et al., in press) allow an estimate of the maximum thickness to be between 3-4 km (Petterson, 1995b).

The MVG is exposed in the cores of inliers (or periclinal structures) three of which were mapped by Petterson (1995b): the Kwara'ae, Fateleka and the biggest in central Malaita, the Kwaio Inlier (Figure 3.1). The southern end of the Kwaio Inlier (Dorio) was mapped by Hughes and Turner (1976), who also mapped two other small inliers, one in southernmost









Malaita ('Are'are) and the second on the island of Maramasike (Figure 3.1).

The MVG has been dated at 121-125 m.y. using ⁴⁰Ar-³⁹Ar geochronology methods (Tejada et al., 1996; Duncan et al, unpublished data). This corresponds to the major plateau building event of 122.3±1.0 m.y. as drilled on the OJP by ODP, Leg 130, Sites 807 (Mahoney et al., 1993) and DSDP, Leg 30, Site 289 (Stoeser, 1975). To date no 90 m.y. aged basalts have been dated from Malaita associated with the second plateau building eruptions, as have been found on Santa Isabel and recovered by ODP, Leg 130, Site 803 (Mahoney et al., 1996).

3.5.2 Lithological description

The MVG consists of a basaltic extrusive sequence including both pillowed and nonpillowed, sometimes poorly jointed, sheets. These sheets vary from 40 cm up to 70 m in thickness and appear to be only rarely interrupted by intercalated sediments (<<1%). Exposure along rivers often reveals impressive lava/sheet sequences up to 100 m in thickness. Hughes and Turner (1976) recorded an 85 m thickness for an individual pillow sheet in the Unua River from Southern Malaita and a series of massive sheets in the Hito River in Northern 'Are'are reaching 456 m in total thickness, initially giving the appearance of an individual sheet. However, sequences of pillowed and non-pillowed (massive or columnar jointed) alternating layers is more typical of the MVG.

Some of the massive sheets show basalts grading into dolerites in the centre of the units. Basalts and dolerites alike are tholeiitic in nature (Chapter 7) and the bulk of their mineralogy is made up of augite and plagioclase with rare fresh olivine (Chapter 5). The ophitic/subophitic nature of the coarser grained dolerites occasionally gives the MVG a mottled appearance.

Very few dykes were observed and to date no sheeted dyke complexes have been found on Malaita. Petterson (1995b) mapped a small number of dykes on the southeastern side of the Fateleka Inlier which are roughly N-S or E-W. These dykes mostly intrude another MVG intrusive phase, one of a group of gabbro bodies which are restricted to the Fateleka Inlier. Hughes and Turner (1977) suggest that single flows cropping out within the rivers are difficult to distinguish from dyke or sill intrusions. They also noted it is often difficult to ascertain whether the massive units are flows or sills (hence they are termed sheets).

Pillow lavas make up to 70% of the basement units (Plate 3.1). The fresher, thicker



Plate 3.1. Pillow lavas make up approximately 70% of the MVG exposure (Ala'olo River).



Plate 3.2. Poorly columnar jointed thick (\geq 10 m) sheet (Singgalo River).



Plate 3.3. Rare sediments between an upper, poorly jointed sheet (as in Plate 3.2) and lower pillowed sheet. Flat surface of pillows below imply a possible tectonic surface (Singgalo River).



Plate 3.4. Pillow striations (the top of the Kwaleunga River)

pillow units were often spectacular sights, with beautifully preserved chilled glassy rims, radial and concentric jointing and the pillow interstices not always infilled. Pillow morphology was at times sharp providing excellent wayup criteria.

Amygdales, though abundant (30-40%) in the outer zones of some pillows, are rare/ absent in massive flows. Amygdales rarely exceed 3 mm in diameter and are often partially or completely filled with secondary minerals, e.g. calcite, zeolite and chlorite. In some large pillows, pipe vesicles (1 cm wide by 10 cm long) caused by the upward movement of gas were present. Pillow striations (Plate 3.4) were observed in the deepest regions of the Kwaleunga River, showing the effects of extrusion of a fresh pillow through a fractured skin.

The non-pillowed sheets are sometimes columnar jointed (Plate 3.2), structureless/massive or grading into both. Flow banding is present periodically as is the exposure of blocky flow tops. Overlying sheets are occasionally seen to have chilled against the lower, cooler sheet top and the moulding of an upper massive sheet over the convex form of the underlying pillows is a clear indication of a successive extrusion for the eruption of the MVG (Hughes and Turner, 1976). Some of the massive sheets are seen to grade into pillow flows suggesting that this may be the common termination of massive flow units.

The lava sequences are only interrupted by minor (less than 1%) intercalated sediments. When present these occasional hiatuses are fine grained chert, soft mudstone or rarer calcilutite, and only 1-2 cm in thickness (Plate 3.3). This lack of intercalated sediments and their highly siliceous nature reveals a lot about the eruption environment of the MVG (Section 3.5.5).

Preservation within the MVG lava pile varies and the alteration of the MVG is discussed in Chapter 6. The colour of the basalts varies, the more altered basalts appearing greenish due to a higher percentage of smectite and secondary clay minerals. Often seen on joint surfaces and sometimes in the filled amygdales was chalcopyrite, along with quartz veins and centres. Occasionally seen were large white/pink zeolites (XRD analysis showing these zeolites to be a mixture of thomsonite and tetranatrolite, Section 6.2) with radiating needles up to 3 cm long.

3.5.3 Fateleka Gabbro Intrusions

Approximately five intrusive gabbro bodies are found in the Fateleka Inlier, ranging from approximately 1 km to 7 km in diameter (Petterson, 1995b). The gabbro body exposed along the Ngwanaa River that we observed shows multiple intrusive phases (Plate 3.5). The original



Plate 3.5. Multiple intrusive phase gabbros (Ngwanaa River).



Plate 3.6. Basalt containing feldspar megacrysts.

gabbro intrusion has been cut by a later stage almost pegmatitic gabbro with feldspar crystals >1 cm in length and 0.5 - 2.0 cm diameter clinopyroxenes which increase in abundance towards the centre of the intrusive body. With the exception of a small isolated gabbro outcrop on Small Malaita (Hughes and Turner, 1976) these gabbro bodies appear to be restricted to the Fateleka Inlier, though why is unclear. The estimates of thickness of basalt folded in the Fateleka Inlier is about 2 km, far less than the estimates for the Kwaio Inlier (where 3.5-4 km of basalt have been folded (Petterson et al., in press)), clearly indicating that depth is not the controlling factor for the emplacement of the Fateleka gabbro bodies but is more likely to be a regional effect.

3.5.4 Spotty Rock

Hughes and Turner (1976) were the first to described a "peculiar 'spheruloid texture' being present on an otherwise massive outcrop of basalt". Often found within the river float, and only rarely in situ, this striking facies has been termed the 'orbicular basalt' or 'spheruloidial basalt' by Petterson (1995b) though known more affectionately as the 'spotty rock'. (In this study we prefer to use the term megacrysts as spherulites implies a radial texture). The spotty rock consists of a dolerite/basalt host rock with large (commonly up to 5 cm in diameter) xenoliths of gabbros and monomineralic plagioclase (Plate 3.6) and pyroxene megacrysts.

The host basalt/dolerite is geochemically and mineralogically identical to the MVG lavas and the plagioclase megacrysts (An₈₈-An₈₃ from electron microprobe analysis) could have possibly precipitated from the same primary lavas as the MVG and therefore they are not considered to be xenocrysts (Section 5.3.2). Hughes and Turner (1976) suggest that the megacrysts are the result a late stage cooling event. We strongly disagree with such interpretations and suggest that the megacrysts were grown while residing in a magma chamber. Evidence of regular, concentric zoning in the plagioclase megacrysts, with little variation in An composition (<5%) across 6 cm scale crystals suggests a long period of growth in an open system magma chamber (Section 5.4) before being caught up in an erupting magma.

3.5.5 Eruption Environment of the Malaita Volcanic Group

Virtually no pyroclastic material has been observed in the MVG, apart from a small volume of pillow breccia belonging to the Fo'ondo Clastics by Rickwood (1957). The paucity of intercalated sedimentary layers, little pyroclastic evidence and thick massive sheets up to 70 m thick (which may be representative of magma ponding) all suggest voluminous 'flood' magmatism occurred. Drilling closer to the crest of the OJP has revealed similar thick lava sheets with few sedimentary horizons (Mahoney et al., 1993) supporting the theory that lava eruption was rapid, widespread, voluminous in nature and submarine.

The overlying water column into the MVG erupted is believed to be of considerable depth (Hughes and Turner, 1976; Saunders et al., 1993). Vesicles in basalts allow a crude estimate of the seawater pressure and depth at time of eruption. Seawater pressure at depths in excess of approximately 2000 m is usually sufficient to prevent the exsolution of dissolved gases. Pillow vesicularity varies from \approx 10% to 40% when erupted at depths of less than 500 m and 5% at 1000 m (McBirney, 1963; Thorpe and Brown, 1985). However, no detailed study of the size, distribution and state and amygdales was carried out so only the most elementary of conclusions can be drawn, that is the depth of eruption is likely to be greater than 1000 m as the pillow vesicularity is < 5%. Hughes and Turner (1976) observed in the south that the younger basalts were more vesicular than the older series in the north, implying an extrusion depth greater for the older MVG basalts than that of the southern younger basalts.

The eruption of the MVG is therefore considered to have been into water of considerable depth, with evidence pointing to below the carbonate compensation depth (CCD) (Hughes and Turner, 1976), which is thought to have been around 2.8-3.0 km deep during the Aptian (Arthur et al., 1985). This evidence includes (i) pillow basalts (with less than 5% vesicularity) make up a large portion of the basement sequence; (ii) absence of pyroclastics implies eruption sites in deep (>500 m) water, far removed from subaerial vents; (iii) rare sediment intercalations are usually fine grained chert mudstones or calcilulite; (iv) the older basalt sequences are overlain conformably (?) by a highly siliceous mudstone, or deep sea pelagic calcilutite known as the Kwara'ae Mudstone Formation (Section 3.6.1).

The indication that at least part of the plateau was constructed below the CCD raises many questions in light of the association of CFB and oceanic plateau formation with large vigorous plume systems (Coffin and Eldholm, 1993). Such a plume would be expected to elevate a large area of the developing plateau to shallow sea levels, if not above sea level. However, to date no evidence of subaerial eruption of the OJP has been drilled on the main body of the plateau or exposed within the Solomon Islands. ODP Leg 130 found basalts interbedded with limestones at Site 807, suggesting magma deposition at or above the CCD (Mahoney et al., 1993). Radiolarian-enriched interbeds within the overlying sedimentary sequences were found at Sites 807 and 803, deposited near or below the CCD which may have been the result of latter stage subsidence of the plateau. Therefore, it could be that the main body of the OJP did erupt around or slightly above the Aptian CCD as indicated by the limestone interbeds recovered by the ODP, i.e. at shallower depths than seen exposed in Malaita. However, the idea that Malaita may be the deep water flank of a shallower structure (Saunders et al., 1993) is one which needs to be addressed. But it is noteworthy that the crust beneath Malaita is thought still to be in excess of 30 km thick, although how much of this is due to tectonic thickening is unclear.

3.6 Malaita's cover sequence

3.6.1 Kwara'ae Mudstone Formation

The Kwara'ae Mudstone Formation directly overlies the MVG, though it is gradually pinched out southwards on Malaita to the extent that it is absent on Small Malaita, where the MVG is directly overlain by the Alite Limestone Formation (Hughes and Turner, 1976). A maximum thickness of 270 m for the mudstone unit was recorded by Hughes and Turner (1977) and still remains the maximum for the stratigraphic range of 100 - 270 m calculated by Petterson (1995b).

Thin sections of the mudstone show varying percentages of radiolaria present which, to date, have been of no biostratigraphic value and the date of formation of the mudstones is still speculative. The most accurate means of dating the mudstone is by the restricting upper and lower stratigraphic units which sandwich it. The underlying MVG give the Kwara'ae Mudstone Formation a maximum age of 121-124 m.y. (Mahoney et al., 1993; Tejada et al., 1996) or lower Aptian. The overlying Alite limestone Formation has been more successfully dated by its abundant foraminifera as Maastrichtian/Cenomanian.

The banded greenish grey mudstone (Plate 3.7) displays thinly laminated beds and thicker, more massive beds (Hughes and Turner, 1976) causing it to be subdivided into two facies. Facies 1 forms the thicker (10 - 30 cm) hard bands which tend to be massive or finely laminated and are more siliceous than the Facies 2, which is a softer, thinner (averaging about 8 cm) fine grained silt and is more calcareous in composition than Facies 1 (Petterson, 1995b). The mudstone is classified as a highly siliceous ooze, mostly inorganic in origin (Hughes and Turner,



Plate 3.7. Kwara'ae Mudstone Formation showing the two different facies (Ala'olo River).



Plate 3.8. Alite Limestone Formation, resilient chert bands shows folding related to thrusting (Ala'olo River).

1976) with the lack of calcite indicating an origin of formation below the Aptian CCD. The carbonate composition of the mudstones increases up sequence implying later, shallower, depths of deposition with the youngest mudstone units forming at depths at or possibly above the CCD (Hughes and Turner, 1976).

3.6.2 Alite Limestone Formation

Conformably above the Kwara'ae Mudstone Formation lies the Alite Limestone Formation which varies in thickness from 300 to 1000 m across Malaita (Rickwood, 1957; Hughes and Turner, 1976; Petterson, 1995b). The Alite Limestone Formation has been dated by its well-preserved foraminifera by several authors (Van Deventer and Postuma, 1973; Hughes and Turner, 1976; Hine, 1991 and Barron, 1993). The oldest units are Maastrichtian/Cenomanian and younging up to the Middle/Upper Eocene. Also the upper units of the Alite Limestone Formation are intruded by and overlain locally by the Maramasike Volcanics. Both the alkali suites, the Southern Younger Volcanics and Northern Malaita Alkali Suite, which make up the Maramasike Volcanic Formation have been dated by ⁴⁰Ar - ³⁹Ar methods at 44 m.y. (Tejada et al., 1996), a good age constraint marking the end of Alite Limestone Formation deposition on Malaita.

The Alite Limestone Formation is a well-lithified white limestone formation containing thin interbeds and lenses of yellow, pink, grey, red chert bands. At their purest they contain 95% CaCO₃ (Hughes and Turner, 1976) down to \approx 77% (this study). The resistant chert bands stand out well in contrast to the paler limestones clearly showing folding, shear zones and other structural deformations (Plate 3.8). The presence of the Alite Limestone Formation conformably overlying a deep sea ooze such as the Kwara'ae Mudstone Formation is a further indication of a shallowing in the deposition environment to above the CCD. However, the presence of chert bands, which if primary deposits of radiolarian ooze as suggested by Petterson (1995b), could indicate that the lowermost units of the Alite Limestone were also deposited in part below the CCD.

3.6.3 Maramasike Volcanic Formation

In south Malaita and on Maramasike (Small Malaita) a suite of younger, highly vesicular basalts occur overlying and interbedding the hard, porcellanous Alite Limestone Formation (Hughes and Turner, 1976). A similar suite of younger basalts intrude (?) the same sedimentary horizon in northern Malaita (Barron, 1993; Petterson, 1995b). Because of their similar stratigraphic relations they have both been grouped under the same formation, the Maramasike Volcanic Formation by Petterson (1995b) though in this PhD thesis they are subdivided into the Southern Younger Volcanics (SYV) and Northern Malaita Alkali Suite (NMAS) because of their geochemical differences (Section 7.6). The SYV and NMAS are both highly alkaline basaltic suites, very different to the tholeiitic, evolved nature of the Malaita Volcanic Group (Section 7.3); each alkali suite is also isotopically distinctive indicating different source regions (Tejada et al., 1996). Unfortunately during our trip to Malaita we didn't observe either suite of alkali basalts so only a brief description summarised from reports by Hughes and Turner (1976 and 1977) for the SYV, and Barron (1993) for the NMAS follows.

The Southern Younger Volcanics (from Hughes and Turner, 1976 and 1977)

In their mapping of Small Malaita, Hughes and Turner noted an approximately 300 m to 500 m thick sequence of highly vesicular pillowed and non-pillowed basalt sheets which overlie the Alite (Apuloto) Limestone Formation. Thin bands and small patches of the limestone unit are included within portions of the SYV. In southern Are'are the basalts are only 100 m to 150 m thick and die out northwards. Hughes and Turner (1976) observed in the south that the younger basalts were more amygdale enriched than the older series (MVG) in the north, implying a shallower extrusion depth for the SYV than that of the northern MVG. Alternatively the magmas could have been more gas rich. The SYV are commonly brecciated with the occurrence of peperitic sills in the southern Are'are region.

The Northern Malaita Alkali Suite (from Barron, 1993)

The NMAS shows similar lithological features to those described by Hughes and Turner (1976, 1977) for the SYV. They are reported as highly weathered and brecciated lavas consisting of both pillowed and massive sheets of a highly vesicular nature. Amygdales make up 30-60% of the rock volume and are infilled with calcite, chlorite and zeolites. Bedded hyaloclastics and interbedded baked limestones are also recorded and the evidence points to an eruption environment similar to that of the SYV.

Eruption environment of the Maramasike Volcanic Formation

The highly vesicular and commonly brecciated, largely altered nature of both subdivisions of the Maramasike Volcanic Formation makes them distinct from the Malaita Volcanic Group (MVG) with their rare amygdales and reasonably unweathered appearance. The high percentages of amygdales, and their interbedding with limestone units, possibly implies a shallower eruption environment for the Maramasike Volcanic Formation compared to the Cretaceous MVG (Hughes and Turner, 1976; Barron, 1993). The commonly brecciated nature of the Maramasike Volcanic Formation is indicative of an explosive emplacement mechanism (Barron, 1993). The occurrence of peperitic sills in southern Are'are indicate the fluidization and turbulent mixing of wet sediments has occurred on lava intrusion (Kokelaar, 1982).

The NMAS are restricted to the Faufaumela Basin (along with the alnoites) and to the NW tip of Malaita (to the NW of the Toambaita Inlier). The restriction of the alnoites and partial restriction of the NMAS to the Faufaumela basin may be related to Eocene extension (Petterson, 1995b). This is a likely explanation for the reason that the alnoites appear to crop out only in this region, but the volume of the NMAS cropping out in the Faufaumela basin is small in comparison to the NMAS found in the very NW region of Malaita. However, it also doesn't explain why the SYV outcrop in southernmost Malaita and Small Malaita so prolifically unless previous mappers have missed important pre-obduction extension features or they have been completely destroyed or rendered unrecognisable by the later stage obduction deformation.

3.6.4 Haruta Limestone Formation

Conformably overlying the Alite Limestone Formation, in the absence of the localised Maramasike volcanic Formation, is the Haruta Limestone Formation which is in turn overlain by the Suafa Limestone Formation. As with most of Malaita's sedimentary units, the thickness of the Haruta Limestone Formation is variable, between 500 and 1000 m.

The Haruta Limestone Formation has been dated by abundant planktonic foraminifera as being late Eocene/early Oligocene (Hughes and Turner, 1977). The base of the Haruta limestone Formation can be dated by the Maramasike Volcanic Formation, which intrude and overlie the Alite Limestone Formation locally, to <44 m.y. (Tejada et al., 1996).

The Haruta Limestone Formation shows well-lithified pale calcite-rich layers containing dark brown/grey silt/mudstones layers. The mudstone layers often have a straight siltstone base and grade up into clay sediments at the top of the individual mudstone layers. The tops of the mudstone, in contrast to their sharp bases, are clearly bioturbated, implying a deposition by distal ends of turbidity currents.

Thin sections show the mudstone/siltstone bands to contain glass shards, vitric clasts

and plagioclase laths (Petterson, 1995b). These mudstone bands have been separated and geochemical analysis implies a possible arc volcanic source (i.e. Nb troughs on multi-element plots and high Ba/Nb and K/Nb ratios). The mudstones are believed to have formed by erosion of and transport by turbidity currents of volcanoclastic material from some distal source. It is possible that at the time of deposition of the Haruta Limestone the OJP had moved into the vicinity of the Vitiaz Arc.

3.6.5 Suafa Limestone Formation plus local stratigraphic units

The Haruta Limestone Formation is overlain by the Suafa Limestone Formation which is in turn unconformably overlain by several units of local extent. During our field season to Malaita we did not observe or sample the Suafa Limestone Formation or any of the overlying units above the Haruta Limestone Formation, therefore only a brief summary follows.

The Suafa Limestone Formation is poorly exposed and therefore the least studied of all the major formations on Malaita. The Suafa is a soft cream-brown limestone containing abundant pelagic and benthic fauna (Rickwood, 1957; Hughes and Turner, 1976; Petterson, 1995b). Several authors have suggested that the Suafa Limestone Formation is of upper Miocene to lower/middle Pliocene age (Van Deventer and Postuma, 1973; Hughes and Turner, 1976; Hine 1993).

The maximum thickness of the Suafa Limestone Formation is not possible to determine. The top is not found, because the Suafa Limestone Formation is unconformably overlain by a series of local formations. A maximum exposed thickness of the Suafa Limestone Formation on Malaita is approximately 850 m in the Faufaumela Basin; elsewhere the thickness does not exceed 300 m (Rickwood, 1957; Hughes and Turner, 1976; Petterson, 1995b).

The Suafa Limestone Formation represents a major change in depositional environment, from the deep water environments responsible for the formation of the Kwara'ae Mudstone, Alite and Haruta Limestone Formations moving to more shallow waters during the Miocene and eventually to a period of deformation, uplift and erosion responsible for the island-wide mid-Pliocene unconformity.

From the mid Pliocene to the Pleistocene a series of local formations (e.g. Lau Limestone, Tomba Limestone, Hauhui Conglomerate and the Rokera Limestone Formations) were deposited. They vary in thickness from around 100 m for the Tomba Limestone Formation up to 250 m for the Lau Limestone Formation (Rickwood, 1957; Hughes and Turner, 1976; Barron, 1993; Petterson, 1995b). These units cover small regional areas and either unconformably overlie the mid Pliocene Suafa Limestone Formation or, as in the case of portions of the Lau Limestone Formation, the Haruta Limestone Formation.

3.6.6 Silica Budget of the Pacific Ocean during the Eocene

Deposition of the Alite Limestone Formation ended around the time that the emplacement of the Maramasike Volcanic Formation began and was followed by the formation of the Haruta Limestone Formation, a well lithified limestone with mudstone interbeds. However, the chert bands present in the Alite Limestone Formation are notably absent in the Haruta Limestone Formation. The source of seawater silica, enabling the formation of the chert layers in the Alite Limestone Formation, has been suggested as being directly linked to the magmatic activity producing the upper/middle Eocene Maramasike Volcanic Formation (Hughes and Turner, 1976). However, the abundance of chert bands within the Alite Limestone Formation decreases with height, with the uppermost beds showing almost no chert layers and a transitional facies with the overlying Haruta Limestone Formation when the Maramasike Volcanic Formation is absent (Petterson, 1995b). This decrease in the abundance of chert in the Alite Limestone Formation with height is not consistent with the Maramasike Volcanic Formation being the source of silica saturation. Also, if the Maramasike Volcanic Formation was the source of silica allowing chert formation, this would indicate the formation of the chert in the region of Malaita was a local event, not a widespread deposition over the whole plateau (see below).

Sedimentation through the Upper Cretaceous-Eocene (Alite Limestone Formation), in contrast to the Oligocene (i.e. Haruta Formation) shows far greater diversity. Drilled cores from ODP Leg 130 also show diverse lithologies during the Eocene (Berger et al., 1992). Both drilled material recovered from the main body of the OJP and Malaitan sediments reveal that before the upper Eocene, siliceous material was important, defined by interbedded radiolarian bearing chalks, limestones and chert, indicating the silica saturation was not an event local to Malaita. Berger (et al., 1992) state that these deposits are readily seen as strong reflectors (the Ontong Java Series), the top of which indicates a reorganisation of the silica budget of the ocean, marking a transition from a global silica excess to silica deficiency from 40 m.y. to the present day.

3.7 The Malaita alnoites

Ironically, there is a abundance of published literature on the rarest of Malaita's formations, the alnoites. First described by Rickwood (1957), numerous subsequent studies have resulted because of the uniqueness of these small scale intrusive rocks (e.g. Nixon and Coleman, 1978; Nixon and Boyd, 1979; Nixon, 1980; Neal, 1988; Neal and Davidson, 1989). The Malaitan alnoites are also unique as they are the only know alnoites emplaced in an oceanic environment. Alnoites contain xenolithic suites similar to those found in kimberlites set in continental cratonic settings (Nixon and Coleman, 1978) with the presence of garnet Iherzolite mantle xenoliths exceptional considering the oceanic setting of the alnoites (Nixon and Coleman, 1978; Nixon and Neal, 1987).

There are four, possibly five, of these alnoite pipes cropping out in the Faufaumela Basin region of central-northern Malaita (Figure 3.3). The alnoite breccia is blue-green and superficially resembles kimberlite 'blue ground' (Nixon and Coleman, 1978). They are pipe/sill like intrusions explosively emplaced into the Alite Limestone (Neal and Davidson, 1989) and as high as the upper-mid Haruta units, displaying up to 15 cm thick chilled margins (Petterson, 1995b).

The mineralogy of the alnoites includes essential groundmass melilite, extremely large megacrysts of garnet (up to 8.2 kg) and diopside, as well as augite, orthopyroxene (bronzite), ilmenite, phlogopite and minor zircon (Nixon, 1980; Neal and Davidson, 1989). U-Pb dating has been carried out on zircons giving the Malaitan alnoites an age of 34 m.y. (Davis, 1977). They contain well rounded clasts, including abundant country rock xenoliths, ranging in abundance from 5 to 40% (Petterson, 1995b).

3.8 Structural deformation of Malaita - a brief outline

3.8.1 Fold axial trends

The basement and its overlying sediments form *en-echelon* periclines outlining the well developed NW-SE structural grain which dominates the topography and geology of Malaita (Figure 3.1 and 3.3). The structural grain is especially well represented by the fold axis orientation of Malaita (Figure 3.4). The periclines have upright to overturned axial planes with vergence predominately towards the NE (Figure 3.5); the western limbs of the periclines typically have shallow dips to the south west ($\approx 25^{\circ}$) while on the eastern side they tend to dip at a far steeper

angle, 70° to vertical, to overturned, to the north east (Figure 3.5). Notably, trends of the fold axes in some areas, e.g. the Fiu Re-entrant Zone (Figure 3.6), are sigmoidal in nature; this causes the plane of strike of the fold axis to vary for the calculated mode of 128° to approximately 110°. The cause of this refolding is probably major sinistral strike slip between the OJP and the Solomon Plate (Petterson, 1995b; Petterson et al., in press), i.e. the convergence forming the Malaita Anticlinorium is transpressive in nature (Petterson et al., in press). There is a similar, though less developed sigmoidal rotation of fold axes indicating a *dextral* strike slip component on Maramasike.

3.8.2 Thrust faults - Hughes and Turner (1970 - 1974) and Barron (1993)

Faulting within the lava sequence is common both parallel and perpendicular to the main NW-SE structural grain, although thrust faulting appears to be rarer. Hughes and Turner (1976) reported one instance of a thrust fault in their mapping of South and Small Malaita, as well as documenting folding, faulting, jointing and a number of instances of overfolding.

Barron (1993) interprets the structure of northernmost Malaita as being dominated by a major SW-directed thrust complex, the To'ambaita Duplex, plus extensive faulting in the associated footwall, horse and hanging wall sequences. These thrust complexes are shown on Figure 3.7, north of the X-Y line marking the southern boundary of Barron's (1993) mapping area. Portions of Barron's (1993) mapping area was mapped as part of the 1993 to 1995 mapping seasons by M. Petterson and co-workers. M. Petterson (1995b) does not recognise the thrust complex in North Malaita as suggested by Barron, indeed not a single thrust fault is recorded by the group north of the X-Y line (Figure 3.7).

3.8.3 Thrust faults and structural areas - Petterson et al (1992 - 1995)

Petterson (1995b) records a number of thrusts striking parallel to the NW-SE structural grain of Malaita, mostly showing topwards movement (or vergence) towards the NE (Figure 3.7). The main exception is the Kwaio Backthrust Belt which shows a small grouping of SW-directed thrusts. The only other area with SW topwards directed thrusts is the southernmost Auluta Thrust Belt, south of the Fateleka Anticline (Figure 3.6).

The Auluta Thrust Belt and the adjacent Faufaumela Basin is believed to exert important controls on the volcanism (Petterson, 1995b; Petterson et al., in press) being the only known areas where the alnoite pipes crop out and one of only two locations were the Northern Malaita





Chapter 3: The Geology of Malaita

Figure 3.5




Alkali Suite (NMAS) occurs (Section 3.6.3). Petterson (1995b) models the Auluta Thrust Belt and Faufaumela Basin as predominantly extensional features enabling the ascent of the younger alkaline volcanics and alnoite pipes. However, the NMAS occurs in a greater volume to the very NW of the Toambaita Fold Belt (Figures 3.3 and 3.6), furthermore the SYV alkali basalts of a similar age, prolifically outcrop in southernmost Malaita and Small Malaita. Therefore, another deep sourced mechanism controlling the intrusion of Malaita's younger alkali volcanics is required, as well as the Faufaumela Basin formation.

The Kwaio Backthrust Belt on the western limb of the Kwaio Inlier is of major significance to the work currently being carried out at Leicester. Of the river sections we sampled, three were on the western limb of the Kwaio Inlier and are in close proximity to this thrust belt. The tectonics involved in the formation of the Malaita Anticlinorium may have caused imbrication and stacking of NE thrust sheets followed by later stage SW-directed thrusting in the Kwaio Backthrust Belt. This would cause problems when trying to reconstruct depths sampled into the lava pile when working in such a highly thrusted area. The lack of any metamorphism (Section 6.2) indicates that the depth of penetration into the lava sequence was only shallow, i.e. the upper few kilometres as a maximum estimate (Section 4.5).

Chapter 4

Growth and obduction models of oceanic plateaus: the formation of the Malaita Anticlinorium

4.1 Introduction

Oceanic crust - including oceanic plateaus - is tectonically recycled on a time scale of <200 m.y., and as a result no oceanic plateaus can exist in a submarine environment indefinitely. In contrast, the preservation potential of continental flood basalts is not limited by such tectonic recycling processes, although they are subject to normal processes of erosion. Therefore, the preservation of oceanic plateaus such as the OJP is ultimately controlled by their potential to resist subduction. The extent of obduction and/or accretion of oceanic plateaus is governed by the inter-related parameters thickness, age, thermal state of the plateau and the nature of the subduction zone involved (Cloos, 1993). The method of plateau growth (Section 4.2) also plays an important role as it may establish potential décollements. Décollements (Section 4.4) are important for nappe emplacement, create weakness and ductile zones facilitating orogenic processes, and are likely to play an important role in the obduction of oceanic plateaus.

Today, the geological record shows clear preservation of several oceanic plateaus, e.g. the Caribbean Plateau and the Ontong Java Plateau, although the percentage of oceanic plateaus preserved prior to the Cretaceous may be negligible. Therefore, it is highly speculative to predict future events which may affect the OJP, i.e. will it subduct piece-meal (Saunders et al., 1996) or be accreted as an allochthonous terrain (Ben Avraham et al., 1981; Nur and Ben Avraham, 1982)? The future of these Cretaceous plateáus is linked to the debate as to whether Archaean greenstone belt sequences represent the most ancient of oceanic plateaus (Storey et al., 1991). If greenstone belts are imbricated Archaean plateaus this may indicate whether plateau eruption and hence preservation is periodic, an idea suggested by the unrepresentatively large proportion of oceanic plateaus within the Pacific Oceanic and accreted around the Pacific Rim which were erupted during the Cretaceous superchron (Larson, 1991;

Larson and Olson, 1991). Today Archaean greenstone belts make up significant portions of the continental crust, suggesting that plateau eruption and mantle dynamics are directly linked to future crustal growth rates (Stein and Hofmann, 1994). Are the numerous Cretaceous plateaus in today's ocean basins tomorrow's greenstone belts?

The obduction of the Ontong Java Plateau in the Solomon Islands is one of the best examples of an oceanic plateau which has resisted subduction. The initial docking of the Ontong Java Plateau with the North Solomon/Vitiaz Trench, occurring approximately 20 m.y. ago, was a 'soft docking' event. No major compressive deformation occurred until about 6-4 m.y. when the coupling between the Solomon Arc and the OJP intensified and resulted in the formation and eventual subaerial emergence of the Malaita Anticlinorium, i.e. the 'hard docking' event (Section 2.5.6). There is considerable speculation about how a portion of the OJP obducted to form the approximately 150 km wide by 600 km long Malaita Anticlinorium. Recent workers (e.g. Kroenke, 1984; Petterson et al., in press; Mann et al., 1996; and this study) have focused on two main obduction models (i) a flake tectonics model (Section 4.6.1) and (ii) a NE thrusting and imbrication model (Section 4.6.2). The flake tectonics model implies that the OJP has effectively been sliced by its collision with the Solomon arc block, the upper portion scraped off and obducted on to and over the Solomon arc while the lower portions continued to subduct. The second model developed (and favoured) in this study, implies that the leading edge of the OJP, on collision with the trench, caused the onset of NE thrusting, subduction backstep and eventual imbrication of the upper layers of the plateau as a tectonic accretionary prism.

4.2 The formation of large igneous provinces

The mechanisms by which LIPs are emplaced can be generalised into lateral growth (Section 4.2.1) and vertical growth (Section 4.2.2). Lateral growth includes the entrapment of a spreading ridge above a mantle plume such as in Iceland, while vertical growth involves intraplate hotspots, which are responsible for the eruption of Hawaii and continental flood basalts (CFBs). The mechanisms by which LIPs are emplaced, whether via lateral or vertical growth, have important implications not only towards their chemistries, volumes and eruption rates, they also have serious implications for preservation potential, especially in the case of oceanic plateaus because they can dictate the formation of possible décollements

within the plateau's lava pile (Section 4.4).

Which growth model effected the eruption of the OJP is unclear. Geochemically, the lack of heavy rare earth element depletion in the OJP lavas (the OJP lavas typically display flat, near chondritic REE patterns, Section 7.5) implies that melt segregation occurred largely in the spinel lherzolite field. Such shallow melt segregation for the OJP lavas precludes any thick lithospheric cap during their production, unlike in the production of intra-plate hotspot lavas such as in Hawaii. Therefore, it seems the OJP erupted at a spreading ridge and its growth was likely to have been analogous to the present day ridge-centred plume eruption in Iceland.

4.2.1 Lateral growth models

Iceland is the only active subaerial segment of a mid-oceanic ridge and therefore presents geologists with a unique location, not only to investigate oceanic ridge spreading processes, but also to research an actively melting plume and the physical growth of an oceanic plateau. The Icelandic plume originally began melting in a continental intra-plate setting, erupting flood basalts on West Greenland (Clarke and Pederson, 1976; Larsen et al., 1992), eastern Greenland (Larsen and Jakobsdóttir, 1988; Larsen et al., 1994; Fitton et al., 1997) and along the western margin of Europe (Roberts et al., 1984; Eldholm et al., 1987). Eventually, a purely oceanic spreading environment developed as seen today where the actively melting Icelandic plume has captured the Mid-Atlantic Ridge (e.g. Schilling et al., 1982; Gibson et al., 1982; White and McKenzie, 1989; Saunders et al., 1997).

The geology of Iceland is divided into the Quaternary axial rifts or neovolcanic zones, which represent the subaerial portion of the North Atlantic Ridge, and the Tertiary lava pile (Pálmason and Saemundsson, 1974; Saemundsson, 1986). The Quaternary neovolcanic zones are distinguished by two parallel volcanic belts in the southern half of Iceland and a single belt to the north. These volcanic belts are the areas of main crust formation in Iceland and are composed of lavas younger than 0.7 m.y. (i.e. within the Bruhnes chron; Saemundsson, 1978). The geology within the active volcanic belts is complex, with localised volcanic centres related to linear zones of faulting and fissure eruptions (e.g. Krafla). These fissure swarms often have an *en-echelon* arrangement within the axial rift zone (Saemundsson, 1978). The Tertiary lava pile appears very different to the neovolcanic zone, having the appearance of continental flood volcanism with its well bedded, unfaulted, sub-horizontal basaltic sequences well exposed along

fjord coastlines and glacier carved valleys.

Walker (1959, 1960) first suggested, from geological studies of eastern Iceland, that the dipping of the Tertiary lava pile towards the active rift zone was due to sagging caused by lava deposition into the rift zone. Bodvarsson and Walker (1964) expanded these ideas to included Iceland's crustal spreading being strongly influenced by the injection of dykes into Iceland's crust. These initial interpretations were backed up by Saemundsson (1978) and later numerically modelled by Pálmason (1980, 1986). Gibson and Gibbs (1987) suggested that listric normal faulting and lava pile rotation, rather than sagging determined by isostatic conditions, is Iceland's dominant control on volcanic and spreading processes. Dyke emplacement and the associated listric faulting effectively tilts the pre-existing lava pile towards the axial zone (Figure 4.1a and b), achieving the riftward dipping lava sequences (or SDRs). With subsequent eruptions the lavas are rotated down into a deeper position (Figure 4.1c). In contrast Pálmason (1980) does not attribute normal faulting as playing a major part in the dipping of lavas towards the rift zone, focusing more on the addition of material to the rift zone causing sagging of the lava pile and lateral migration of the crust via dyke injection (Figure 4.2). Dyke injection undoubtedly plays an important role in modelling Iceland's growth as shown by the prolific abundance of fissures in the neovolcanic zone, but the actual number of dykes visible in the Icelandic Tertiary lava pile is small, though some dyke swarms are occasionally exposed. This paucity of dykes exposed in Iceland's Tertiary lavas is likely to be the result of rotation of the equivalent to Iceland's Tertiary Neovolcanic zone dykes deep (>6 km?) into Iceland's lava pile away from the axial rift zones.

Past rotation of neovolcanic zone lavas towards the axial rift which caused the synclinal structure of the Iceland crust (Figures 4.1 and 4.2) was undoubtedly the result of an intimate relationship between listric faulting and dyke intrusion (Gibson and Gibbs, 1987). However, this rotation of the pre-existing lavas and dykes relies on a detachment (or décollement) surface below the lava pile. Gibson and Gibbs (1987) suggest that along the magma chamber/crustal boundary, a crystal magma mush zone near to the rifting axis will accommodate the deformation by flowage, and farther away from the rift axis (approximately 5 km below the Tertiary lava pile) a sheared gabbroic layer will form the detachment surface (Figure 4.1c). The formation of such detachment surfaces within oceanic plateaus has wider implications than the function they have served in Iceland's growth models as they may provide separation surfaces during subsequent



plateau - trench collision. Detachment surfaces also form potential décollements playing an important role in plateau obduction (Section 4.4).

Iceland is the only example of a modern day oceanic plateau erupting via a spreading centre; there is no doubt that Iceland-based studies help towards our current understanding of mid-oceanic ridge processes in general. The subaerial exposure of Iceland also helps to highlight the important differences between oceanic plateaus and ophiolites, preserved within the geological record. For example, a lack of dykes is noted within the Icelandic Tertiary lava pile. Obducted oceanic plateaus such as the Ontong Java Plateau and the Caribbean Plateau show a paucity of dykes, even when their deepest portions are exposed as in the Caribbean Plateau (Millward et al., 1984; Burke, 1988; Aitken and Echeverria, 1984). This is in marked contrast to the well-preserved obducted ophiolite suites such as the Troodos and Oman where sheeted dykes are a prominent feature of the geology and ophiolites growth models (e.g., Gass, 1968; Allerton and Vine, 1991; Coleman and Hopson, 1981).

4.2.2 Vertical growth models

There is no doubt that CFBs such as the Siberian Traps and Columbia River Basalt Group add considerable volumes of basaltic rock to continents (Coffin and Eldholm, 1993, 1994), not only by the obvious surface flood volcanism but also by considerable lateral injection at depth into the crust (White et al., 1987) (Figure 4.3). Similarly, submarine intra-plate plumes such as the Hawaiian plume emplace magmas through a thickened lithospheric lid (80 km in thickness (Watson and McKenzie, 1991)) and add substantial volumes to the oceanic crust in the form of seamount chains. The main source of lava eruption in the oceanic intra-plate settings, such as Hawaii, comes from volcanic centres such as the Kilauea and Mauna Loa shield volcanoes and their associated fissure systems. CFBs rely more on fissure eruptions as their main feeder systems, though shield building eruptions do also occur in continental flood basalt provinces (e.g. Mohr and Zanettin, 1988).

The extensive feeder systems responsible for the emplacement of CFBs can often be seen preserved as dyke swarms. The feeder dykes which erupted the Columbia River Basalt Group are all oriented in a similar NNW trend with individual dykes often being correlated to specific flows erupting over the 11 m.y. period of magmatic activity (Hooper et al., 1984; Hooper, 1988). The remarkably constant trend in the dykes over 11 m.y. of their formation suggests that fissure formation was dictated by a regional stress system, and not by local stress fields created by shallow crustal magma reservoirs (Hooper, 1988).

The Deccan Traps show no obvious feeder dyke swarms as seen in the Columbia River Basalt Group (e.g. Auden 1949; Mahoney, 1988). Numerous dykes do exist in the western Deccan Trap flows but they are rare elsewhere. This, along with the thickening of individual flows towards the west, an increase in the number of flows and an increase in the number of laterally restricted flows towards the west, supports evidence of a source area for the Deccan Traps being situated off the west coast (e.g. Raja Rao et al., 1978; Marathe et al., 1981; Mahoney, 1988).

The Deccan Traps and Columbia River Basalt Group have the advantage of being directly emplaced into a continental environment and therefore their long term preservation is not influenced by the recycling which affects oceanic crust and consequently the oceanic plateaus embedded in them. CFBs are, however, susceptible to erosional processes, for example the Deccan Traps which today cover approximately 500,000 km² are estimated to have covered an original area greater than 1,500,000 km² (Krishnan, 1960; Raja Rao, 1978). However, oceanic intra-plate plateaus, such as Hawaii, are subject to the same poor preservation potential as oceanic plateaus erupted at mid-ocean ridges because of oceanic crust recycling.

4.3 Criteria for preservation of oceanic plateaus

The preservation potential of LIPs erupted into an oceanic environment, whether via a ridge-centred or intra-plate plume, is determined at least in part by the tectonic recycling affecting the oceanic crust in which they are embedded. Therefore, to be preserved for time scales greater than 200 m.y. (the oldest age of the oceanic crust) they need to be accreted as allochthonous terrains (Ben-Avraham et al., 1981). Ultimately, this is brought about by the collision of the plateau with a subduction zone, be it an island arc or continental convergent margin, enabling the oceanic plateaus to be either preserved via obduction and imbrication, or destroyed by subduction and subsequently recycled into the mantle; a fate similar to that of the bulk of normal-thickness oceanic crust. Whether these overthickened portions of the oceanic crust are obducted and accreted is dependant on their reluctance to subduct. This in turn is directly related to the plateau's isostatic buoyancy.

Views expressed by Ben-Avraham et al. (1981) that many of the oceanic plateaus which exist in todays oceans are "fated to be incorporated at active continental margins" are rejected by Cloos (1993). Cloos argues that only oceanic plateaus which have crustal thickness of greater than 30 km on a 100 km lithospheric keel, such as the OJP (Kroenke, 1984), have the potential to cause orogenesis when they reach subduction zones. These arguments are influenced by calculations predicting when oceanic crust will become negatively buoyant because of density increase. Such a density increase would allow the oceanic crust to subduct, i.e. be susceptible to slab pull once reaching an active convergent margin. Such calculations predict that normal oceanic lithosphere older than 10 m.y. has the potential to subduct, and oceanic plateaus with a crustal thickness of 17 km or less will also subduct, without orogenesis.

Therefore, according to Cloos the crust of the Caribbean Plateau, which is thought to have been originally between 10 and 15 km thick (Edgar et al., 1971; Burke, 1988), should have been efficiently subducted. Obviously, this was not the case because fragments of the Caribbean Plateau have been imbricated along the western coast of Colombia and accreted around the Caribbean Basin (Millward et al., 1984; Berrangé and Thorpe, 1988; Burke, 1988; Kerr et al., 1997). The Caribbean Plateau is unique in that portions of the plateau, which crop out on the island of Gorgona, expose the only Phanerozoic komatiites discovered to date (Aitken and Echeverria, 1984). Furthermore in the Bolívar region in west/central Colombia sheared gabbro layers, serpentinised dunite and banded lherzolites are exposed, which may represent the very deepest layers of the exposed plateau (Nivia, 1992; Kerr et al., 1997). Because plateaus are believed to have formed at higher temperatures than normal oceanic crust, they remain thermally buoyant for longer, often continuing to be thermally heated by the underlying plume. As the Caribbean Plateau is thought to have collided with the subduction zone within a few million vears of formation (Burke et al., 1978; Hill, 1993), it was therefore still very much thermally buoyant. The buoyancy aid in the case of the Caribbean Plateau was the Galapagos hotspot, which was responsible for its eruption (Duncan and Hargraves, 1984; Hill, 1993), and may have remained a distal thermal influence on the plateau's lithosphere. After imbrication of the Caribbean Plateau, subduction backstepped to the edge of the plateau (Figure 4.12) resulting in the intrusion of arc volcanic rocks, such as the tonalite intrusions of Aruba off the coast of Venezuela (e.g. Beets et al., 1984), into the oceanic plateau.

In comparison to accretion and imbrication of the younger and thermally hot Caribbean plateau, the obduction of the OJP occurred some 90 m.y. after the main plateau building event (though this time scale may be too high because of the younger ages of the Makira plateau lavas, Section 2.8). Only the upper portions of the plateau have been exposed (Section 4.5: Petterson et al., in press) and no high temperature picrites and komatiites, such as those exposed in the imbricated sections of the Caribbean Plateau, have been discovered to date. The OJP, thought to be greater than 30 km thick (Furumoto et al., 1976; Hussong et al., 1979; Gladczenko et al., in press), successfully blocked the Vitiaz Trench (Section 2.5.4). It does seem likely the thickness of the OJP was the main cause of obduction, with no thermal influence such as that responsible for the accretion of the Caribbean Plateau, possibly explaining the lack of exposure of the deeper regions of the OJP.

Briefly, other considerations which also influence the obduction of oceanic plateaus need to be mentioned. Firstly the nature of the subduction zone may have played an important role in the distinctly different obduction of the Caribbean and Ontong Java Plateaus. The Caribbean Plateau collided with a major continental margin, where as the OJP choked up the subduction zone of an island arc. If the Caribbean Plateau had collided with an island arc, obduction of its deeper portions may have not been so successful. Also, if the 11-15 km thick Caribbean Plateau had waited 100 m.y. before colliding with the South American continent then the plateau may have been subducted in its entirety. Furthermore, Cloos (1993) does not consider the composition of the lithospheric mantle, and how changes in its composition may influence its density and the plateau's buoyancy. For example, if plateaus have a lithosphere with a lower density than normal asthenospheric mantle, i.e. depleted harzburgite, then it will provide a reduced density lithospheric keel and the plateau will more effectively resist subduction (e.g. Saunders et al., 1996). A low density lithosphere keel under oceanic plateaus would provide an explanation for the shallower subsidence curves in oceanic plateaus compared to normal oceanic crust (Coffin and Eldholm, 1994).

4.4 Formation of décollements in oceanic plateaus

Section 4.3 outlines the necessary criteria for plateau preservation, which includes thickness, age, thermal state, lithospheric keel and the nature of the convergent margin involved.

The nature of the obduction itself may rely on the presence of décollements within the oceanic plateau.

"A décollement is a surface or zone of structural disharmony" (Moore, 1989). The concept of décollement in the Swiss Alps, implied in the formation of nappes, dates back to 1884 (Bertrand, 1884). Therefore, décollements have been known to play important roles in thrust tectonics and mountain orogenesis long before the existence of oceanic plateaus and large igneous provinces was known. It is known that sedimentary layers acted as ductile zones and that these facilitate the formation of thrust zones. The high pressures generated in the fluids within these sedimentary layers is thought to allow nappes to glide efficiently over each other (e.g. Hubbert and Rubey, 1959; Kehle, 1970; Barker, 1972; Gretener, 1981). At convergent margins potential décollements within the internal structure of an oceanic plateau, whether it is of a sedimentary nature or layered and sheared gabbros, provide essential zones of weakness necessary for thrust surfaces to develop. The formation of such zones of potential décollement within oceanic plateaus is related to their growth models (Section 4.2).

Figure 4.3 suggests some possible potential décollements within intra-plate plateaus, though some may also form potential décollements in ridge-erupted plateaus. For example, sedimentary layers intercalated within the lava pile form potential oceanic plateau décollements and are present in both intra-plate and ridge-erupted oceanic plateaus. Such sedimentary layers are not as likely to expose such deep regions as the magma chamber detachment layers suggested by Gibson and Gibbs (1987). However, sedimentary layers may be less common in highly effusive systems because of the more rapid and voluminous production of lava, leaving fewer hiatuses between eruptions needed for the deposition of sedimentary horizons. None the less, boundaries between individual flows in the lava pile form potential décollements, though if not interrupted by sedimentary horizons they would not form a ductile layer, more of an exploitable physical weakness in the lava pile.

Intra-plate provinces are emplaced into and on top of pre-existing crust, especially important in the case of oceanic hotspots because the ocean crust which they erupt through and onto may be topped by a considerable thickness of sediments. For example, Hawaii, is currently erupting through and onto Upper Cretaceous oceanic crust (Hilde et al., 1976). Although the magma may have assimilated most of the sediment around and under the developing seamounts,



Figure 4.3. Intra-plate hotspot and the formation of possible décollement layers (not to scale).

some may have remained and are potentially important décollements, possibly responsible for the scraping off of seamounts onto the fore-arc region of the northern most Kuril Arc (Ben-Avraham et al., 1981).

Porous layers in oceanic crust have also been suggested as possible planes of décollement (Kimura and Ludden, 1995). The décollement zone, originally within the overlying sediments, steps down to the more altered oceanic crust layers as it enters the subduction zone. High pore pressures result within the basalts and shear strengths are reduced, causing the uppermost layers of oceanic crust to be peeled off before being subducted. The alteration of basalt glass, and the infilling of vesicles with smectites and possibly carbonates causes a porosity contrast, an important aspect of Kiruma and Ludden's (1995) model in forming a décollement which allows the peeling off of oceanic crust slices at accretionary prisms. A thick oceanic plateau may similarly use altered, reduced shear strength layers as a potential décollement during obduction and imbrication. However, the extent of alteration reached throughout oceanic crust and plateau lavas is governed by the geothermal gradient of the basalts after eruption. The higher the geothermal gradient, the higher the grade of alteration reached in shallower depths of the oceanic crust/oceanic plateau.

4.5 Structure of the Malaita Anticlinorium - clues to obduction 4.5.1 Shallow level thrusting in the MVG - shallow level detachment layers?

Along the Singgalo River two thrust faults were observed, both having propagated along green/brown, highly siliceous mudstones intercalated between lava flows (Plate 3.3). A pillowed lava flow lying directly beneath one of the mudstone intercalations shows the uppermost pillows have flat surfaces (Plate 3.3). These pillows have either been eroded prior to the mudstone deposition, or their tops have been sheared off as a result of the over thrusting of the overlying massive flow. The latter seems the most likely scenario because the deposition of sedimentary horizons is uncommon in the MVG, therefore it is unlikely that the major time period needed for such basalt erosion would have occurred. However, this thrust/erosional contact shown in Plate 3.3 is poorly exposed at river level. Further upstream a better-developed thrust fault was mapped, clearly propagating along and shearing an intercalated mudstone unit in the MVG.

Shallow thrusts exposed within the MVG, as seen along the Singgalo River, developed

from the deeper rooted, high-angle thrusts which are believed to be responsible for the asymmetrical folds forming the Malaitan inliers (Rickwood, 1957; Hughes and Turner, 1976). These deeper structure are calculated to have formed at detachment surfaces between 1 and 4 km deep (Petterson et al., in press) in the underlying basaltic crust (Figure 4.5). However, these deeper rooted detachment surfaces are not interpreted as major décollements by Petterson et al (in press).

Shallow thrusts, which sometimes formed along sedimentary horizons between MVG lava flows, may have facilitated the obduction of the OJP lava pile exposed on the island of Malaita. However, because the MVG contains very few sedimentary horizons (Section 3.5.5), thrust development within the lava pile may have been hindered. Furthermore, while obvious exposed sedimentary intercalations are rare in the MVG, it is possible that where there is no exposure along the rivers, or where only individual flows are present with obscured upper and/or lower margins, sedimentary horizons and possible thrust zones could be disguised. If this is the case, the extent of thrusting throughout the MVG may be greater than originally thought.

The maximum alteration grade of the MVG appears to be into the zeolite facies (Chapter 6). Early reconstructions of the vertical depth within the MVG sampled was calculated to be as much as 3-4 km in the deeper regions of the Kwaio Inlier (Babbs and Saunders, 1995). Such depths would be expected to place the basalts normally well into the greenschist facies, and the apparent lack of alteration seemed therefore to require a very low geothermal gradient. However, the imbricated nature of the MVG resulting from shallow level thrusting, especially present in the Kwaio Inlier (Figure 4.5a - c) causes some repetition within the lava sequence exposed along stream segments, as indicted by geochemical traverses (Section 4.5.2). This repetition of shallow slices of MVG nicely explains the low metamorphic grade reached within the lavas exposed deep into the inliers.

Frustratingly, the true nature of the deeper detachment surface within in the MVG is still enigmatic. Petterson et al. (in press) observed no major décollements within the MVG. Therefore, while it is fair to say that no classic décollements are present in the MVG such as those responsible for the emplacement of the Jura of the Swiss Alps, the small thrusts seen within the shallow levels of the MVG could be interpreted as minor décollements assisting the imbrication of the MVG.



Figure 4.4. Main structural areas of Northern Malaita (Petterson, 1995b) with the location of the cross sections shown in Figure 4.5. The *en-echelon* alignment of the periclines outlined in the figure indicates a sinistral strike slip movement, i.e. transpressive deformation occured during the hard docking event causing the obduction of the Malaita Anticlinorium.



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Figure 4.5a-h. Cross sections of north and central Malaita (section locations in Figure 4.4) showing predominant NE fold and thrust vergence. Figure 4.5b. shows a small percentage of SW directed, latter stage thrusts located in the Kwaio Backthrust Belt. Calculated detachment depths range from 4 to 1 km deep and percentage shortening from 34-46% (Petterson et al., in press).

Figure 4.5a-h

4.5.2 Geochemical evidence for shallow-level thrusting in the MVG

Attempts to reconstruct the MVG lava sequence is hindered because of the restricted exposure on Malaita. However, it is reasonable to speculate from the small number of thrusts which are revealed in the MVG, and the lack of any high to moderate grade of metamorphism (Chapter 6), that imbrication of the MVG lava pile has occurred, and has probably been extensive. One way of considering the exact extent of this imbrication comes from studies of geochemical profiles drawn along the rivers we sampled. Geochemical profiles were reconstructed by, first determining the average strike of the MVG, and then projecting the sample positions to a line drawn perpendicular to the average strike. The results of these sample projections along the rivers sampled can be seen on the sketches in the back pocket. By projecting the samples to a line perpendicular to strike, the effects of river bends can be minimised. Geochemical river profiles have been drawn up for the three rivers which we sampled that cut through the western limb of the Kwaio Inlier, these are the Singgalo, Kwaleunga and Ala'olo Rivers. (The Singgalo was a continuation of sampling by Clive Neal and John Mahoney (unpublished data) in the field season prior to ours).

Figure 4.6 outlines the geochemical profile for the Singgalo River, both including the lower Singgalo samples which were collected by Clive Neal and John Mahoney (unpublished data), and Figure 4.7 shows both the Kwaleunga and Ala'olo River profiles. The distance, shown in meters, is not a true vertical depth into the lava pile, but rather the horizontal distance along which we sampled from the beginning of the MVG's exposure. For a vertical distance into the lava pile I estimate values of ~3700 m for the Singgalo River (calculated from a average dip of 40°), ~3900 m for the Kwaleunga (average dip 37°) and ~ 2000m for the Ala'olo River (average dip of 30°). The geochemical profiles suggest that these estimates are not true vertical depths and that extensive imbrication has occurred. Hence, there may be no need for an unrealistically low geotherm to explain 4 km thickness of basalts which do not reach metamorphic grades greater than brownstone and into zeolite facies (Chapter 6).

Cycles in the chemistry of the MVG basalts along the river profiles reveal that tectonic imbrication has caused repetitions in the lava sequence. However, an obvious alternative is that these cycles are the result of repeated fractionation, magma replenishment and eruption resulting from magma chamber processes (RTF). The best example of what appears to be a sequence of

48 56 64 3 5 7 8 13 18 0.8 1.3 1.8 0 0 a) b) Mg. no. Nb (ppm) c) Al₂O₃/TiO₂ d) (La/Sm)n 1000 1 000 d d i 2000 S 2000 s t t a n an С С е (m) е (m) 3000 3000 1.6 0.6 1.1 e) Ni/Zr 4000 4000 5000 5000 6000 6000 Figure 4.6a-e. Geochemical profiles for the Singgalo River. Shaded bars represent possible magma chamber replenishment events, and/or tectonic imbrication (Upper Singgalo samples; J. Mahoney and C. Neal unpublished data).



Figures 4.6a-e



Figures 4.7a-h

magmas erupting from a fractionating magma chamber is the upper portions of the Singgalo River profile, i.e. up to 3000m up river. The lavas sampled towards the top of the lava pile show increasingly evolved compositions, reflected by decreasing Mg numbers (Figure 4.6a) and Al_2O_3/TiO_2 ratios (Figure 4.6c), as well as a systematic increase in the concentrations of incompatible elements (Figures 4.6b, d and e). Deeper portions of the Singgalo River profile appear to show smaller scale RTF cycles and possible hiatuses (indicated by the shaded bars). Nevertheless, these apparent cycles at deeper levels in the Singgalo profiles are most likely due to tectonic imbrication, as indicated by the small number of thrusts that were mapped (Section 4.5.1).

Geochemical profiles from the Ala'olo and Kwaleunga Rivers are also shown for comparison. Unlike the Singgalo they do not show obvious RTF cycles along the lava pile, though in both cases the lavas gradually become more evolved in their compositions towards the uppermost sections. Both the Ala'olo and Kwaleunga show shaded bars, but in this case they do not represent what appears to be obvious RTF breaks, as in the Singgalo. The shaded bar represents a break in the MVG lava sequence by a section of sediment which has been tectonically imbricated.

Therefore it seems that, while some of the geochemical profiles appear to show well developed RTF sequences, tectonic imbrication has also influenced the geochemical profiles. Thin section and XRD studies (Section 6.2) along the Kwaleunga River and deeper portions of the Singgalo River reveal that there is no obvious increase in alteration with depth. Hence I conclude that the tectonic imbrication has caused at least several repetitions in the lavas sequences of the MVG, though RTF processes have occurred, and have been particularly well preserved in the upper portions of the Singgalo.

4.5.3 Structural deformation on Malaita - clues to the obduction mechanism

The intense or hard docking event which resulted from the increased coupling between the OJP and Solomon Block (Section 2.5.6) was probably caused by the onset of subduction of the thermally buoyant, young Woodlark Basin (Petterson et al., in press). The NW-SE structural grain of Malaita is in accordance with a NE-SW directed oblique compressional event which resulted in sinistral strike slip movement as indicated by the *en-echelon* arrangement of the anticlines (Figure 4.4) and the sigmoidal fold axial trends to the north of the Kwaio Inlier (Figure 3.4; Section 3.8.1). This transpressive deformation phase which resulted in the formation and eventual emergence of the Malaita Anticlinorium (Rickwood, 1957; Hughes and Turner, 1976; Petterson, 1995b) is stratigraphically bound by the Suafa Limestone Formation, dated around upper Miocene to lower/middle Pliocene (Section 3.6.5). The Suafa Limestone Formation, which represents a major shift to shallow water deposition with its current bedding and benthic fauna, was the youngest unit to be affected by the hard docking deformation event. In turn the deformed Suafa Limestone Formation is unconformably overlain by localised Upper Pliocene/Pleistocene to Recent units which remain undeformed (Petterson, 1995b). The transpressive deformation is consistent with the oblique convergence occurring presently between the Pacific Plate moving north-west and the Indo-Australian Plate moving north-eastwards, currently affecting the whole of Melanesia.

Malaita is dominated by a well-developed NW-SE structural grain, with an average strike of 128° (Section 3.8). Fold axis orientation follows this NW-SE grain, with the upright to overturned vergence of these folds being predominately towards the NE (Petterson et al., in press). Indeed 90% of the fold vergence in northern Malaita as mapped by Petterson (1995b) is towards the NE with the Toambaita Inlier having the only SW verging fold axis (Figure 4.5h). Hughes and Turner (1976) report 4 out of 5 mapped inliers in south and Small Malaita show fold vergence is to the NE. The asymmetrical folds accommodate between 34 and 46% crustal shortening (Figure 4.5); the larger folds are interpreted as being the tips of blind thrusts from a detachment surface 1-4 km beneath the cover sequence (Petterson, 1995b). Thrust vergence is therefore predominately towards the NE, though a small group of SW-directed thrusts occur in the Kwaio Backthrust Belt (Figure 3.7). Petterson (1995b) interprets these SW-directed thrusts as a later stage, SW-directed, event as they appear to cut pre-existing NE-directed thrusts (Figure 4.5b).

The dominant NE-verging structures have been discovered elsewhere on the Malaita Anticlinorium. Across the Ramos Ridge, which lies between Santa Isabel and Malaita, NE-verging folds in the sedimentary cover sequence have been imaged by enhanced seismic reflection (Kroenke, 1972; Bruns et al., 1986, 1989) (Figure 4.10 insert). The timing of these thrust and deformation events is unclear because of their submarine environment. It may be therefore that the deformation which formed the Malaita Anticlinorium was a longer event than is recorded

within Malaita.

It is suggested that these NE-verging folds and thrusts dominating the structure of the Malaita Anticlinorium originally formed as normal faults when the OJP arched on collision with the North Solomon Trench and were later reactivated as thrust faults (Coleman and Kroenke, 1981; Petterson et al., in press). This certainly may be the case with thrusts that are restricted to the sedimentary intercalations. On collision with the North Solomon Trench and intense transpressive deformation they are reactivated as thrust faults.

Finally, the most important information on the obduction of the Malaita anticlinorium may be found not on Malaita itself, but on Santa Isabel. The Kaipito-Korighole Fault (KKF) system on Santa Isabel (Hawkins and Barron, 1991) is the obducted major terrain boundary between the OJP and Solomon Block. This boundary is thrust faulted, trending in a NE direction, the same vergence which dominates the structure of Malaita. No accretionary prism is recorded on Santa Isabel, but a 62 - 46 m.y. ophiolite assemblage (Tejada et al., 1996) is exposed. The exact nature and origins of the ophiolite assemblage obducted on Santa Isabel are unclear, although the young ages indicate it is unlikely to be part of the Pacific Plate.

4.6 Obduction models for the Malaita Anticlinorium

4.6.1 The tectonic flake model

The later-stage SW thrusts seen in the Kwaio Backthrust Belt on Malaita, have been interpreted as the result of a later-stage, SW-directed, detachment of a 5-10 km thick upper 'flake' of the OJP across the North Solomon Trench and overriding onto the Solomon Arc forming the Malaita Anticlinorium (Petterson, 1995b; Petterson et al., in press). Furthermore, sidescan images by the South Pacific Applied Geoscience Commission (SOPAC) of the seafloor surrounding Malaita reveal NW-SE ridges with numerous SW-directed thrust (Figure 4.8), the direction of these thrusts being determined from the sediment off-loading from the frontal thrust ridge into their associated foreland basin (SOPAC, 1994; Kroenke, 1995).

In the 'flake tectonic' model (Figure 4.9), the predominant NE-directed thrusts form initially as a result of the increased coupling between the Solomon Block and the OJP some 6-4 m.y. ago. As subduction of the lower levels of the OJP continued the Solomon Block acted as a knife, cutting the OJP effectively in two, allowing the subduction of the lower levels to continue, while

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Figure 4.9a&b. The flake tectonic model of the obduction of the Malaita Anticlinorium as outlined in Petterson et al (in press). a; the NE thrusts propagate through the SW margin of the OJP as it is thrust into the Solomon Block intialising the onset of formation of the Malaitan Anticlinorium. SW-directed subduction of the lower levels of the OJP continues. b; the anticlinorium begins to overthrust the NE edge resulting in a 6-10 km thick flake of the OJP being obducted over the leading edge and onto the Solomon arc block. (Exaggerated vertical scale).

an upper flake of plateau crossed the North Solomon Trench and is thrust south-westwards over and onto the arc (Petterson et al., in press). Evidence for the transport of an upper flake of the OJP across the North Solomon Trench is given by seismic images that show apparent continuous draping of sediment across the North Solomon Trench (Kroenke, 1972; 1989).

4.6.2 NE thrusting and imbrication of the OJP

Alternatively, the formation of the Malaita Anticlinorium may have been due to the imbrication of upper portions of the OJP by NE-directed thrusts such as those which dominate the structure of the anticlinorium as seen exposed on the island of Malaita (Figure 4.10). The anticlinorium is built of NE-verging thrust sheets, with the SW-directed backthrusts (which appear on Malaita only in a small numbers) not being considered to play a major role in the obduction, possibly being later-stage pop-up or displacement structures. These predominant NE-verging structures, and later stage SW-directed structures, may have been the result of onset of SW-directed subduction.

Figure 4.10 insert (Kroenke, 1972) shows the predominance of the NE-verging thrusts across the anticlinorium offshore, northwards of Malaita. This second model, while complementing the results of such seismic imaging as shown in the insert to Figure 4.10 also stems from direct interpretation of the geology as exposed and mapped on Malaita.

4.6.3 Discussion

The flake tectonics model (Section 4.6.1) uses the small percentage of SW-directed thrusts within the Kwaio Backthrust Belt and numerous SW thrusts mapped on SOPAC maps as evidence for a SW-directed, major detachment of a 5-10 km thick portion of the Malaita Anticlinorium. The SW-directed flake is believed to have been the result of the lower Solomon Block effectively slicing the OJP in two, with the lower portions continuing to subduct. This model accounts for the presence of a large number of SW-directed thrusts on the sea floor off the east coast of Malaita (Figure 4.8). The apparent direction of sediment transport into the foreland basins, depicted on the SOPAC images, is used to help determine the direction of thrusting. If the direction of sediment transport were misinterpreted and the thrust directions wrong, then the actual percentage of SW-directed thrusts are required to be deep-rooted structures if they are responsible for flake detachment and transport. However, while they cut earlier NE-directed



Figure 4.10

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thrusts, they appear to be shallow structures with a maximum depth rooting back to the original NE-directed detachment planes (Figure 4.5b). They certainly do not appear to have been formed as deep as 5-10 km.

Flake tectonics was first hypothesised by Oxburgh (1972) when he proposed that a slice of crust is split from the down-going slab and thrust (in the opposite sense of normal subduction related thrusting) on top of the overriding continent. Some (e.g. Hamilton, 1988) argue that not only is such "obduction" mechanically impossible, but also has yet to be proved to have operated anywhere. For the Solomon Block to have been capable of slicing the OJP effectively in two, a major décollement would have been needed at depth. No evidence of such an effectively ductile zone exist in the exposure of the OJP on Malaita, nor does Santa Isabel show such a décollement along the Kaipito-Korighole Fault (KKF) system. Furthermore, why would the ophiolite assemblage on Santa Isabel be preserved thrust against the OJP with a NE sense if the OJP has been thrust in a SW direction over the arc? The presence of the ophiolite is more consistent with the accretion of arc material, and surrounding sea floor, jamming up against and possibly overriding the margins of the OJP.

Further evidence in support of the flake model is the apparently continuous draping of an unbroken sedimentary cover on top of the OJP across the North Solomon Trench and up the western slopes of the anticlinorium (Kroenke, 1972; 1989). However, the North Solomon Trench is an active trench, as shown by the seismic cross sections which indicate current subduction of a SW-dipping slab below Santa Isabel (Figure 2.4). Therefore, the North Solomon Trench is a major thrust zone, accommodating movement on the scale of kilometres, not a passive feature. Consequently, the sediment across its width and continuing up the NE slopes of the Malaita Anticlinorium cannot be truly continuous.

In contrast, the imbricated thrusting model (Section 4.6.2) has evolved almost purely from the structure of the anticlinorium as exposed on the island of Malaita. The later stage SW-arcdirected, thrusts only occur as a small minority of the overall thrusting direction on Malaita and therefore are not considered to have formed a major structural control on the obduction. These SW thrusts could be the result of pop-up (Charvet and Ogawa, 1994) or small flower-like structures which resulted from intense, later stage transpressive and strong sinistral strike slip motions between the Solomon Block and the OJP. Small scale backthrusting is a common feature of accretionary tectonics, seismically imaged for example off the north coast of Panama and across the Sunda Arc in Indonesia (Silver and Reed, 1988) as the sedimentary accretionary prism is wedged against the forearc basin. Major scale landward verging backthrusts also occur in the accretionary prisms off Washington and Oregon (Charvet and Ogawa, 1994) however these backthrusts occur in sedimentary accretionary wedges, not thick piles of basalt.

The imbrication model also requires no splitting of the OJP by the Solomon Block, a difficult process to physically explain or prove as island arcs are typically composed of andesite and volcanic ash material. Andesites and volcanic ash are less dense and physically weaker material than basalt and therefore unlikely to have caused the splitting of a major pile of basalts. Therefore, less dense, siliceous arc material is not likely to have remained so undeformed as depicted in Figure 4.9, and being more buoyant than basaltic crust would have overridden the OJP.

The depth reached by the SW-dipping subducting slab below Santa Isabel is approximately 200 km (Figure 2.4D), depths not inconsistent with subduction having been occurring for approximately 2-3 m.y. By 2-3 m.y. the main obduction of the Malaita Anticlinorium may have already occurred. It may have been that subduction back-stepped slightly to a more NE position (as shown by Kelleher and McCann, 1977, Figure 6, page 120) hence creating a new subduction zone and a new trench, or subduction may have just resumed along the relict North Solomon Trench. The onset of new or resumed SW subduction may have been responsible for the emergence of the anticlinorium above sea level as the subducting plate successfully pushed beneath the newly formed anticlinorium. This newly subducting, underlying plate beneath the anticlinorium may also have been responsible for the formation of the later SW-directed thrusts as an uplift and/or lateral pressure increase caused localised pop-up or flower structures. I speculate that the whole of the Malaita Anticlinorium started life as an accretionary prism and went on to undergo large scale (as well as local) pop-up or flower deformation because of the resumed subduction, rather than flake detachment and emplacement. This may provide a possible explanation for the numerous SW-directed thrusts on the SOPAC images.

It is the latter of the two obduction models, the NE imbrication of the anticlinorium with latter SW-directed pop-up structures, which is preferred in this text. Doubtless, until the real deep structure of the Solomon Block is discovered, possibly via deep seismic imaging, questions about how the OJP obducted will remain.

4.7 The future of the OJP - potential plume or greenstone belt?

Will the OJP eventually undergo wholesale subduction, recycling back into the mantle and removing any trace of the plateau which may be preserved in a future geological record? Or will it obduct on a bulk scale and form a major component in the growth of a future continent? If the OJP was to sink down into the mantle on a large scale, then it may ultimately be recycled, possibly forming an important component in a future plume (e.g. Hofmann and White, 1982). Alternatively, if portions resist subduction then the OJP may form an accreted allochthonous terrain, possibly a future greenstone belt (e.g. Storey et al., 1991; Kusky and Kidd, 1992).

The origin of Archaean greenstone belts is controversial and has important implications for oceanic plateau preservation. Greenstone belts with their thick sequences of sub-aqueous tholeiites and ultramafic komatiites, led to early speculations on their origins as allochthonous remnants of Archaean ophiolites (Burke et al., 1976; de Wit et al., 1987). Others envisaged an autochthonous relationship between the greenstone belts and their associated continental terrains (e.g. Nisbet, 1987). For example, geochemical analysis of the Belingwe greenstone belt in Zimbabwe seems to imply assimilation of the underlying continental crust, suggesting that the lavas erupted through the Archaean crust in a continental rift environment (Blenkinsop et al., 1993). Kusky and Kidd (1992) suggested that the Belingwe greenstone belt was the remnants of an allochthonous oceanic plateau. They identify a major mylonite décollement zone along the base of the high temperature ultramafics, which allowed emplacement of the plateau over the shallow water sediments and gneiss basement.

The presence of high temperature, ultramafic komatiites in most greenstone belts has been attributed to the higher mantle temperatures thought to have existed in the Archaean, and which may have caused higher rates of sea floor spreading and thicker oceanic crust (Burke et al., 1976; Sleep and Windley, 1982; Bickle, 1986). The only Phanerozoic komatiites discovered to date occur on the island of Gorgona off the east coast of Colombia (Gansser et al., 1979; Echeverría, 1980). The komatiites, and associated picrites, gabbros, tholeiites and tuffs, represent an accreted portion of the Cretaceous Caribbean Plateau (e.g. Duncan and Hargraves, 1984; Hill, 1993; Kerr et al., 1996). Therefore, these rare Cretaceous komatiites may provide a vital link between predicting the future of Phanerozoic oceanic plateaus and understanding the origins of Archaean greenstone belts (Storey et al., 1991).

The Caribbean plateau, accreted along the western coast of Colombia (i.e. Gorgona) and throughout the Caribbean Basin, is thought to be analogous to greenstone belts such as the Belingwe greenstone belt in Zimbabwe. The island of Aruba, lying to the north of Venezuela, also represents an imbricated portion of the Caribbean Plateau. While Aruba shows no evidence of komatiites, it does expose younger, arc-related volcanics which punch through the obducted plateau lavas. Similar, arc-related, later stage calc-alkaline volcanics appear to have erupted through the Abitibi greenstone belt in Canada (Desrochers et al., 1993). The Abitibi greenstone may have formed when a series of Archaean oceanic plateaus collided with a subduction zone forming an ever-growing Superior Province. After the plateau accreted an extensional ridge reached the subduction zone erupting the younger calc-alkaline volcanics (Desrochers et al., 1993).

The obduction of the OJP which formed the islands of Santa Isabel and Malaita has been compared to greenstone belt formation (Tejada et al., 1996). However, no komatiites have been discovered to date either on the obducted portions of the OJP in the Solomon Islands or drilled on the main body of the plateau. No major ductile, mylonite décollement has been discovered along the KKF in Santa Isabel, the major terrain boundary between the OJP and the Solomon arc block, such as Kusky and Kidd (1992) described along the base of the Belingwe greenstone belt and found in the Bolívar region of the Caribbean Plateau (Nivia et al., 1992; Kerr et al., 1997).

Subduction of the OJP may have occurred on a small scale, explaining the presence of seismically active SW-dipping slab beneath Santa Isabel (Figure 2.4). In addition, subduction of the southernmost margins of the OJP along the Vitiaz/Cape Johnson Trench to the east of Makira (San Cristobal) may have been occurring for at least 2 m.y. Figure 4.11 shows the SOPAC sidescan images for the Melanesian arc gap to the east of Makira which show 3 distinct lines of volcanic centres. The southernmost chain, the San Cristobal magmatic arc, is believed to be related to the subduction of the Indo-Australian Plate in a northwards direction below the Solomon arc block at the San Cristobal Trench. The magmatic centres labelled as the Rejuvenated Vitiaz Magmatic Zone are related to the southwards subduction of the Pacific Plate



Figure 4.11

along the Vitiaz Trench. There is a third and final chain of volcanic centres, labelled the synrift volcanic zone, which is believed to be extension related, possibly the result of slab roll back of one or both of the opposing polarity subduction zones. Volcanic centres related to the Vitiaz Trench indicate that the trench was active at least 2 m.y. ago because volcanism on Anuta and Fatutaka (volcanic centres to the east of the area shown in Figure 4.11) is dated at 2.2 ± 0.1 m.y. (Jezek et al., 1977).

The apparent partial subduction of the SW margins of the OJP at the North Solomon Trench and the southern margins along the east Cape Johnson Trench may cease in the future as thicker portions of the plateau (estimated to be as much as 42 km thick, Furumoto et al., 1976; Hussong et al., 1979) more efficiently jam up the trenches. If the subduction does cease completely or the Pacific Plate push is too great then the subduction may backstep to the margin of the OJP, a similar scenario to the tectonic setting which affected the Caribbean Plateau in Aruba. Eventually an active volcanic arc may form on the margin of the OJP and arc lavas may intrude though the plateau. Figure 4.12 outlines some of the possible scenarios of initial plateau obduction, i.e. via the flake tectonics (Figure 4.12a) or thrusting and imbrication. Figure 4.12 also outlines future predictions which may effect the OJP, depending on whether the current obduction occurred via flake tectonics (Figure 4.12b) or by an accretionary style thrusting and imbrication (Figure 4.12d).

A good example of how oceanic plateaus may cause subduction to backstep occurred in the Bering sea. During the late Mesozoic/Early Tertiary, subduction at the original continental margin of the Bering Sea shifted to the Aleutian Trench (Cooper et al., 1976). Three thickened oceanic edifices today sit to the north of the Aleutian trench, the Umnak Plateau, Bower and Shirshov Ridges (Ludwig et al., 1971; Cooper et al., 1980). These edifices are thought to have caused the subduction to backstep to a more favoured regime (Ben-Avraham et al., 1981).

If such an oceanwards backstep was to affect the OJP in the future, continued subduction of the Pacific Plate may occur beneath the plateau. This could result in other oceanic plateaus, embedded in the Pacific Plate such as the Nauru Basin and Magellan Seamounts, to eventually dock with the subduction zone, possibly colliding with island arcs which may have formed along the margins. Further backsteps may occur if any of these later docking plateaus are thick enough to resist subduction, possibly resulting in a series of accreted plateau terrains, with

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arc volcanics sandwiched between, a similar scenario to that envisaged by Desrochers (et al., 1993) to have formed the Abitibi greenstone belt in eastern Canada. If a plateau does cause subduction to backstep then fluids from the subducting slab will be introduced to the overlying plateau. This addition of slab water to the plateau eventually may induce the lower levels of the plateau to undergo transformation from garnet granulite to eclogite (Saunders et al., 1996). An increase in the density of the lower levels of the plateau may bring about spontaneous subduction of the lower levels of the plateau in a piece-meal fashion.

The OJP and other oceanic plateaus such as the Caribbean, Manihiki Plateaus and the Nauru Basin were erupted during the Cretaceous superchron (Larson, 1991; Larson and Olson, 1991). Cretaceous oceanic plateaus cover some 3% of the ocean floor (Sandwell and MacKenzie, 1989) and appear to have formed during an episode of unusually active plume eruptions, suggested to have been the result of enhanced mantle convection (Stein and Hofmann, 1994). Similarly, a large portion of what potentially may have been Archaean oceanic plateaus are preserved as greenstone belts. If these greenstone belts are the most ancient of oceanic plateaus, then the geological record of oceanic plateaus preserved between the Archaean and Cretaceous is rather poor. It could be that Archaean Plateaus were favourably obducted because of higher temperatures of formation, as indicated by their high temperature mafic komatiites, or there may have been more plateaus erupted in the Archaean as occurred in the Cretaceous. Therefore, a large portion of the Cretaceous plateaus that are beginning to accrete around the Pacific rim may successfully obduct as allochthonous terrains, forming potential greenstone belts in a future geological record. A future episode of rapid crustal growth?

4.8 Conclusions

• Oceanic plateaus form via lateral or vertical growth models. Lateral growth requires the melting plume head to capture a spreading axis, a modern day analogue of which is the Iceland plume which has entrapped the North Atlantic spreading ridge. Vertical growth models of oceanic plateaus, i.e. such as responsible for formation of continental flood basalts and the Hawaii-Emperor seamount chain, involves a plume erupting in an intra-plate setting. Some plumes remain within one tectonic setting for their entire eruption history, though most swap to involve both eruption models. For example, the Kerguelen plume initiated small scale melting under the
thick Indian continental lithosphere, then it switched to an oceanic spreading ridge environment erupting the bulk of the lavas, finally moving to its current oceanic intra-plate setting in the South Indian Ocean.

Growth models are important as they control formation of potential décollements within oceanic plateaus. Such décollements are thought to form along; i) sedimentary intercalations in the lava pile; ii) alteration (hence porosity) changes with increasing depth throughout the basalts;
 iii) upper gabbro/magma chamber layers at the base of the plateau's lava pile.

• The preservation of plateaus relies on their being able to resist subduction. To resist subduction plateaus require positive buoyancy, which is directly related to their thickness, age and thermal condition at the time of collision. Normal 7 km thick oceanic crust is neutrally buoyant some 10 m.y after formation, oceanic plateaus that are >30 km thick are positively buoyant, regardless of age, and hence inherently obductable.

• The OJP is believed to have erupted via a spreading ridge between 122 and 90 m.y. ago. On initial collision with the North Solomon Trench approximately 100 m.y. after formation the OJP was cold, having long since moved away from the thermal effects of the plume responsible for eruption. However the OJP passed over the Rarotongan or Samoan Plume around 43 m.y. and this may have reheated the lowermost layers on a regional scale, as suggested by the presence of small scale, intrusive, alkali basalts (Tejada et al., 1996). At the time of hard docking around 6-4 m.y., any of these later stage thermal influences are likely to have dissipated. In comparison, the Caribbean Plateau was thought to be thermally buoyant when it collided with the south American continent, allowing the lowermost regions of the plateau to be imbricated and preserved.

• The island of Malaita shows a structure dominated by a NW-SE grain. Thrust and fold vergence is predominately towards the NE with a later-stage small-scale SW vergence. The *enechelon* periclines are consistent with an oblique convergence causing strong sinistral strike-slip motion throughout the Malaita Anticlinorium. This transpressive deformation began during the Pliocene and is well bracketed by the Suafa Limestone with only the uppermost recent local strata being undeformed.

The Malaita Anticlinorium is essentially an accretionary prism formed by NE-verging thrust

imbrication. Models which propose a later stage, SW-directed detachment and transport of an upper 5-10 km thick flake of the OJP across the North Solomon are not consistent with the structure of the OJP as seen exposed on Malaita. The KKF, the major terrain boundary between the Solomon arc block and the OJP is a NE-directed thrust and is in the wrong sense to support the emplacement of a tectonic flake of the OJP over the arc. Small-scale SW thrusts on Malaita are interpreted as pop-up or flower structures, possibly related to resumed (or backstepped) subduction along the North Solomon Trench. However, until deep seismic imaging can provide conclusive evidence for the deep-seated structure of the anticlinorium, i.e. does the arc underlie the OJP, the flake tectonics model can be neither substantiated nor dismissed.

• Some subduction of the OJP seems to be occurring today along the North Solomon Trench and possibly along the westernmost regions of the Vitiaz Trench. In the future subduction may jam completely causing the subduction to backstep to the north-eastern margin of the OJP. Arc lavas may erupt through the plateau as a result of subduction backstep. Introduction of slab water to the lowermost regions of the plateau may cause transformation of garnet granulite to eclogite which may then cause spontaneous subduction backsteps may deter this metamorphism and eventual piece-meal plateau subduction for long enough to allow it to be accreted as a continental allochthonous terrain.

• The unrepresentatively large proportion of Cretaceous oceanic plateaus existing the ocean basins and being accreted around the Pacific Rim may cause an anomaly in the preservation of oceanic plateaus in future geological records.

• The island of Gorgona, with its rare Phanerozoic komatilites, is believed to be an obducted portion of the Caribbean oceanic plateau and is therefore an important analogue to the origin of Archaean greenstone belts. The predicted future of oceanic plateaus accreting around the edge of the Pacific rim today may be directly analogous to how Archaean greenstone belts were emplaced.

Chapter 5

Petrography and mineral chemistry of the Malaita Volcanic Group

5.1 Introduction

This chapter gives an outline of the petrography of the MVG basalts and dolerites, including phenocrysts and groundmass assemblages. Textural features are described ultimately enabling the establishment of the crystallisation sequence of minerals in the MVG. Also presented are the results of electron microprobe analysis carried out at Leicester University (full results are presented in Appendix G) which enable a detailed study of the chemistries of the minerals within the MVG, including both phenocryst and groundmass phases, allowing an insight into magma chamber processes and pressures of crystallisation.

The grade and extent of alteration displayed by the MVG basalts are low, and have generally only affected olivines and any mesostasis. Optical and electron microprobe studies were carried out on the alteration phases and are included in Chapter 6, along with the results from XRD studies on the alteration products of the MVG basalts.

5.2 Petrography

The MVG basalts have been crudely divided into those that contain obvious phenocryst phases, and those that don't. However, it is essential to point out that the dolerites, common in the centres of the thicker units, may hide any magmatic phenocrysts because of their coarser grained textures. Only the finer grained, more rapidly cooled facies are likely to reveal the presence of phenocryst phases.

5.2.1 Petrography of the phenocryst-bearing basalts

Approximately 10% of the MVG basalts show any phenocrysts. Observable phenocrysts are predominately plagioclase, less commonly pseudomorphed olivine (I discovered no fresh olivine phenocrysts in the basalts), augite and minor opaques. The phenocrysts can be found as isolated individual crystals, but are more often grouped forming glomeroporphyritic clusters (Plate 5.1). Plagioclase is the most abundant phenocryst mineral in the MVG basalts, with compositions ranging from An₆₆ to An₇₈. They are typically anhedral with albite twinning, and some of the larger phenocrysts also show oscillatory zoning with more sodic rims (Section 5.3.2). The size of these plagioclase phenocrysts varies, but none exceeds 2-3 mm in length. This section excludes the plagioclase megacrysts which are several centimetres in diameter and discussed in a separate section below (Section 5.2.3).

The sparse olivine phenocrysts have been totally pseudomorphed to brown clays, predominately saponite (see Section 6.3). These olivine pseudomorphs vary from anhedral to well rounded, and the size of these pseudomorphs varies up to a maximum of 3 mm.

The clinopyroxene phenocrysts are typically subhedral to anhedral in form, with a slight brown body colour, and they are the least abundant phenocryst phase found within the MVG basalts. In one sample (KW7), two of the augite phenocrysts were found to enclose very small plagioclase crystals (Plate 5.2). These small plagioclase inclusions have high anorthite contents (An₇₂ to An₇₆), which fall within the range of An values displayed by the rest of the phenocrysts within the same basalt (Figure 5.2). This enclosure of plagioclase phenocrysts by augite phenocrysts, and the rarity of augite phenocrysts, is good evidence that plagioclase crystallised out as a phenocryst phase in the MVG basalts before clinopyroxene. The paucity of clinopyroxene phenocrysts within the MVG basalts is contradictory to their chemistries which indicates that clinopyroxene was an important precipitating phase in the magmas prior to eruption (Section 7.4).

The bulk of the groundmass mineralogy consists of mostly plagioclase and clinopyroxene. Plagioclases and pyroxenes coexist in an intergranular texture in the finer grained groundmass (<0.25 mm), while in the coarser grained basalts (≤1 mm plagioclase laths) clinopyroxene often subophitically encloses groups of radial plagioclases. Olivine may have been present as a mineral phase in the groundmass or as microphenocrysts, but, as with the olivine phenocrysts, it has now been entirely pseudomorphed by brown saponite. The remainder of the groundmass consists of smaller amounts of titaniferous magnetite and intersertal mesostasis which has now been replaced by predominately smectites such as saponite and possibly minor celadonite (the composition of the alteration products is



Plate 5.1. Glomeroporphyritic cluster of plagioclase phenocrysts, plus one clinopyroxene and a pseudomorphed olivine (ol) phenocrysts (sample KW7). Field of view 3 x 2 mm.



Plate 5.2. Augite phenocryst found to ophitically enclose small plagioclase phenocrysts (sample KW7). Field of view 3 x 2 mm.

discussed in Sections 6.2 and 6.3). The titaniferous magnetite varies in size and form, ranging from skeletal to well rounded and anhedral forms. In one basalt (KW7) quartz is also found as a minor, groundmass mineral phase.

5.2.2 Petrography of the dolerites

The majority of the MVG lavas (>80%) do not show obvious phenocryst phases but because this is more likely to be a factor of their coarser grained nature hiding any original magmatic phenocrysts, this section is sub-headed dolerites. For example, the compositions of some of the plagioclases in the dolerites overlaps with the plagioclase phenocrysts in the basalts (Section 5.3.2), therefore it seems fair to say that these magmas do contain phenocrysts.

As with the finer grained basalts, the dominant mineral phases in the dolerites are plagioclase and augite. Smaller amounts of altered olivine crystals are present, although only one sample (ML219 with a whole rock MgO = 10.02 wt%) was found to contain fresh olivines. Some of these olivines in ML219 have fresh cores and are surrounded by an alteration corona of saponite (Section 6.3).

The rest of the mineralogy is made up of opaques, mostly titanomagnetite (but very fine ilmenite needles have been found as a minor phase in some samples), and mesostasis (now replaced by alteration products). Minor later-stage mineral phases within the basalts (<1%) also include quartz crystals and less commonly (<1‰) apatite and ilmenite. Sulphide blebs were also discovered in one of the clinopyroxenes (sample KW13).

The texture of the dolerites varies between intergranular to ophitic, often incorporating both, with intersertal glass. The subophitic to ophitic texture of the dolerites often means that they have a slight mottled effect (Plate 5.3). Plagioclase may have been precipitated out of the magma before clinopyroxene, hence the common ophitic texture (Plate 5.4), with the plagioclase forming a nucleation point for the rapid clinopyroxene growth in the cooling dolerites. Plagioclase continued to co-precipitate along with clinopyroxene, this is evident as the size of some of the plagioclases within the ophitic texture increases from the middle to the edge of the enclosing clinopyroxene, hinting that surrounding plagioclases continued to grow until trapped within a proximal expanding clinopyroxene. However, in at least one sample (KW9, the most evolved dolerite with MgO = 5.7 wt% and Zr = 160.6 ppm) chadacrysts of



Plate 5.3. Ophitic texture within the MVG dolerites is common, meaning they often have a slightly mottled effect, or 'ophimottled' texture (sample TLB25). Field of view 10 x 6.7 mm.



Plate 5.4. Single clinopyroxene crystal (as shown in Plate 5.3, to the right of view) ophitically enclosing plagioclase laths and titanomagnetite (sample TLB25). Field of view 3 x 2 mm.



Plate 5.5. Chadacrysts of clinopyroxene encased in oikocryst of plagioclase (sample KW9). Field of view 1.5 x 1 mm.

are encased in oikocrysts of plagioclase (Plate 5.5). These plagioclase oikocrysts are more sodic (An_{57} - An_{52}) than the phenocryst plagioclases, a result of the more chemically evolved nature of this dolerite.

5.2.3 Plagioclase megacrysts

One of the most striking aspects to have emerged from the petrological studies is the unusual optical zoning displayed by the plagioclase megacrysts. The rocks containing these plagioclase megacrysts were found rarely in situ, more commonly in the river float (Section 3.5.4). Each megacryst appears to be a single radial plagioclase, ranging in size from a couple of centimetres up to 6 cm in diameter, displaying beautiful small scale (>0.1 mm), regular oscillatory zoning (Plate 5.6), which occasionally becomes convoluted (Plate 5.7). The plagioclase contain small clinopyroxene stringers along crystal twinning planes (Plate 5.8).

The anorthite content of these plagioclases $(An_{81}-An_{88})$ is higher than any of the plagioclases probed in other MVG samples (as seen in Figure 5.2), with the exception of one plagioclase with an An_{81} content which was probed in sample ML219 (the only MVG basalt

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Plate 5.6. Spectacular, small scale oscillatory zoning within the plagioclase megacrysts. Field of view 10 x 6.7 mm.



Plate 5.7. Zoning in the plagioclase megacrysts more rarely becomes convoluted. Field of view 3 x 2 mm.



Plate 5.8. Clinopyroxene crystals strung out between the crystal twinning planes in the megacryst. Field of view 3 x 2 mm.



Plate 5.9 Sodic overgrowth rim (An₅₅) surrounds the megacryst which has a more calcic composition (An₈₁ - An₈₈). Field of View 3×2 mm.

plagioclase with an An₈₁ content which was probed in sample ML219 (the only MVG basalt to contain obvious fresh olivines). Therefore, because their anorthite content slightly overlaps with other plagioclases analysed with the basalts I do not consider them to be truly xenocrystic.

The small scale oscillatory zoning in the plagioclase megacryst shows only very minor chemical variation between the lighter more calcic zones and the darker sodic zones. The rim of the megacryst (Plate 5.9) becomes slightly more sodic, varying from An_{81} to An_{55} (across a sub mm scale), and is interpreted to be a later stage overgrowth as this rim value corresponds to the An values of the plagioclases within the surrounding host dolerite (no resorption of the megacryst has occurred). The chemistries of the plagioclase megacryst are discussed in Section 5.3.2 and the mechanics of growth, and its relationships to magma chamber processes are considered further in Section 5.4.

5.3 Mineral chemistry

5.3.1 Chemical mineralogy of the MVG olivines

Although olivine may have been quite common throughout the MVG basalts and obviously occurred as a phenocrysts phase, only one sample was found to contain any fresh olivines. Therefore only tenuous conclusions about the chemistries of the MVG olivines can be reached. However, the sample (ML219) which did contain the fresh olivines was one of two samples with MgO contents greater than 10 wt%.

Figure 5.1 shows that the fresh olivines within ML219 have forsterite contents between Fo₆₉ and Fo₇₆ (probe data in Appendix G). Little variation is seen in forsterite content across individual olivines (\leq 2 Fo unit) and, where seen, the zoning is normal. Minor oxides include CaO, MnO and NiO, but little or no (<0.05%) Cr₂O₃, TiO₂, Al₂O₃, Na₂O and K₂O was detected. CaO, as well as NiO, shows little variation with decreasing Fo content (though CaO possibly shows a slight increase with increasing Fo content). In the most evolved olivines (Fo_{<70}), NiO contents do, however begin to drop off to below 0.05%. MnO and Fo contents show a negative correlation (Figure 5.1). Substitution of Mn²⁺ for Fe²⁺ is likely, following the reasoning that Fe²⁺ contents in the magma increases with evolution of the magma (as shown by the whole-rock values in Figure 6.4).





5.3.2 Chemical mineralogy of the MVG plagioclases

The chemistry of the MVG plagioclases is plotted in Figure 5.2. The first triplot (Figure 5.2a) shows the clear separation in An content between the plagioclase phenocryst (An₆₆ to An₇₈) and groundmass phases (An₄₃ to An₆₄). However, there is a single plagioclase phenocryst analysis which plots down at An₅₅; this analysis is from a later stage overgrowth rim, hence it has a more sodic chemistry and is similar in composition to the groundmass plagioclase analyses.

Figure 5.2b shows the composition of the small plagioclases ophitically enclosed by pyroxene phenocrysts (Plate 5.2). Even though small in size (approximately the same size as the groundmass plagioclases), these plagioclases have An values which overlap with the larger phenocryst values. This implies they are indeed phenocryst plagioclases whose growth was retarded by being entrapped in a pyroxene phenocryst. One of the analyses of these plagioclases has a much lower anorthite content of An₄₄. This analysis was taken from a plagioclase which was partially protruding from its pyroxene oikocryst, towards the rim of the crystal, hinting that the analysis was from a latter stage overgrowth. All analysed plagioclases have very low Or contents (<2 Mol % with the exception of one point $An_{21}Ab_{66}Or_{13}$).

Finally, the plagioclase analyses from the MVG dolerites are plotted in Figure 5.2c. Comparisons with Figures 5.2a and b reveals a complete overlap with both the phenocryst and groundmass plagioclases from the finer grained basalts and the plagioclases analysed in the dolerites. The dolerite plagioclase analyses range from An_{81} to An_{37} , with a single point at $An_{21}Ab_{66}Or_{13}$ (correspondingly there is an increase in Or content with the lower An content plagioclases). The more calcic plagioclases, which overlap with the phenocryst phases in the basalts, are likely therefore to be phenocrysts.

	KW7	KW9	KW6	KW6	KW6	KW6	TLB25	TLB25	ML219
	plag	plag	plag a	plag b	plag c	plag d	plag a	plag b	plag
An (Mol %)	69.1	58.2	74.0	74.0	71.3	74.9	61.4	62.6	81.6
core								/	
	73.1	49.0	65.5	65.5	/5./	70.3	64.9	57.1	78.3
	70.9	39.4	-	-	55.4	71.8	57.9	56.6	-
	-	-	-	-	-	70.7	-	53.9	-
An (Mol %)	-	-	-	-	-	71.3	-	40.7	-
rim									

Table 5.1. Chemical zoning within individual plagioclase crystals.



The chemical zoning within the plagioclases (phenocrysts, groundmass and dolerite phases) is usually normal, though some reversed zoning does occur. Table 5.1 shows a summary of the chemical zoning in the plagioclases. If just core and rim values are considered, all plagioclases, with the exception of KW7 plag, show normal zoning. However, TLB25 plag for instance has a core composition of An₆₁, an An₆₅ value between the core and rim, and a rim value of An₅₈.

The plagioclase megacrysts show higher anorthite contents than is seen in all but the most calcic of plagioclases analysed in the MVG dolerites (Figure 5.2). There is **no** progressive increase in albite content across a 6 cm diameter crystal, until a sub mm sized, higher sodic, overgrowth rim is reached (Figure 5.3). The only chemical zonation is within the small scale oscillatory zoning, Ca rich zones vary between $An_{86} - An_{88}$ contents while the more sodic zones show similarly restricted compositions of $An_{83} - An_{84}$. Table 5.2 shows the results of probe analyses which were taken across some of the oscillatory zoning. The main point that Table 5.2 highlights is the small change in anorthite content between sub mm scale dark and light zones (as seen on the backscatter electron image (BSE) of the electron microprobe which distinguishes chemical changes), never varying more than 2 or 3 Mol % anorthite.

Plagioclase megacryst	An (Mol%)			
lighter (BSE) oscillatory zone	85.5			
dark (BSE) oscillatory zone	82.9			
lighter (BSE) oscillatory zone	87.6			
dark (BSE)oscillatory zone	83.8			
lighter (BSE) oscillatory zone				
dark (BSE)oscillatory zone				
lighter (BSE) oscillatory zone	87.1			
light (BSE) oscillatory zone at rim	85.8			
dark (BSE) oscillatory zone at rim	81.1			
dark (BSE) sodic rich overgrowth rim	65.5			
dark (BSE) sodic rich overgrowth rim				

Table 5.2. Plagioclase Megacryst chemical zoning.

5.3.3 Chemical mineralogy of the MVG pyroxenes

Pyroxene appears as the third phase on the liquidus, after olivine and plagioclase, and is the rarest phenocryst phase in the MVG basalts. The clinopyroxene analysis of the phenocryst bearing basalts are shown in Figure 5.4 and the dolerites in Figure 5.5. In Figure



Figure 5.3

5.4b, all of the pyroxene phenocrysts analysed are calcium-rich augites, with the groundmass pyroxenes showing a spread to more iron rich compositions. The movement from Ca and Mg rich pyroxene compositions to a higher Fe content (i.e. decreasing Mg/Fe ratios) shown by the later stage groundmass pyroxenes is to be expected with increased amounts of fractional crystallisation. The three calcium-poor phenocryst analyses, which overlap with the groundmass analyses, are from the very rims of phenocrysts. These rims typically have more TiO₂, Al₂O₃, MnO and less Cr_2O_3 and are considered to be more evolved overgrowth rims.

The dolerites (Figure 5.5a), compared to the basalts, show further degrees of pyroxene evolution and some overlap with the ferroaugite and subcalcic augite fields, though the majority (>90%) of pyroxenes are augites. The composition of the pyroxenes in the dolerites, as with the plagioclase compositions, overlaps completely with the compositions of the pyroxenes which are phenocryst and groundmass phases in the basalts.

The compositions of the pyroxenes are strongly influenced by their crystallisation rates, an extreme example of which is the Skaergaard intrusion (pyroxene evolution trends of the Skaergaard intrusion are plotted on Figures 5.4 and 5.5 for comparison). Extremely slow crystallisation rates enable stable partitioning between Ca, Mg and Fe, hence allowing the subparallel trends of orthopyroxene and clinopyroxene to develop (e.g. Brown and Vincent, 1963). At a Mg:Fe ratio of 70:30, pigeonite begins to precipitate instead of orthopyroxene, while in such slow crystallisation systems the clinopyroxene trends from diopside to hedenbergite (Gribble and Hall, 1992). Co-precipitation of orthopyroxene and clinopyroxene swith the MVG dolerites is not seen (though one sample, ML219, may contain pigeonite and is discussed later).

Two extreme examples of dolerites are shown in Figures 5.5b and c, ML219 which is about the most primitive basalt we sampled in this study, while at the more evolved end of the MVG tholeiites is KW9. KW9 (Figure 5.5b) shows mostly augite, with some overlap into the subcalcic and ferroaugite fields. The domination of high-Fe pyroxenes is unsurprising considering the high Fe content of the whole rock (Fe₂O_{3(t)} = 18.43 wt%). In comparison, the most primitive sample, ML219 (Figure 5.5c), shows a wide spread from low Fe, high Ca augite, to a single pigeonite analysis. The petrographic texture of this sample is medium grained (\approx 1 mm grain size) with subophitic and intergranular pyroxenes, indicating that cooling

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Figure 5.4a-c



was not exceptionally slow. Seven analysis points (all from smaller, intergranular pyroxenes, i.e. the last to crystallise) are circled on Figure 5.5c. A group of three analyses falls at the high calcium region of the ferroaugite field and they define a trend similar to that of the Skaergaard intrusion, as would be expected in a slowly crystallising system, i.e. trending off towards hedenbergite. However, three analyses plot well below this trend, indicating a trend towards the ferrosilite apex which is more representative of a rapid crystallisation history. As for the single pigeonite analysis, one point alone does not contribute a great deal, except perhaps contradicting the idea of a reasonably rapid crystallisation for ML219 (suggested by the subophitic texture and subcalcic ferroaugite pyroxenes). A hint that co-precipitation of orthopyroxene and clinopyroxenes, although far from large in scale, may have occurred?

The stoichiometric elements in pyroxenes (Fe, Mg, and Ca) can be substituted by a host of minor elements (Deer et al., 1992). The MVG clinopyroxenes include minor oxides such as Cr_2O_3 , MnO, Al₂O3, TiO₂ and Na₂O (Figure 5.6). All the phenocryst pyroxene analysis are shown in open squares on Figure 5.6, and they typically plot at the low clinoferrosilite end (i.e. higher Mg:Fe ratios), overlapping with some dolerite pyroxene analyses. The three analysis of the overgrowth rims of the pyroxene phenocrysts show clinoferrosilite contents of >20% and are clearly precipitated from a latter stage, more evolved melt.

 Cr_2O_3 shows a remarkably sharp decrease after the clinoferrosilite content within the MVG pyroxenes exceeds 20%, a similar trend to that which is seen in several other basalt groups, e.g. Scoresby Sund, Keweenawan and Hawaiian basalts (Larsen et al, 1989; Papike and White, 1979; Fodor et al., 1975). The crystallisation of chrome spinel will strongly deplete the magma with respect to Cr content (e.g. Gibb, 1973), though no chrome spinels were found in the MVG lavas. The lack of any chrome spinel does not rule out its involvement as an early precipitating phase. Chrome spinels are usually found in basalts with Cr > 350 ppm (Sigurdsson and Schilling, 1976), whereas the MVG basalts have Cr values less than 300 ppm. Furthermore, chrome spinels are often found within olivines which, as I have already described, are not common in the MVG basalts and dolerites. Usually, it is the phenocrysts and cores of the larger ophitic texture augites that contain more Cr, which follows reasoning that Cr is removed by the precipitation of chrome spinels. However, rarely the Cr contents within some of the groundmass pyroxenes were also found to have been elevated, e.g. >





Figure 5.6. Minor element variations within the MVG clinopyroxenes. % Ferrosilite is calculated by the oxide analysis of (Fe/Fe+Ca+Mg)*100 (Mol %). The closed squares represent basalt groundmass and all dolerite clinoproxene analyses, while the open squares are phenocrysts. Al and Ti values are calculated to 6 oxygens.

0.1% to 0.35% Cr₂O₃. An increase in the partitioning coefficients for Cr_{augite}/Cr_{liquid} with decreasing temperature (Irvine, 1975) could explain these high Cr content groundmass pyroxenes. A similar reasoning is put forward by Larsen et al. (1989) to explain the high Cr content of the pyroxenes within some of the more evolved basalts from Scoresby Sund. The high Cr contents of the pyroxenes in the evolved basalts, which are probably too evolved to have precipitated chrome spinels, overlaps with the high Cr content of some of the pyroxenes in the more primitive basalts (which have higher MgO and Cr whole rock values and therefore considered to have been crystallised at higher temperatures and hence more likely to have precipitated chrome spinels).

In comparison Mn shows a straightforward positive correlation with increasing clinoferrosilite content. Such a correlation is best explained by simple Mn^{2+} substitution for Fe²⁺ in the pyroxenes. Na content shows little correlation with percentage clinoferrosilite, except perhaps for a slight drop as the magma evolved to lower Mg:Fe ratios.

 Al_2O_3 and TiO₂ appear to have a more complex relationship, not only with magma evolution but also each other. Al substitution into pyroxenes could have been inhibited by plagioclase fractionation, which may offer a possible explanation for the drop in Al_2O_3 with increasing clinoferrosilite content. However, because plagioclase is on the liquidus before pyroxenes and is consequently precipitating out of the magma prior to pyroxene formation, if plagioclase removal was the sole control of the Al content of pyroxenes, it cannot explain a small initial increase in the Al_2O_3 content (i.e. in the low clinoferrosilite region) across some individual pyroxenes probed.

The TiO₂ contents of the pyroxenes are less than 2 wt%, and within the typical range found in augites (i.e. titanaugites usually have TiO₂ contents >3%). TiO₂ shows a similar distribution to Al₂O₃, with an initial rapid increase, followed by a decrease. This suggests that the behaviour of TiO₂ relies on its coupled substitution with Al₂O₃ (Larsen et al., 1989), i.e. the dominant control on Ti is (Fe²⁺, Mg)²⁺ + 2Si = Ti + 2Al (Gibb, 1973; Larsen et al., 1989). In other words, the Al and Ti content of the pyroxenes are both controlled by the Ti content of the liquid. Therefore, a drop in Ti availability levels in the magma, when titanomagnetite reaches the liquidus, could result in inhibited substitution of both Ti and Al into pyroxenes (Gibb, 1973). Because titanomagnetite reaches the liquidus after pyroxene, this would provide a suitable explanation for the initial rapid increase in AI and Ti values in pyroxenes because it isn't until later that the substitution of AI and Ti into pyroxenes is inhibited.

If Al and Ti substitution in pyroxenes was perfect, the Al:Ti ratio would be 2:1 (Figure 5.6). Recent work on alkali basalts has indicated that the pressure of crystallisation has a controlling effect on the Al:Ti ratios in pyroxenes. Thy's (1991a) experimental work on the Al:Ti ratios at various pressures shows extensive variation in the substitution ratios, with higher Al:Ti ratios prevailing at higher pressures. At pressures of 1 atm, pyroxene compositions vary between 4Al:Ti and 2Al:Ti, while at 10 kbars compositions fall between 10Al:Ti and 6Al:Ti, and at higher pressures (12.5 to 20 kbars) between 10Al:Ti and 24Al:Ti. The MVG pyroxene analyses fall between ratios of 2Al:Ti and 10Al:Ti, with the majority between 4Al:Ti and 10Al:Ti (Figure 5.6) The pyroxene phenocrysts in the basalts (open squares) typically display high Al:Ti ratios, indicting they formed at high pressures, possibly as much as 10 kbars. The rest of the dolerite and basalt groundmass pyroxenes (filled squares) show a full range between high and low Al:Ti ratios, these analyses with higher Al:Ti ratios are likely to be from dolerites with pyroxenes which may have been magmatic phenocrysts.

5.3.4 Chemical mineralogy of the MVG opaques

Opaques are found commonly within the groundmass and rarely as microphenocrysts. With only one exception, all of the opaques within the MVG basalts are titanomagnetite (Figure 5.7, filled diamonds), the results of the probe analyses can be found in Appendix G and are plotted on Figure 5.7. The one exception was the discovery of ilmenite needles in the groundmass of ML219 (depicted by open diamonds in Figure 5.7).

The electron microprobe data were recalculated to 4 oxygens for the titanomagnetite and 3 oxygens for the ilmenite. The Fe^{2+}/Fe^{3+} ratios were calculated using the general equation of Droop (1987);

F = 2.X.[1 - (T/S)] where

X - number of oxygens per unit cell
(i.e. 4 in titanomagnetite and 3 in ilmenite)
T - ideal number of cations per unit cell (3 for titanomagnetite and 2 for ilmenite)
S - observed total (per X oxygens)

[NB. The co-existence of titanomagnetite and ilmenite was only found in one sample, therefore no attempts at oxygen fugacity (fO_2) or geothermometry were made].



5.4 Discussion of the petrography and chemical mineralogy *Crystallisation order of the MVG tholeiites*

In the MVG it seems there is a consistent crystallisation sequence:

```
OLIVINE ⇒ PLAGIOCLASE ⇒ CLINOPYROXENE ⇒ Ti MAGNETITE ⇒ ± MINOR PHASES
(+ OLIVINE?) +PLAGIOCLASE (+CLINOPYROXENE?) i.e. Apatite, Quartz
(+OLIVINE?) (+PLAGIOCLASE?) and ilmenite
(CPX/PLAG?)
```

Pressure of formation of clinopyroxenes in the MVG

Figure 5.8 is an expanded portion of Figure 5.6, showing the AI:Ti ratios of the pyroxene phenocrysts in the MVG basalts (open squares) and the rest of the pyroxenes from the dolerites and basalt groundmass (filled squares). In addition, fields from pressure

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experiments on pyroxene AI:Ti ratios are shown (Thy, 1991a), in addition to the augite phenocrysts from the south-eastern rift zone, Sudurland (open circles; Thy, 1983), groundmass augites from Heimaey (striped field; Thy, 1991b), and a representation of the pyroxene probe analyses from the Scoresby Sund tholeiites (grey diamonds; Larsen et al., 1989). All AI and Ti cation values and pressure fields have been recalculated to 6 oxygens.



Figure 5.8

The pressure fields of Thy (1991b) outlined in Figure 5.8 were calculated on alkaline glassy lavas from the 1965 Surtsey eruption, and from experimental results. The Al:Ti ratios of the augite phenocrysts from Sudurland and the groundmass pyroxenes from Heimaey both show lower Al:Ti ratios compared to the MVG phenocrysts. However, the range of Scoresby Sund tholeiites overlaps with the MVG basalts, with perhaps some of the MVG pyroxenes (mainly the phenocrysts) showing slightly higher Al:Ti ratios.

Values of 10AI:Ti for the MVG phenocrysts implies pressure of formation was at a

maximum of 10 kbars, corresponding to depths of 30 km or less, while the majority of pyroxenes were formed at shallower levels as they plot towards the 4AI:Ti line which represents pressure of formation nearer to 1 atm.

Magma chamber processes and plagioclase megacrysts

The fine scale chemical zonation and the large size of the plagioclase megacrysts implies that they spent time in a magma chamber. The small compositional changes which forms the superb oscillatory zoning, and the lack of any gradual change to more sodic compositions towards the rim (except the overgrowth rim) may be a direct reflection of the chemical fluctuations within a magma chamber. These fluctuations may have been only small, varying from more calcic to slightly more sodic at times, always returning to a pre-existing state and causing no major changes (i.e. not greater than a few An mol %). Such chemical changes in the chemistry of a magma chamber may have been caused by replenishment and eruption events, but would they have been so steady and rhythmic and only changed the overall composition of the liquid in the magma chamber by such a small An scale?

Further consideration of the small chemical changes in the oscillatory zoning, and lack of any general trend to greater albite contents could reflect local phenomena due to diffusioncontrolled growth. Allègre et al. (1981) outlined the three variables which would effect the formation of a crystal from a liquid ; (i) the crystallographic features of the growing crystal and any associated interface kinetic effects, (ii) the temperature distribution in the liquid suspending the crystal, and (iii) the element distribution in the liquid. Their conclusions were that interface kinetics are highly sensitive to concentration changes in the melt due to chemical diffusion, not temperature diffusion which is far too efficient to be a limiting factor for the growth rates considered (Allègre et al., 1981; Bottinga et al., 1966) and, hence, the time between interface kinetics and diffusion-controlled growth is on a measurable crystal length. This diffusioncontrolled growth for the plagioclase megacrysts may certainly explain the small changes in An content. If the megacryst resided in an open magma chamber with a steady influx and eruption of magma, this could possibly keep the system at a steady rate until the megacrysts were caught up in an ascending magma and erupted. (A closed system would be expected to undergo gradual fractionation and cause a trend towards more sodic compositions across the megacryst). Also, some of the small plagioclase phenocrysts (though admittedly of lower An

contents) show some oscillatory zoning too, backing up speculation that magma chamber processes and chemical diffusion both are important in controlling the An content of the plagioclases.

Chapter 6

The effect of alteration and element mobility on the Malaita Volcanic Group

6.1 Introduction

Alteration of the MVG is low grade and within the brownstone and zeolite facies. The general characteristics of the alteration and the nature of secondary mineralisation of the MVG basalts has been established via a combination of petrographic, X-ray diffraction (XRD) and electron microprobe studies.

The secondary alteration which has affected the MVG basalts has helped in attempts to establish the extent of tectonic imbrication within the MVG lava pile (Chapter 4). It is also essential to consider any effects the alteration may have had on the whole-rock compositions of the basalts. If the original composition of the MVG basalts has been modified by secondary element mobilisation, this would have far-reaching consequences for the succeeding geochemical studies trying to establish primary magmatic and tectonic processes.

6.2 X-ray diffraction studies of the MVG

XRD analysis (for methods see Appendix B) was carried out on clay separates (<2 μ m) from selected MVG samples from the Singgalo and Kwaleunga Rivers. The aim was to gauge alteration levels within the MVG basalts and establish whether there was any variation/increase in alteration levels with depth in the lava pile. The alteration would hopefully provide vital clues to whether the geochemical cycles seen, for example along the Singgalo River (Section 4.5.2), were the result of magma chamber processes or purely a result of tectonic imbrication.

Clay separates were analysed by XRD in preference to whole-rock samples because there was a tendency for the poorly crystalline clays (e.g. celadonite) to be masked by the peaks generated by the major mineral phases in the basalts. The peaks were measured in terms of 20 (theta) and converted into lattice (d) spacings via:

nλ=2dSinθ

[Bragg equation].

As all minerals possess a unique XRD pattern, identification of peaks, d spacing, and hence minerals was via standards published yearly by the JCPDS-International Centre for Diffraction Data.

The results of the XRD analysis carried out on the clay separates from the MVG basalts are tabulated in Table 6.1, and two examples of the XRD traces (samples TLB20 and TLB24) produced by the MVG clay separates are shown in Figure 6.1.

Sample	Saponite	Celadonite	Zeolite	Chlorite	Notes
Singgalo River	_				
TI B1	\checkmark	\checkmark	x	x	<u>-</u>
TI B7	Ň	Ń	x	×	-
TLB9	, V	×	x	x	-
TLB11	Ň	√ √	x	x	-
TLB12	V	Ń	\checkmark	x	analcite
TLB14	\checkmark	\checkmark	х	x	-
TLB17	\checkmark	\checkmark	х	x	-
TLB19	\checkmark	\checkmark	x	x	-
TLB20	\checkmark	\checkmark	x	x	see Figure 6.1
TLB21	\checkmark	\checkmark	x	x	-
TLB22	\checkmark	N	x	x	-
TLB24	N.	√.	\checkmark	x	analcite (see Figure 6.1)
TLB25	√	\checkmark	х	x	-
TLB26	N	?	х	x	poorly crystalline saponite
TLB27	√.	N	х	x	-
TLB28	N	N	x	x	v. poorly crystalline saponite
TLB31	√,	N	x	х	-
ML680	,	√,	x	x	-
ML684	N	\checkmark	x	x	-
ML687	N	×	x	x	-
ML689	N	N,	x	x	-
ML692	V	٦	x	x	-
Kwaleunga River					
KŴ9	\checkmark	\checkmark	x	x	-
KW13	\checkmark	\checkmark	x	x	-
KW14	\checkmark	\checkmark	x	?	phillipsite possibly present
KW18	\checkmark	\checkmark	x	\checkmark	-
ML696	\checkmark	\checkmark	x	x	-

Table 6.1. Results of the XRD analysis on clay separates

In the MVG clay separates, XRD traces are dominated by intense saponite peaks, corresponding to a d spacing of \approx 15Å which is typical of smectites (12-15Å). On exposure to ethylene glycol (HOCH₂CH₂OH) basal spacing is expanded to 17Å because the larger glycol molecules exchange (mainly with water) into the inter-layer spaces of the smectite,

causing a corresponding peak shift (glycolated traces in Figure 6.1). No such swelling is seen in chlorites or illites when glycolated, an important identification point, especially as chlorites have a d spacing around 14Å which overlaps with smectites. Two heating steps followed glycolation as the exposure of smectites, such as saponite, to temperatures in excess of 300°C causes the smectite structure to collapse to 10Å d spacing, i.e. they collapse to an illite structure.

Two examples of the XRD results are shown in Figure 6.1. Some of the sub-2 µm peaks labelled (i.e. plagioclase and augite) are a product of grinding and not alteration products. TLB20 (plot a) is typical of the majority of the MVG basalts, with saponite displaying the most intense peak, showing typical swelling when glycolated and collapsing to 10Å d spacing when heated. Smaller peaks of celadonite (a member of the same group as illite) are also labelled. It should be mentioned that although the intensity of the peaks (i.e. peak height or area) for celadonite are typically less than for saponite, it does not necessarily imply that the concentrations of saponite exceed celadonite in the samples. The difference in peak intensity may merely reflect difference in degree of crystallinity, and different crystal systems have different diffracting abilities based on mass attenuation characteristics. It could be that celadonite has a poorly crystalline form in the MVG basalts compared to saponite, so caution is important in quantitative analysis of XRD traces. However, in the MVG, the majority of secondary minerals analysed on the electron microprobe (Section 6.3) did comprise saponite, so it appears reasonable to assume that saponite is a dominant alteration product in the MVG basalts.

Plot b in Figure 6.1 is from sample TLB24, one of the few (along with TLB12 and possibly KW14) to show the presence of zeolites in the XRD traces. The zeolites identified are analcite, in TLB24 and TLB12, and possible phillipsite in KW14. Also collected in the field was a large (~5 cm in diameter) zeolite centre found within a pillow lava, identified as thomsonite by XRD whole mineral analyses. Very few zeolites were detected in the MVG basalts. This is surprising as they were often present in amygdales in some pillow lavas. However, our sampling, by its avoidance of the most altered portions of basalt, naturally meant we also avoided sampling zeolites.

The results of the XRD studies on the alteration products of the MVG basalts showed





no increase in the grade of alteration with increasing depth into the lava pile. This lack of any variation in the alteration grade of the MVG is consistent with i) a low geotherm for the MVG basalts (Section 6.5) and/or ii) tectonic imbrication which is preventing us from seeing to depths greater than 2 km into the lava pile (Section 4.5.2).

6.3 Chemistry of the MVG secondary minerals

Several of the MVG basalts and dolerites had their alteration phases probed, the results of which are tabulated in Appendix Gv. Probe analysis of the alteration minerals, whether from mesostasis, olivines or replacement of other mineral phases, typically have totals which are less than the considered perfect "100%". This is a simple matter of the high water content of clay minerals, and the intensity of the probe beam drives off any such volatiles. For this reason, totals of 85% and higher are considered good totals. Analysis with totals lower than this are discarded and not plotted in Figure 6.2. Individual clay minerals are almost impossible to probe, because they comprise inter-stratified and overlapping layers.

The results of the analysis of the secondary clay minerals within the MVG basalts are summarised by plotting on Velde and Meunier (1987) triplots (Figure 6.2). These triplots show the compositional fields of the clay minerals and allow easy identification of secondary phases in the MVG tholeiites. No surprises which were not primarily identified in XRD analysis (Section 6.2) appear on the triplots. The majority of the clays fall within the saponite field, with some spread across the glauconite, phengite and celadonite fields and towards the MR3 and 2R3 join where zeolites and albites typically plot.

Saponite is a magnesium trioctahedral smectite in which the inter-layer cation is predominantly calcium. Saponite is a commonly identified alteration product of olivines, in both hydrothermal and surface weathering systems (Shirozu et al., 1975; Righi and Meunier, 1995). Celadonite, on the other hand, is a member of the illite group (as are glauconite and phengite), and its formation relies on potassium enrichment (typical of illites) and high magnesium contents. Celadonite has an average composition and properties indistinguishable from glauconite, but it commonly forms in basalt environments, i.e. in vesicles or as replacement of ferromagnesian constituents (e.g. Hendricks and Ross, 1941; Deer et al., 1963; Deer et al., 1992). It should be noted that 'sericitization' or the surface weathering of feldspars (typically K

Chapter 6: The effect of alteration and element mobility on the Malaita Volcanic Group



Plate 6.1. The rim of an olivine (ol) being replaced by saponite (sample ML219). Field of view 0.75 x 0.5 mm.



Plate 6.2. Well formed chlorite needles filling an amygdale in a MVG dolerite (sample KW18). Field of view 3 x 2 mm.





Figure 6.2

rich feldspars) produces illites, and illite/smectite mixed layer clays as a result of the breakdown of smectites (Deer et al., 1992). However, celadonite is reported within secondary veins from the OJP lavas drilled on ODP Leg 130 (Mahoney et al., 1993). Because the basalts drilled by the ODP have remained in a submarine environment, it is likely the trend towards the celadonite compositions of some of the MVG alteration products is a result of submarine processes, though subaerial weathering of the MVG basalts cannot be completely ruled out.

Both saponite and celadonite are low-grade alteration mineral assemblages (e.g. Wood et al., 1976; Inoune, 1995), common in the brownstone and zeolite grades of alteration (Cann, 1979). The dominance of saponite as an alteration product within the MVG lavas ties in with the suggestion that the secondary alteration in the MVG lavas is largely due to the replacement of olivine (Plate 6.1) and glass (plus smaller extents of plagioclase alteration). Other alteration analyses plotted in Figure 6.2 (KW18, KW13, KW7) are from altered glass and feldspars and appear to represent a trend towards various percentages of mixed-layer saponite/celadonite clays. The last glass to freeze within tholeiites is typically enriched in K and Si (Cox et al., 1993). This potassium enrichment in glass would certainly facilitate formation of illite group minerals, such as celadonite. In the case of feldspars to form illite group minerals, so the formation of celadonite may require addition of K⁺ from seawater. In addition, one sample, KW18, shows the probe analyses from an amygdale fill (Plate 6.2), an example of the rare occurrence of chlorite in the MVG lavas.

6.4 Secondary element mobilisation affecting the MVG

6.4.1 Introduction

Before a detailed study of the magmatic geochemistry of the MVG is undertaken, assessment of any secondary alteration processes needs to be established. Screening the MVG, indeed any suite of rocks, for the effects of post-magmatic alteration is essential. The original composition of lavas may have been modified as a result of element mobilisation, having far-reaching consequences for establishing primary magmatic and tectonic processes.

The mineralogy of the MVG (as outlined in Chapter 5 and Sections 6.2 and 6.3)

indicates that although alteration has occurred, it is largely restricted to the replacement of the mesostasis by secondary smectites such as saponite (and possibly celadonite and/or celadonite/saponite mixed layers). Olivines, when present, have almost entirely been pseudomorphed by secondary saponite while the clinopyroxenes and feldspars (both phenocrysts and groundmass phases) are only very rarely altered. The secondary alteration is therefore almost exclusively smectite mineralization, with only rare zeolites (mostly veins and centres) and chlorite-filled amygdales. It therefore appears the MVG only reached the lowest of alteration grades, predominantly brownstone facies (Cann, 1979) and at a maximum into the zeolite facies.

The following sections are subdivided on the basis of elemental groups which behave similarly within the extent of alteration that has affected the MVG lavas. Trace and major elements were measured on whole-rock powders on either a Philips PW1400 or an ARL 8420+ wavelength-dispersive XRF-spectrometer. Sample preparation, machine conditions, standard controls and data quality are all outlined in Appendices B and C. The full sets of XRF major and trace element results from the Malaita (plus Makira and Ulawa) samples can be found in Appendix D.

A number of samples were selected for rare-earth element (La, Ce, Nd, Sm, Eu, Gd, Tb, Yb, Lu) plus Th, Ta, Hf, U, Sc and Co analysis via instrumental neutron activation analysis (INAA). The counting and standard results used in the INAA analyses are outlined in Appendices B and C, the full set of INAA results for the Malaita samples can been found in Appendix E.

Radiogenic isotope analyses for selected samples were carried out at NERC Isotope Geosciences Laboratories (NIGL), Keyworth. The laboratory techniques are described in Appendix B. Isotope systems Sr/Rb, Nd/Sm and U/Pb were determined on two multi-collector mass spectrometers, a Finnigan MAT 262 and a VG 354.

Twelve MVG samples were analysed for platinium group elements, Au, Re, Ni and Cu at the Australian National University in Canberra using ICP-MS techniques, the methods are outlined in Appendix B.

6.4.2 High field strength elements (i.e. Nb, Zr, Y, P, Ti, Ta, Hf)

There is a consensus that high field strength elements (HFSE) are immobile in low to
moderate alteration grades (e.g. Cann, 1970; Hart et al., 1974; Wood et al., 1976; Floyd and Tarney, 1978; Ludden et al., 1982; Gibson et al., 1982). For example, in the Icelandic Research Drilling Programme (IDPR) the deepest lavas recovered, which were extensively epidotized and into greenschist facies, have retained their original magmatic HFSE concentrations (Gibson et al., 1982). While the Icelandic lavas can be considered geologically young, comparison with far more ancient complexes, for example Archaean greenstone belts, is useful. The Abitibi Greenstone Belt, containing lavas which have undergone metamorphic grades of zeolite to lower amphibolite facies, represents a far more extreme case for element mobility studies. At the lower grades of metamorphism, i.e. zeolite facies, HFSE and rare-earth elements (REE) are considered immobile in the Abitibi Greenstone Belt basalts (Ludden et al., 1982). Even at amphibolite grade the lavas have retained their magmatic Zr/Y and Zr/Ti ratios, although the REE have been leached and redeposited (Ludden et al., 1982). The epidotization which has affected the Icelandic basalts and the amphibolite facies metamorphic grades within the Abitibi greenstone lavas both represent far more extreme alteration grades than has effected the MVG lavas. Bivariate plots of HFSE values for the MVG lavas are shown in Figure 6.3. The tight - generally colinear - data arrays shown by the MVG indicate that the HFSE have remained unaffected and as such represent primary igneous values. In the following account I have used Zr as an index of fractionation because of, (i) its immobility during alteration and (ii) its high abundances which thus avoid analytical uncertainly. Variations in HFSE ratios are good indicators for establishing different magmatic sources or degrees of partial melting.

6.4.3 Major elements (Si, Fe, Al, Ca, Mg, Mn)

Major elements in the MVG plotted in Figure 6.4 show that AI, Mg and Fe appear to have suffered little or no modification in the MVG lavas, Ca may have suffered only minor alteration effects whereas Si and Mn show the most variation due to secondary processes. Al concentrations are controlled by plagioclase, therefore accumulation of plagioclase may cause a spread in AI values which may be confused with secondary alteration (Morrison and Thompson, 1983). Denoted by grey symbols in Figure 6.4 are several samples which showed anomalously high AI and Si values and low Mg and Fe values. These samples appear in microscope studies to contain high percentages of plagioclase crystals, possibly



Figure 6.3. HFS elements, such as Zr, Ti, P, Y, Nb, Ta, Hf, are immobile in the low alteration grades which have affected the MVG basalts. This immobility of the HFS elements provides us, not only with an essential gauge in assessing the extent of possibly mobility in other elements, but is also an understanding of the fractionation history and, possibly, the source characteristics of the lavas.

Figure 6.3



explaining some of the anomalous spread of data.

6.4.4 Transition metal elements (V, Zn, Ni, Cr, Cu)

Transition metal elements (TME) may be mobile, especially at higher temperatures (Seewald and Seyfried, 1990; Rollinson, 1993). In the MVG basalts secondary alteration of Ni and V is minor, Cr and Zn show slightly more secondary spread, while Cu varies the most (Figure 6.5). The concentrations of Ni, Cu and Cr decrease with increasing Zr abundances, similar to observations made by Gibson et al. (1982) in the Icelandic basalts. Fractionation of olivine and augite is likely to be responsible for this trend but, as suggested by Gibson et al. (1982), the original variations in Cr, Ni, and Cu are difficult to estimate and hence secondary alteration more difficult to distinguish from original magmatic signatures. In contrast the Zn and V values increase the more evolved the lavas become, implying Zn and V behaved incompatibly with distribution coefficients remaining low and unchanged as fractionation proceeded, similar to the HFSE.

6.4.5 Large ion lithophile elements (Sr, Rb, Ba, K, Na)

The large ion lithophile elements (LILE), although mostly incompatible in basaltic melt systematics, are more prone to secondary alteration because of their large size and low ionic charge/potential (have small size and high ionic charge compared to HFSE) and hence, their easily soluble state. Unsurprisingly, the Cretaceous MVG basalts (Figure 6.6) show some degree of secondary alteration which has resulted in the enrichment of the LILE. It is well established that low temperature seawater exchange in basalts enriches concentrations of LILE such as Ba, K and Rb (e.g. Hart et al., 1974), while higher temperature alteration is thought to result in the leaching of LILE (Saunders et al., 1979; Stern and Elthon, 1979). No such high temperature LILE leaching has obviously occurred in the MVG lavas, and the majority of the basalts (>90%) show low values for LILE which represent primary magmatic values.

Sr has the ability to substitute for Ca in plagioclase, therefore care must be taken when discriminating between secondary alteration of Sr values and the results of feldspar accumulation. The samples which showed major-element variations because of large amounts of plagioclase (Figure 6.4) are enriched in Sr, suggesting that some of the spread in Sr may have been the result of excessive plagioclase fractionating prior to eruption, not purely



Figure 6.5. The transition metals show some secondary mobilisation, especially in the case of Cu. Ni, Cr and Cu all behave similarly with respect to Zr, probably due to the fractionation of olivine and augite. Zn and V show an opposite behaviour compared to Ni, Cr and Cu, acting incompatibly and increasing in concentration as the magma evolves (as do the HFS elements).

Figure 6.5



secondary alteration. However, in the same samples Rb, K and Na values are also elevated implying secondary processes may have also caused the dispersal of Sr, Na, K and Rb abundances; for example, the presence of zeolites, such as analcite and thomsonite, which contain high concentrations of Sr, Ba, Rb and K, may elevate the concentrations of these elements (Berger et al., 1988; Wood et al., 1976). Sample TLB12 (Figure 6.9) contains secondary analcite veins (determined though optical and XRD methods) and shows clear enrichment in Sr (304 ppm), Ba (123 ppm) and K (K₂O=0.543 wt %). Wood et al. (1976) noted that in the Icelandic lavas, intense zeolitisation strongly decreased concentrations of SiO₂ and MgO. I see no such reductions in SiO₂ and MgO concentrations within the MVG lavas. This leads me to conclude that, while trace element abundances (i.e. Rb, Sr, Ba) in the MVG may have been affected by secondary zeolitisation, the major elements (with the exception of Na and K) have remained largely unaffected.

Approximately ninety percent of the MVG lavas have total loss on ignition (L.O.I.), with values of 1.5 wt% or less, most frequently ranging between 0.5 and 1 wt % (Figure 6.7). There is no obvious correlation between increasing L.O.I. values and concentration of LILE elements (Figure 6.7). In fact, K, Na, Rb and Ba all show some secondary enrichment over the entire range of L.O.I. values, with some of the higher LOI basalts (>1.5 wt%) apparently preserving low, primary magmatic LILE values.

6.4.6 Rare-earth elements

The mobility of rare-earth elements (REE) in low grade metamorphic rocks has been addressed in numerous papers (e.g., Wood et al., 1976; Ludden and Thompson, 1979; Ludden et al., 1982; Gibson et al., 1982; Dostal and Strong, 1983; Bienvenu et al., 1990; Marsh, 1991). REE, being highly to moderately incompatible, are usually concentrated in the mesostasis and thus have the potential to be modified (e.g. Humphris et al., 1984). Therefore, it follows that REE mobility often has more to do with the crystallisation history of the rock, rather than the metamorphic grade reached. For example, a glassy basalt is far more likely to have suffered modification of REE abundances than would a highly crystalline basalt. REE mobility is dependent not only on the ability of the fluid phases to extract REEs from the altering glass/crystals, but also the retention capacities of the secondary minerals formed (Humphris, 1984). Thus, if any secondary alteration has occurred, as with the MVG basalts,



Figure 6.7

care must be taken to establish whether or not the REE have been changed.

Wood et al. (1976) claimed that the concentrations of light REE in the Icelandic basalts may have been mobilised to some extent by zeolite facies metamorphism, i.e. Ce/Yb was modified but Yb and Lu remained unaffected. In contrast Gibson et al. (1982) concluded that in the zeolite facies, and the higher grade greenschist facies basalts, the REEs in the Icelandic basalts have suffered no modification. Gibson et al. (1982) compared multiple samples taken from the same lava flow at the various metamorphic grades. The lack of variation in the REE and HFSE concentrations leads to the conclusion that most variations in Ce/Yb in the Iceland basalts could be explained by igneous processes alone. Supporting Gibson et al. (1982), Ludden et al. (1982) claim that the REE have remained immobile in the low grade metamorphic grade lavas of the Abitibi greenstone belt. Only the highest grade rocks in the Abitibi greenstone, which have suffered extensive alteration by carbonate-rich and K-rich fluids, have had their REE elements affected (Ludden et al., 1982).

More recently, Bienvenu et al. (1990) compared fresh and seawater-altered basalt glasses. With the use of extended "Coryell-Masuda diagrams" (e.g. Bougault and Treuil, 1980) they concluded that changes in Sm/Zr or Sm/Hf (Figures 1 and 2, Bienvenu et al., 1990) are an indication of alteration, as too are anomalously high Ce values in REE patterns. They found that the REE abundances in the basalt glasses, which had suffered secondary alteration, had been modified, while HFSE ratios such as Hf/Zr remain unaffected throughout. They concluded, therefore, that HFSE ratios such as Ta/Hf and Nb/Zr be used for studying magmatic processes in preference to the REE (Bienvenu et al., 1990). Bienvenu et al. (1990) also carried out analysis on the secondary smectites, concluding that the smectites did not quantitatively retain the REE, although Ta/Hf ratios in the smectites were the same as in the fresh basalts.

The MVG basalts, while having suffered replacement of their mesostasis by clays have apparently suffered from no secondary REE mobilisation. Figure 6.8 shows that when various REEs (both light and heavy) are plotted against Zr, concentrations systematically increase. Constant REE/Zr ratios is consistent with the REE having remained immobile. Figure 6.9 compares (La/Sm)_n (normalised to McDonough and Sun, 1995, C1 chondrite values) with HFSE, such as Zr and Ti, which are assumed to be incompatible. Sample TLB12 is





Figure 6.9

highlighted on Figure 6.9 because it contains secondary zeolite veins (thus explaining the elevated Sr and Ba values). TLB12 also has a low $(La/Sm)_n$ value, but this is likely to be consistent with a primary magmatic value as the Zr and TiO₂ concentrations of TLB12 are accordingly depleted. Finally, the MVG display sub-parallel REE patterns (Section 7.5) which are also consistent with a lack of alteration (Sun and Nesbitt, 1978).

The lack of secondary alteration of the REE in the MVG basalts, even though the basalts have suffered replacement of any original glass by smectites, is likely to have been the result of several factors. Firstly, a low temperature and a low fluid/rock ratio may not have been sufficient to cause REE mobilisation and removal. Thus, the smectites may have quantitatively retained the REE originally in the glasses (c.f. Bienvenu et al., 1990). Secondly, the MVG lavas originally contained only minor amounts of glass, and those MVG basalts with higher percentages of glass showed no differences in their REE/Zr ratios than the more crystalline dolerites analysed. It also should be considered therefore that the REE may not have been solely concentrated in the final freezing glass of the MVG lavas, minerals such as augite and minor mineral phases such as apatite will contain a high percentage of REE too.

6.4.7 Radiogenic isotopes

It follows, because of the involvement of such elements as Rb, Sr, Pb, U and Th, radiogenic isotope ratios may be affected by secondary processes. The effects of secondary alteration on the isotope systematics in altered oceanic plateau lavas have been summarised by Mahoney (1987), although he doesn't include Pb isotopes in the study. As with all submarine basalts, seawater can modify ⁸⁷Sr/⁸⁶Sr values and addition of Rb by low temperature alteration has the same effect (Staudigel and Hart, 1981). Sm-Nd isotopes do not seem to be affected to the same extent through alteration (Mahoney, 1987; Castillo et al., 1991; Mahoney et al., 1993), though some modification can occur in the Sm and Nd abundances. Results of previous studies on the Pb isotopes of the OJP lavas show, as with Nd-Sm isotopes, that they may have suffered very little modification (Mahoney and Spencer, 1991; Mahoney et al., 1993).

ICPMS analyses of MVG basalts for U, Th, Pb and Zr, carried out at the Australian National University, Canberra, are shown in Figure 6.10. Pb concentrations show a spread of data indicating Pb has been affected by secondary processes. However U and Th, when





Figure 6.10

plotted against Zr (Figure 6.10), show remarkably consistent ratios, implying that, while Pb was apparently extensively mobilised, whole-rock concentrations of U and Th were not obviously modified.

Samples analysed for Pb, Sm-Nd and Sr-Rb isotopes were subjected to leaching experiments prior to isotope dilution processes (the isotope dilution and leaching methods are described in Appendix B). The results of these experiments are shown in Figure 6.11, with individual samples being joined by tie lines.

The Sr isotope values of the MVG are typically lowered by the leaching experiments (with the exception of KW13), though the shift was small in comparison with previous studies which used multi-step acid leaching experiments (in this study only one-step leaching was carried out) for the Sr isotope determination (e.g. Mahoney, 1987). The age-corrected values of the unleached samples in Figure 6.11a show a substantial shift to lower ⁸⁷Sr/⁸⁶Sr values. The leached samples are assumed to contain no Rb (because the leached residue consisted almost entirely of plagioclase and clinopyroxene), and therefore present day ⁸⁷Sr/⁸⁶Sr values are approximate age-corrected values.

The most interesting aspect from leaching experiments has arisen from the Pb isotope results. Mahoney and Spencer (1991) and Mahoney et al. (1993) recorded in their early leaching experiments on the OJP basalts (both drilled and recovered from Malaita) that shifts in the Pb isotopes were minor in leached, compared to unleached, samples and thus concluded that Pb isotopes had suffered very little secondary modification. In this study a considerable shift is seen towards more radiogenic Pb isotope values (Figures 6.11b and c), leading me to conclude that the Pb isotopes (as well as Pb abundances as shown in Figure 6.10) have in fact been modified by seawater alteration. This modification is far from insignificant, the shift experienced by the MVG basalts in this study has placed some the samples into a different isotope group than is suggested by their Sr and Nd isotopes (Section 7.6). Previous leaching experiments on Pb isotopes of the MVG basalts by Mahoney and Spencer (1991) and Mahoney et al. (1993), however, did use a more selective method of hand picking mm-sized chips and weak acid/water washing procedures, prior to crushing, which may offer one explanation for the lack of any Pb isotope shift in their study. The initial discarding of any alteration veins, etc., in this study was limited to mm to cm scale chips, and





samples were washed only in de-ionised water prior to crushing.

6.4.8 Platinum group elements (Pd, Pt, Rh, Ru, Ir)

The platinum group elements (PGEs), plus Au, Re, Ni and Cu, for the MVG basalts are plotted in Figure 6.12. PGE values tend to decrease in the more evolved basalts. All of the PGEs and Au can be mobilised under certain metasomatic/hydrothermal conditions (e.g. MacDonald, 1987; Barnes et al., 1985); with the exception of one sample showing excessively high Re concentrations, this does not appear to be the case in the MVG basalts. Sample KW13, as highlighted in Figure 6.12, seems to show high concentrations in Au, Pd, Pt, Rh, Ir and Pd/Ir. However, KW13 has not undergone PGE modification, and the high PGE values are considered primary, for reasons discussed in Section 7.8.

6.5 Conclusions of alteration studies of the MVG basalts

The MVG only reached the lowest alteration grades, possibly into the lower part of the zeolite facies, though most of the alteration is restricted to smectite replacement of the mesostasis. (Though it is still unclear how much of the low alteration grade which affects the MVG lavas is indeed the a result of a low geotherm or due to the imbricated nature of the basalt pile, thus both must be considered).

A low geothermal gradient in the MVG basalts may have been the result of lavas erupting quickly and being removed from any intense hydrothermal activity associated with ridge volcanism. Field studies of the MVG lavas failed to reveal any wide-scale veining which would have been expected if the lavas had been subjected to extensive hydrothermal activity. It is possible that the MVG lavas are either the result of distal eruptions, away from the ridge and its intense hydrothermal systems, or the accretion rate was such that the lavas were moved away from the ridge rapidly before major hydrothermal systems could be established. Rapid eruption, quick removal from intense ridge activity and lack of intercalated sediments could have resulted in a low water/rock ratio and restricted pore water exchange (Batiza et al., 1981) within the MVG.

The alteration grade reached in the MVG is far less than that calculated for Icelandic basalts. In the samples recovered from the IRDP, it is at depths of 2400 m into the lava pile that abundant epidote appears, the result of burial metamorphism (though the proximal



Figure 6.12. PGE, Au, Re, Ni and Cu abundances for the MVG basalts above appear to show little or no secondary alteration. As the lavas become more evolved the abundances of PGEs decrease, the exception to this is sample KW13 (ringed).

Thingmuli centre caused small scale introduction of epidote before this depth because of geothermal systems (Mehegan et al., 1982)). Mehegan et al. (1982) assume a temperature of 260°C for the top of the epidote zone, and therefore estimate a geothermal gradient in excess of 100°C/km. The MVG basalts reach at most zeolite facies (>50-100°C and <250°C), and, assuming no tectonic imbrication, we can estimate that the lava pile in the Kwaio Inlier is at least approximately 3 km thick (Section 4.5.2). But if the MVG basalt pile is imbricated I may have only been seeing the upper 1 km, possibly 2 km of the lava pile. Therefore, if I assume that a 1 to 2 km section of the MVG reaches zeolite facies metamorphism, then based on the Icelandic lavas which reach a similar grade of metamorphism after approximately 600m (Gibson et al., 1982; Mehegan et al., 1982) the geothermal gradient of the MVG could range from 60°C/km (assuming a lava pile of 1 km) down to a minimum of 30°C/km (for 2 km depth). Such a low geotherm would explain why all but a few element abundances in the MVG basalts have retained their original magmatic values.

Chapter 7 Petrogenesis of the Malaita Volcanic Group

7.1 Introduction

This chapter documents the geochemical characteristics of the MVG and considers the magmatic plumbing processes which affected the MVG basalts after mantle melting. Understanding the nature of fractionation and magma chamber processes prior to eruption is vital before any attempts to understand mantle source region signatures and melting processes are made (Chapter 8).

Previous workers have tended to subdivide the OJP basalts into two groups, Units A and C-G, after ODP Leg 130 Site 807 which recovered and documented basalts with geochemically distinct signatures (Mahoney et al., 1993). However, the division between the two groups which is seen clearly in the basalts from Site 807, is often ambiguous in the major and trace elements of the MVG and is only obvious in the rare-earth elements and isotopic signatures. Therefore, no attempt is made to subdivide the MVG basalts until the latter half of this chapter when the REEs, isotopes and PGEs are discussed.

7.2 Classification of the MVG

The MVG lavas plot in a restricted field on the total-alkali versus silica (TAS) diagram (Figure 7.1a), well within the basalt field (nomenclature boundaries from Le Maitre et al., 1989). The samples have been screened for secondary alteration; i.e. samples with Na₂O>2.5 wt% and K₂O>0.3 wt% have been removed as these LILE elements have been enriched due to secondary processes (Section 6.4.5). Further classification plots, SiO₂ vs. K₂O and Al₂O₃ vs. alkali index, confirm the MVG lavas as low-K tholeiites (Figure 7.1b and c).

The screening of MVG samples as described above, i.e. samples with Na₂O>2.5 wt% and K₂O>0.3 wt%, is assumed from this point on, unless stated otherwise. One exception is the inclusion of samples MB2 (Na₂O = 3.31 wt% and K₂O = 0.29 wt%) and



Figure 7.1a-c. The MVG lavas, after basic screening to remove altered samples, fall in a tight array within the low K, tholeiitic basalt range in the various classification plots shown above. The TAS diagram includes a shaded field representing the Icelandic Tertiary basalts (J.G. Fitton and B.S. Hardarson, unpublished data), which covers a wider range than the MVG basalts, for comparison. In plot a, the TAS boundaries are from Le Maitre et al. (1989); in plots b and c the boundaries are from Middlemost (1975).

Figure 7.1a-c

ML348 (Na₂O = 2.69 wt% and K₂O = 0.14 wt%) because MB2 is used in calculations of liquid lines of decent, and ML348 is one of the most evolved basalts sampled.

The CIPW normative mineralogy of the MVG lavas were calculated using the 'CIPW norms 3.2' computer program (D.R. Mason, 1990), assuming a ratio (by mass) of Fe₂O₃/FeO = 0.15. The full results of the CIPW norm calculations for the MVG basalts are included in Appendix Di and are shown in the Diopside(*di*)-Hypersthene(*hy*)-Olivine(*ol*)-Nepheline(*ne*)-Quartz(*qz*) plot (Figure 7.2, after Thompson, 1982; 1984). The CIPW norms are extremely sensitive to Na₂O and CaO values and hence secondary alteration can cause considerable spread in the data. The altered samples (not shown in Figure 7.2) have higher *di/hy*, possibly because of some small scale carbonate addition. The majority of secondary replacement in the MVG basalts is by smectite, however, which would be more likely to trend towards higher *ol* values (i.e. MgO addition); indeed, some of the more altered, excluded samples do show a spread towards high *ol*.

The *di-hy-ol-ne-qz* plot in Figure 7.2a outlines a classification of lavas, which is determined by the portions of their normative minerals (Yoder and Tilley, 1962; Cox et al, 1993). The results of the normative calculations show that over 95% of the MVG are *qz* tholeiites. They are silica oversaturated, typical of a low pressure fractionating system (Thompson, 1972, 1973). Unusually, the MVG basalts plot in an array sub-parallel to the *di-hy* divide (Figure 7.2b), very different to more typical basalts systems, e.g. Snaefell Volcanics (Hards, 1995), Scoresby Sund basalts (Larsen et al., 1989) and the Icelandic Tertiary tholeiites (J.G. Fitton and B.S. Hardarson, unpublished data), all of which plot along a low pressure fractionation array sub-parallel to the 1 atm line. This unusual trend of the MVG basalts along the *di-hy* divide may be the result of fractionation at various pressures, i.e. basalts which formed at higher pressures have higher percentages of *hy*. Figure 7.3a-e shows some of the various geochemical changes with the changing % *hy* (distance along the *di-hy* divide)

% hy = [hy/(hy+di+qz+ol)]x100

and silica index (extent of silica saturation, see Fitton et al., 1991) where:

S.I. = $(Si-(AI+Fe^{[2+]}+Mg+3Ca+11Na+11K+Mn-Fe^{[3+]}-Ti-4P)/2)x100$

[NB. For the S.I. calculations Al, Si, Mg, etc., are the weight percent oxides divided by their



Figure 7.2. The di-hy-ol-ne-qz diagrams (Thompson, 1982) in plot a, show the CIPW normative classification of lavas (Cox et al., 1993) and in plot b, the position of the MVG tholeiites (•). The 1 atm and 9 kbar lines shown are from experimental studies, the arrows indicate falling temperature and the dashed portions of the lines have yet to be accurately determined (Thompson, 1972, 1973). Results of "fertile" dry mantle and MORB source melting experiments at various pressures have also been included, and the MgO wt% of the initial melts at these pressures are shown in brackets (Thompson, 1982, 1984). The CMAS synthetic system cotectics are included for comparison because some of the MVG basalts plot close to the 9 kbar end.

Figure 7.2a&b

respective molecular weights (Fitton et al., 1991) and $Fe_2O_3/(Fe_2O_3+FeO) = 0.13$].

The slight increase in silica saturation index with increasing % *hy* (Figure 7.3a) is a direct reflection on the movement away from the *di-hy* divide and towards the *qz* apex. The MVG basalts which fall further towards *hy*, also see an increase in Fe_{8.0} content (calculated as in Klein and Langmuir, 1987) and a slight increase in MgO values (Figure 7.3b and c). The positive correlation between Fe_{8.0} and *hy* content suggests that the higher % *hy* basalts formed at greater depths of melting (Klein and Langmuir, 1987), a point which is also indicated by higher *hy* content basalts plotting close to the CMAS 9 kbar cotectic. However, neither Mg number, Ni concentrations (Figure 7.3d and e), trace element concentrations nor ratios, suggest that the higher % *hy* might have been the result of higher pressure melts and this is discussed further in Section 7.4.

7.3 Major and trace element chemistry of the MVG

The range of major and trace element data for the MVG basalts is shown in Figures 7.4a-i and Figures 7.5a-j, with Mg number used as a fractionation index:

Mg number = $[(MgO/40.304)/((MgO/40.304) + (Fe_2O_{3(t)}/90.622))]x100$

when $Fe_2O_3/FeO=0.15$.

For comparison, the data fields of the Nauru Basin lavas (Saunders, 1985) and the ODP Leg 130 Site 807, 122 m.y. tholeiites (Mahoney et al., 1993) have been included. Mahoney et al. (1993) showed that the Site 807 basalts could be separated into two groups, Unit A and Unit C-G, largely because of different radiogenic isotope ratios (Section 7.6) and REE abundances (Section 7.5). These two distinct basalt groups show little variation in the majority of their major and trace elements, though Unit C-G basalts are generally more primitive than Unit A basalts; they, therefore, have been grouped together for convenience in this section and in Figures 7.4 and 7.5.

The MVG tholeiites vary between relatively primitive (though none of the samples is considered to represent a primary mantle melt) to highly fractionated basalts, i.e. MgO varies between 10.38 and 4.93 wt%, Mg number ranges from 65.6 and 35.3, and Ni between 160 ppm and zero. All of the major elements analysed in this study, plotted in Figures 7.4a-i, overlap and extend beyond the limits outlined by the ODP Site 807 lavas (with the possible



Figure 7.3a-e. The positive correlation between $Fe_{8,0}$ of the MVG basalts indicates that higher pressure of melting was responsible for the increase in normative hypersthene content. However, neither Mg number, Ni nor trace element contents show any indication this may have been the case.

exception of Na₂O and K₂O). Nevertheless, the vast majority of samples (>90%) display restricted major element compositions, e.g. TiO₂ (1.1 - 2.1 wt%), Al₂O₃ (12.8 - 15.5 wt%), total Fe₂O_{3t} (13.4 -14.6 wt%), CaO (9.7 - 12.5 wt%) and P₂O₅ (0.09 - 0.17 wt%), with a few more evolved basalts showing TiO₂ >2.3 wt%, Al₂O₃ <12 wt%, Fe₂O_{3t} >18 wt%, CaO <9 wt% and P₂O₅ >0.2 wt%.

Liquid lines of descent (LLD) from various possible 'end member' samples are also shown on Figures 7.4 and 7.5. They were calculated using the TRACE3 program of Nielsen (1988), which assumes a pressure of 1 atm (details of the program are described further in Section 7.4). The system was set for no recharge, fractional crystallisation (i.e. closed system), and fO₂ was set at 1 log unit below the QFM buffer. ML219 and ML216 were chosen as starting magmas because they are the most primitive basalts collected in this study, with MgO contents of 10.02 and 10.38 wt% and Mg numbers of 65.57 and 64.03, respectively. A LLD was also calculated for sample MB2 because it has a high Mg number of 67.13, although this high Mg number is partly a result of its low Fe₂O_{3t} value (Fe₂O_{3t}=8.49 wt%, MgO = 7.72 wt%). Furthermore, MB2 was calculated as a starting composition because it is from an isotopically different group (i.e. affinities with Unit A basalts) than ML219 (affinities with Unit C-G basalts). CUR7 and BE148 are both picrites from the Curaçao Lava Formation which represents an imbricated portion of the Cretaceous Caribbean Plateau (Klaver, 1987; Kerr et al., 1996). CUR7 and BE148 have MgO contents of 19.99 and 18.48 wt%, respectively; both represent a far more realistic primary liquid, or end member, for LLDs and, although they are from a different plume source region, are included for comparison.

The major elements of MVG tholeiites show reasonably well defined LLDs due to low pressure fractionation (i.e. olivine \pm plagioclase \pm clinopyroxene, Section 7.4), especially pronounced in TiO₂, Fe₂O_{3t} and P₂O₅, which fit well with the LLDs calculated for ML219 and ML216 at 1 atm. The modelling suggests that, starting from a basalt of composition ML219, up to 44% of crystals have been removed in order to produce the more evolved, lower Mg number samples (i.e. Mg numbers of 47). The two most evolved samples, KW9 (Mg no. = 41) and ML348 (Mg no. = 35), require the removal of 56% and 64% of crystals, respectively.

None the less, there is some spread of data, most pronounced in CaO, Na₂O and K_2O which indicates that although fractional crystallisation was an important process, it does



Figure 7.4a-c



Figure 7.4d-f



Figure 7.4g-i

not entirely explain the compositional spread seen in the MVG basalts. This may be due to the involvement of different source regions, i.e. clusters of overlapping LLDs in the MVG data set which cannot be related to an individual parental magma. Equally, this spread in some of the MVG data may be the result of magma chamber processes, i.e. repeated tapping, replenishment and magma mixing. Furthermore, TRACE3 is restricted to modelling fractionation at pressures of 1 atm, and it is likely at least some of the fractionation of the MVG lavas occurred at higher pressures, as suggested by the CIPW norms.

Trace elements are plotted against Mg number in Figures 7.5a-j. Ni and Cr values decrease dramatically with increasing extents of fractionation (Figure 7.5a and b), which is to be expected because Ni is strongly partitioned into olivine and clinopyroxene, and Cr is strongly partitioned into clinopyroxene and chrome spinel. Ni does not show such a good fit with the LLD when compared with Cr, tending to plot along a linear trend, rather than the calculated LLD hyperbola. This may reflect mixing between two different parental magmas prior to eruption, or mixing between evolved and primitive magmas during magma chamber processes (i.e. RTF).

V, Zr, Y and Nb (less so Sr) all show an increase in abundance in the basalts with increasing extent of fractionation (Figure 7.5c-f), consistent with the elements behaving incompatibly. Nb and Zr show more deviation from the LLDs than Y, this may be due to Y being slightly less incompatible in its nature than Nb and Zr, i.e. D values of Nb, Zr and Y in clinopyroxene being 0.005, 0.1 and 0.9 respectively (Rollinson, 1993; Green, 1994). Nb and Zr, in comparison with Y, also show a gradual trend away from the calculated LLD with increasing extents of fractionation. Again, as with some of the major elements, the spread in the MVG composition cannot be fully explained by low pressure fractional crystallisation alone, implying the involvement of different magma sources or fractionation at different pressures.

Rb appears to show little change (Figure 7.5h), except perhaps for a slight increase with decreasing Mg number values which is consistent with it behaving incompatibly. Rb, like Ba and Sr, is highly prone to secondary alteration, as shown by the spread in the ODP Site 803 samples (they have not been screened like the MVG basalts). Sc shows little change in abundance with increased evolution of the magma (Figure 7.5i), as predicted by TRACE3, suggesting that Sc abundances are buffered by a precipitating phase (or phases) in the





Figure 7.5a-c



Zr (ppm)









Figure 7.5d-f



Figure 7.5g-j

magma chambers prior to eruption, possibly a combination of plagioclase and clinopyroxene. However, Co (Figure 7.5j) shows very poor correlation with the generated LLD from both MB2 and ML219 as starting compositions.

Multi-element diagrams of a representative collection of MVG tholeiites normalised to Normal (N-type) MORB are shown in Figure 7.6 (Sun and McDonough, 1989). The MVG basalts display similar shaped patterns, with greater enrichment in the highly incompatible elements (i.e. Rb, Ba, K, Th, Nb, Ta, La, Ce) than N-type MORB and resemble enriched (E-type) MORB. The greater data spread on the left-hand side of the plots is probably the effect of secondary alteration. Sr exhibits a small positive anomaly in some samples, e.g. ML219, and may represent plagioclase accumulation (though none of these samples shows anomalously high Al₂O₃ values or obvious plagioclase accumulation in thin section). KW9, the most evolved MVG sample shows a trough at Sr in Figure 7.6b which is consistent with plagioclase fractionation. Figure 7.6c compares the general range of the MVG samples with average values from the ODP samples of Sites 807 (both Units A and C-G), Site 803, and from the Nauru Basin tholeiites drilled on ODP Legs 61 and 89, Site 462A. The MVG tholeiites show complete overlap with the averaged OJP data and the Nauru Basin tholeiites.

Samples normalised to primitive mantle (Sun and McDonough, 1989) are plotted in Figure 7.7. No obvious positive Sr anomalies are seen in the more primitive samples, although KW9 shows a negative anomaly consistent with plagioclase removal. An interesting point is that the MVG basalts display a trough in potassium. This may be a direct reflection of secondary alteration, although secondary alteration in the MVG basalts has tended to increase K values, not deplete them, though some depletion of K may have occurred because of tropical weathering. However, the average of Nauru Basin lavas (Figure 7.7c) shows a similar negative anomaly in K. These samples have apparently remained in a submarine environment, suggesting that low grade submarine alteration may have leached K in the Nauru lavas and therefore a similar processes may have affected the MVG lavas.

7.4 Mineral phases involved in the fractionation of the MVG

The TRACE3 program of Nielsen (1988) calculates an evolving liquid composition (LLD), and the portions of the fractionating mineral phases. According to these calculations, the









fractionation of the MVG is dominated by olivine, plagioclase and clinopyroxene, which is in accordance with the petrological studies of the MVG (Chapter 5). Table 7.1 summarises the MgO (wt%) of the modelled LLD at which individual mineral assemblages begin to fractionate. Olivine is the initial fractionating assemblage in the two MVG basalts (ML219 and MB2) followed by plagioclase and finally clinopyroxene (augite). Orthopyroxene and titanomagnetite do not reach the liquidus until more evolved compositions (lower MgO values) than any of the MVG basalts we sampled. Towards the more evolved compositions of the TRACE3 generated LLDs, inaccuracy tends to be higher as it does not model apatite or alkali feldspar fractionation, although the latter of the two is not found in the MVG lavas and apatite is only found very rarely, and only as a groundmass phase.

Sample	Initial MgO (wt%) of magma	Cr spinel	Olivine	Plagioclase	Clino- pyroxene	Ortho- pyroxene	Titano- magnetite
ML219 MB2 Cur7	10.02 7.71 19.99	18.53	10.02 7.71 19.99	8.50 7.30 8.46	7.72 7.30 8.13	4.35 5.02	3.79 3.62 3.10

 Table 7.1 The MgO (wt%) at which mineral phases begin to fractionate on the TRACE3

 generated LLDs.

Chrome spinel is not precipitated out of the MVG basalts in the TRACE3 program (Table 7.1), presumably because even the most primitive of the MVG samples are too evolved to fractionate chrome spinel. Again this agrees with the observed petrological features of the MVG as no chrome spinels were found in the MVG basalts; probably as they had already been removed via fractionation processes prior to eruption. Table 7.1 also shows the results for CUR7, a picrite from Curaçao (Kerr et al, 1996), for comparison. Olivine and Cr spinel are initial fractionating mineral assemblages, followed by plagioclase and then clinopyroxene, though at slightly higher MgO values than seen in the MVG. An example of the minerals removed and their abundances in the fractionation assemblage, as predicted by TRACE3, is shown in Table 7.2, with ML219 used as the parental composition.

Unfortunately TRACE3 only models at pressures of 1 atm, and while the LLDs fit well for the majority of the MVG basalts, they are not perfect. Fractionation of the MVG at different pressures (as suggested by the CIPW norms) is likely to be the cause of the differences, as it
will cause changes the in composition and proportions of the mineral assemblages removed.

The most likely mineral phase to be affected by higher pressure fractionation is augite, as the composition of augite changes with variations in pressure.

TRACE3 generated LLD for ML219											
LLD MgO (wt%)	Temp (ºC)	Olivine	Ortho- pyroxene	Clino- pyroxene	Plagioclase	Titano- magnetite	% crystallisation				
9.86	1230.3	0.2	-	-	-	-	0.2				
8.5	1190.4	4.4	-	-	0.6	-	5.2				
8.26	1185.9	1.6	-	-	3.4	-	10.2				
7.91	1182.0	1.8	-	-	3.2	-	15.2				
7.72	1178.3	0.6	-	1.8	2.6	-	20.2				
7.49	1175.4	0.4	-	2.4	2.2	-	25.2				
7.24	1171.1	0.4	-	2.4	2.2	-	30.2				
6.95	1167.6	0.6	-	2	2.4	-	35.2				
6.62	1162.5	0.6	-	2	2.4	-	40.2				
6.28	1157.3	0.4	-	2.4	2.2	-	45.2				
5.89	1150.0	0.4	-	2.4	2.2	-	50.2				
5.45	1142.5	0.4	-	2.4	2.2	-	55.2				
4.94	1133.0	0.4	-	2.4	2.2	-	60.2				
4.35	1122.5	-	0.8	2	2.2	-	65.2				
3.79	1113.2	-	0.4	2.2	2.2	0.22	70.2				

 Table 7.2. Percentages of the fractionated assemblages predicted by TRACE3, assuming a starting composition of ML219.

Kerr (1993) showed that in the Mull Plateau Group, which fractionated at pressures of approximately 9 kbars according to CIPW norm values, a large amount of the discrepancies between the TRACE3 generated LLDs and the actual lavas could be accounted for by changes in augite compositions and abundances caused by higher pressure fractionation. The Mull Plateau Group showed poor correlations with the calculated LLDs, especially in CaO, Sc and V values, which was explained as being the result of high pressure fractionation of clinopyroxene. Higher pressure augites tend to have lower CaO and higher Al₂O₃ contents than 1 atm augites (Thompson, 1974). For example, in studies on the Skye Main Lava Series by Thompson (1974) a groundmass pyroxene formed at 1 atm has CaO = 21.9 wt% and Al₂O₃ = 3.17 wt% values, while a pyroxene from the Skye Main Lava Series formed at 17

kbars has CaO = 7.14 wt% and Al_2O_3 = 5.46 wt%. (The Skye Main Lava Series also plots along the 9 kbar CIPW norm cotectic, similar to the Mull Plateau Group).

Kerr (1993) showed by least-squares mass balance calculations that a Mull Plateau Group basalt with MgO \approx 10 wt% could be produced by removing \approx 5% olivine (Fo₈₃) and \approx 2% high pressure augite (Thompson, 1974) from a parental basalt with MgO \approx 12.5 wt%. However, the Mull Plateau Group, similar to the MVG, shows only minor amounts of clinopyroxene phenocrysts in petrological studies (Kerr, 1993). This lack of clinopyroxene in the Mull Plateau Group basalts, even though the composition suggests clinopyroxene was fractionated, may be because i) magmas containing high-pressure phenocryst phases ponded in sub-crustal chambers, resorbing the high pressure phases and was followed by the fractionation of lower pressure phases (Thompson et al., 1980) and/or ii) in-situ or side-wall crystallisation of augite occurred (Langmuir, 1989), which would be difficult to erupt. If augite was fractionated onto the walls of a magma chamber at depth, it would effectively have removed the higher pressure mineral phases from the Mull Plateau Group magmas before moving to shallower levels, as envisaged by Kerr (1993).

The MVG basalts may have resorbed higher pressure mineral phases in shallower magma chambers, explaining the poor fits of some of the LLDs. The extent to which magma reequilibration occurred may have varied, explaining the anomalous CIPW trend because some of the higher % *hy* basalts did not have time to fully re-equilibrate (Figure 7.2b). The paucity of augite phenocrysts in the MVG lavas could also be explained by in-situ crystallisation, possibly at pressures greater than 1 atm; recall that some of the augite phenocrysts have higher Al/Ti ratios than the (1 atm) groundmass augites (Section 5.3.3). None of the augite phenocrysts shows high Al₂O₃ or low CaO values as seen in the Skye Main Lava Series high pressure clinopyroxenes. However, it should be stressed the differences that Kerr (1993) found between the Mull Plateau Group and the generated LLDs were far greater than any differences we see in the MVG and their TRACE3 LLDs. This is possibly because the MVG did largely, though not completely, evolve at shallower depths and lower pressures because of their oceanic setting. It is known that shallow level magma chambers existed because of the presence of large plagioclase megacrysts in some of the MVG basalts.

In summary, variations in the proportions of augite to plagioclase fractionated as

predicted by TRACE3 would also effect the LLDs. A higher ratio of augite to plagioclase would most likely prevail at depth greater than 1 atm, thus an increase in the portion of augite to fractionate would raise Al₂O₃ and lower CaO contents in the LLDs, forming a better fit with the MVG data.

7.5 Rare-earth elements of the MVG

The two distinct source signatures within the 122 Ma OJP lavas were initially documented and discussed by Mahoney et al. (1993) on studies of the OJP basalts drilled by the ODP Leg 130 and DSDP Leg 30 on the main body of the plateau. Distinguishing between these two groups of basalts is often difficult (hence both Site 807, Unit A and C-G lavas are grouped together in Figures 7.4 and 7.5). The most pronounced difference is seen in their isotopic signatures (Section 7.6), although distinctions can also be observed in the REE abundances.

The MVG basalts have been subdivided into Type A and Type C-G on a discrimination basis of La/Sm \geq 1.54 and <1.54 respectively; their REE patterns are shown in Figure 7.8a and b. Type A lavas are more evolved and hence generally display higher abundances of REE, they are also more enriched in the LREE compared to Type C-G basalts with La/Sm \geq 1.54. ML543 and KW13 plot in the stippled field which represents the overlap between Type A and Type C-G MVG basalts in this study. Figure 7.8c also compares the OJP basalts with E and N-type MORB. The OJP basalts appear less depleted in terms of LREE compared to N-type MORB, but do not show the enrichment in LREE typical of E-type MORB.

REE abundances of MVG basalts are shown in Figure 7.9a-e plotted against Mg number. The samples have been divided into Type A and Type C-G basalts on their La/Sm ratios as were the REE patterns (Figure 7.8). The REEs behave incompatibly, displaying a negative correlation with Mg number. The LREE show a greater spread in the data, more clearly showing the sub-parallel trends of the two different basalts groups (outlined by the LLD for MB2 and ML219). The distinction between the two basalt types, clear in the LREE, is not visible in Yb and Lu (with the exception of MB2). The REE abundances in the MVG basalts overlap between the two groups, showing a clear trend oblique to the sub-parallel





Figure 7.8a-c



Figure 7.9a-e

LLDs in Figures 7.9a-c, implying that magma mixing may have occurred to varying extents between the two groups.

7.6 Sr, Nd and Pb isotope ratios of the MVG

Seven OJP tholeiites, two alkali basalts from the Northern Malaita Alkali Suite (NMAS) and a feldspar megacryst from a MVG dolerite were analysed for Sr, Nd and Pb isotopes. The full isotope data set can be found in Appendix F. The range of Sr, Nd and Pb radiogenic isotopes of the MVG tholeiites analysed in this study are limited (Figure 7.10a-e), £Nd_{t=122} ranges from +3.4 to +6.1 and (⁸⁷Sr/⁸⁶Sr)_{t=122} ranges from 0.703700 to 0.704280 (£Nd and (⁸⁷Sr/⁸⁶Sr)_t values quoted above have been age corrected to 122 m.y to give their initial values). Present-day Pb isotopes also show a restricted range, with ²⁰⁶Pb/²⁰⁴Pb values varying between 18.40 and 18.69, ²⁰⁷Pb/²⁰⁴Pb between 15.49 and 15.58 and ²⁰⁸Pb/²⁰⁴Pb between 38.41 and 38.61. This latest set of results adds to the expanding isotope data set for the OJP, but reveals few surprises. The MVG isotopes from this study all fall within the field outlined by previous isotope studies from ODP Sites 807 (Units A and C-G), Site 803 and DSDP Site 289 (Mahoney, 1987; Mahoney and Spencer, 1991; Mahoney et al., 1993). They also show overlap with earlier isotope results from southern regions of the MVG, as well as OJP tholeiites from the islands of Santa Isabel and Ramos in the Solomon Islands (Mahoney and Spencer, 1991; Tejada et al., 1996).

Previous studies on the OJP lavas recovered during ODP Leg 130 subdivided the Site 807 basalts into Units A and Units C-G (Mahoney et al., 1993), with Units C-G typically showing higher ENd, ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb and ²⁰⁸Pb/²⁰⁴Pb values and lower ⁸⁷Sr/⁸⁶Sr values. The basalts from Site 803, which erupted at about 90 Ma, resemble isotopically those belonging to Units C-G, as do the highly altered Cretaceous basalts from Site 289. The clustering of isotopes into two groups was also recorded in more recent studies of portions of the OJP obducted in Santa Isabel, Southern Malaita and Ramos Island (Tejada et al., 1996). The OJP basement lavas on Santa Isabel, the Sigana Basalts, also fall into two isotope clusters, A and C-G type, and age groups, 122 and 90 m.y.. The 122 m.y. Sigana Basalts isotopically resemble Unit A lavas, while a younger, 90 m.y. group of lavas, show Units C-G like isotope signatures. Previously analysed samples from the MVG in Southern Malaita, dated at 122 m.y. (Mahoney and Spencer, 1991; Tejada et al., 1996), also reveal isotopic



Figure 7.10a-c

Figure 7.10a-c. Nd and Sr isotopes have been age corrected, Pb isotope values are present day values. Shaded fields show previous OJP isotope studies (Mahoney and Spencer, 1991; Mahoney et al., 1993; Tejada et al., 1996), dashed lines outline isotope groups from this study. 1 σ error bars for Nd are as shown, 1 σ error for Sr and ²⁰⁶Pb/²⁰⁴Pb is within the size of the symbols. Error bars for previous studies can be found in the individual references, and are generally similar to those reported in this study.



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Figure 7.10d&e

Figure 7.10d&e. Present day Pb isotope values for the OJP are shown above. Shaded fields show data from previous OJP isotope studies (Mahoney and Spencer, 1991; Mahoney et al., 1993; Tejada et al., 1996), dashed lines outline isotope groups from this study. 1 σ error bars for this study are as shown (²⁰⁶Pb/²⁰⁴Pb errors are within the size of symbols). Error bars for previous studies can be found in the individual references, but are generally similar to those in this study.

signatures similar to Unit A, as do one 122 m.y. basalt from the OJP recovered from Ramos Island.

Our studies reveal the presence of both Unit A and C-G type lavas in the MVG tholeijtes, although the Pb isotopes do not fall into such distinctive groups as shown by previous studies. The division of MVG basalts into Type A and Type C-G samples in this isotope study was established via Sr and Nd isotopes (as seen in Figure 7.10a) and trace element ratios, e.g. La/Sm ratios (Section 7.5). Pb isotopes from previous studies show a divide in ²⁰⁶Pb/²⁰⁴Pb values between the Unit A (low ²⁰⁶Pb/²⁰⁴Pb) and C-G (high ²⁰⁶Pb/²⁰⁴Pb) clusters (Figure 7.10d and e). In this study several of the Type A samples display higher ²⁰⁶Pb/²⁰⁴Pb values, more typical of Type C-G values (as shown in Figure 6.11). The present study (Section 6.4.7) has shown that the leaching of the powders may move data from one grouping to another, effectively increasing ²⁰⁶Pb/²⁰⁴Pb values. Furthermore, KW13, a Type C-G basalt (based on La/Sm, ENd and (87Sr/86Sr)t values) showed Pb isotopes which were characteristic of Type A lavas until it was leached, causing it to shift to Pb isotopic values more characteristic of Type C-G lavas. The previous studies on the Pb isotopes of the OJP basalts were carried out on unleached samples, whereas in this study they were leached (Section 6.4.7). Mahoney and Spencer (1991), in their leaching experiments on the Pb isotopes, recorded no change in the isotope values. However, in this study there is an obvious shift to more radiogenic values after leaching leading to the conclusion that the Pb isotopes have in fact been affected by secondary alteration processes.

In addition to the MVG tholeiites, a plagioclase megacryst found within a MVG lava flow was also analysed (Figure 7.11a-e). The megacryst revealed isotopic values, age corrected to 122 m.y., of ENd = +7.4 and $({}^{87}Sr/{}^{86}Sr)_t = 0.703969$, with present day Pb isotope values of ${}^{206}Pb/{}^{204}Pb = 18.37$, ${}^{207}Pb/{}^{204}Pb = 15.56$ and ${}^{208}Pb/{}^{204}Pb = 38.24$. This feldspar megacryst plots away from any of the previously described isotope fields for the OJP basalts, the low (${}^{87}Sr/{}^{86}Sr)_t$ value overlaps with Type C-G basalts. However, the ENd value of +7.4 is significantly higher than the Type C-G basalts, and is the highest reported value for the OJP. Also the Pb isotopes show a mixture between Type A and Type C-G, ${}^{206}Pb/{}^{204}Pb$ and ${}^{208}Pb/{}^{204}Pb$ values, hinting at a Type A signature, but its ${}^{207}Pb/{}^{204}Pb$ values is high, more typical of Type C-G. It is difficult to say which basalt group the plagioclase megacryst most clearly resembles, Type A or C-G, or whether it represents a new isotopic group. The Pb isotopes, typical of neither Type A nor C-G, may represent a truer source signature than suggested by Type A and C-G tholeiites, whose Pb isotopes have been altered by secondary processes.

Two basalts from the Northern Malaita Alkali Suite (NMAS) analysed in this study have $\text{ENd}_{t=44}$ of +4.7 and +4.9, ($^{87}\text{Sr}/^{86}\text{Sr}$)_{t=44} of 0.703786 and 0.703697 and present-day $^{206}\text{Pb}/^{204}\text{Pb}$ of 18.82 and 18.82, $^{207}\text{Pb}/^{204}\text{Pb}$ of 15.56 and 15.55 and $^{208}\text{Pb}/^{204}\text{Pb}$ of 38.85 and 38.85. The isotope results from the two NMAS samples overlap well with previous results of the NMAS, which fall within a very tight data field (Tejada et al., 1996).

The new isotope data from this study for the OJP and the NMAS are plotted along with a selection of oceanic plateaux, OIB and MORB data in Figures 11a-c. The OJP tholeiites in general display an isotopic range typical of an oceanic island or hotspot-related volcance. Also, regardless of all the discussion of the different isotope signatures within the OJP tholeiites, the isotope range for the OJP discovered to date is very restricted in comparison to other oceanic plateaus such as Kerguelen; even the limited sample base from the Nauru Basin shows a wider isotopic range. Considering the geographic range over which the OJP samples were collected, with the ODP and DSDP drill sites being situated ~1600 km away from the portions of the OJP seen exposed in the Solomon Islands, the isotopes, indeed the chemical compositions of the OJP lavas, are remarkably restricted. This is discussed further in Chapter 8.

In addition to the tholeiites, several alkali suites have been sampled from Santa Isabel and Malaita. On Malaita the Southern Younger Volcanics (SYV) and the Northern Malaita Alkali Suite (NMAS) show distinct isotope signatures (Tejada et al., 1996), hinting at the involvement of different source regions. They erupted contemporaneously at about 44 m.y. which lead Petterson (1995b) to group them both into the Maramasike Formation. On Santa Isabel, a group of intrusives, the Sigana Alkali Suite, erupted at 90 m.y., the same time as some of the tholeiites exposed on the island. This is an indication that the 90 m.y. activity was geochemically bimodal. It is likely that both the Sigana Alkali Suite and the tholeiites were derived from the same plume source region by varying degrees of partial melting, whereas the two 44 m.y. old alkali basalt suites exposed in Malaita are unrelated to the OJP plume (Tejada









et al., 1996). The Sigana Alkali Suite display significantly higher ²⁰⁶Pb/²⁰⁴Pb ratios than the bulk of the basalts from the OJP, possibly the result of melting a minor, but highly fusible, component within the OJP source region (Chapter 8). The 44 m.y. old Southern Younger Volcanics from Malaita may be linked to the Samoa or Rarotonga hotspots which the OJP passed over at about 44 m.y. (Yan and Kroenke, 1993; Tejada et al., 1996). The North Malaita Alkali Suite shows isotopes which are not too dissimilar to the OJP tholeiites, with the exception of higher ²⁰⁶Pb/²⁰⁴Pb and ²⁰⁸Pb/²⁰⁴Pb values and it is possible that these North Malaita Alkali Suite basalts represent small degree melts from the OJP lithosphere (Tejada et al., 1996). For a more detailed description of the alkali suites and their source regions the reader is referred to Tejada et al. (1996).

7.7 New Hf isotopes for the MVG

This study reports a new set of Hf isotopes for the MVG tholeiites. Present day ¹⁷⁶Hf/¹⁷⁷Hf isotope values for the MVG vary between 0.283042 and 0.283240 and full results can be found in Appendix F. Figure 7.12 plots these new Hf values against Nd values, showing the MVG plot towards the high end of the OIB field along with Hawaii, contrasting with basalts from the Kerguelen Plateau which plot towards the isotopically low end of the OIB array. The Kerguelen basalts are thought to have been contaminated by a continental lithosphere material (Storey et al., 1988, 1989) with low time-integrated Lu/Hf and Sm/Nd ratios. No such continental contamination appears to have influenced the Hf and Nd isotopes of the MVG (or the Pb isotopes, Section 7.6), in contradiction to suggestions that the lower portions of the OJP may be continental in nature (e.g. Nur and Ben-Avraham, 1982).

The MORB field (Figure 7.12) shows a wide range of Hf isotopes for a relatively restricted range of Nd isotopes. This has been suggested to be the result of a decoupling between Lu/Hf and Sm/Nd in an ancient depletion event of the MORB source region (Salters and Hart, 1991). However, reinterpretation of existing Hf isotopes on MORB samples has shown the high ¹⁷⁶Hf/¹⁷⁷Hf end of the MORB field is only defined by three samples (as shown in Figure 7.12), all of which are from ocean ridge segments in close proximity to known plume localities (Nowell and Kempton, 1996). These three samples hint at the influence of a depleted component which is isotopically distinct from depleted MORB source (DMM),

removing the need for any complex decoupling of Lu/Hf and Sm/Nd in the DMM source. The new data from Iceland and ODP Leg 152 show distinct trends towards high ¹⁷⁶Hf/¹⁷⁷Hf components (G.F. Nowell, P.D. Kempton, J.G. Fitton and A.D. Saunders, unpublished data). The new Hf data from Malaita cannot be split into Type A and Type C-G on the basis of their Hf isotopes alone, unlike their Nd isotopes. Two of the samples, which admittedly have large error bars, plot at high ¹⁷⁶Hf/¹⁷⁷Hf values, trending towards a higher ¹⁷⁶Hf/¹⁷⁷Hf component such as seen in the Leg 152 data. Whether this represents the involvement of a depleted mantle component which is intrinsic to the plume itself, and isotopically distinct from low ¹⁷⁶Hf/¹⁷⁷Hf DMM signature is unclear.

7.8 New PGEs for the MVG

Twelve MVG samples were analysed for PGEs using ICP-MS techniques at the Australian National University in Canberra; full results are as shown in Appendix Eii. The abundances of the PGEs Pd, Pt, Rh, Ru and Ir, as well as Re, Au, Ni and Cu, in the MVG are plotted against Mg number in Figure 7.13a-j. The MVG basalts plotted in Figure 7.13 are divided into Type A and Type C-G on the basis of La/Sm>1.54 and <1.54 respectively as previously described (Section 7.5). Type C-G basalts generally have higher abundances of PGEs, Au, Ni and Cu compared to Type A basalts, though there is some overlap between Re and Ru in the two groups.

A summary of PGE abundances in OIB and MORB are included in Table 7.3. There is a paucity of PGE data for oceanic basalts, and a large number of the data sets contain few elements beyond Pd and Ir. Table 7.3 also includes PGE averages of hotspot-related basalts from the Canada Long Range tholeiitic dykes (Greenough and Owen, 1992) and basalts drilled by ODP Leg 115 from the Reunion hotspot trace (Greenough and Fryer, 1990). PGE averages of komatiites from the Alexo flow in the Archaean Abitibi greenstone belt, and Cretaceous komatiites from Gorgona Island, are also included in Table 7.3 (Brügmann et al., 1987). The average PGE values of MVG Type C-G lavas shown in Table 7.3 are similar to other hotspot lavas, and have average PGE abundances only slightly lower than those of Leg 115 basalts and the Canada Long Range tholeiites. MVG Type A basalts have far lower abundances of PGEs relative to Type C-G lavas, often by as much as a factor of 10. Indeed,



Figure 7.13a-j

the Type A lavas are also more depleted than MORB values quoted in Table 7.3. Note, however, that the MORB values are limited to Au, Pd and Ir, and are average values from a data set which comprises only 18 samples (Fryer and Greenough, 1992).

		Au (ppb)	P d (ppb)	Pt (ppb)	Rh (ppb)	Ru (ppb)	lr (ppb)	Pd/lr
Typical MORB	[1]	0.69	≤0.7	-	-	-	≤0.035	≈100
OIB (tholeiitic)	[1]	2.3	2.2	-	-	-	0.25	8.8
OIB (alk basalt)	[1]	0.5	1.5	-	-	-	0.3	4.5
Leg 115 average	[2]	3.2	8.1	7.3	0.31	0.22	0.11	74.0
Gorgona average	[3]	4.33	12.38	-	-	3.43	2.06	6.0
Alexo average	[3]	2.06	8.67	-	-	2.81	2.19	4.0
Canada thois average	[4]	4.01	10.37	7.11	0.46	0.25	0.12	87.5
MVG Type A average		0.21	0.22	0.37	0.02	0.04	0.02	12.8
MVG Type C-G average		1.83	3.34	3.84	0.22	0.11	0.13	26.1
Primitive mantle	[5]	1.2	4.4	8.3	1.6	5.6	4.4	1.0

[References: 1, Fryer and Greenough, 1992 (and references therein); 2, Greenough and Fryer, 1990; 3, Brügmann et al., 1987; 4, Greenough and Owen, 1992; 5, Barnes et al., 1988 (and references therein)].

Table 7.3. Average PGE and Au values in MORB, OIB, komatiites and MVG basalts.

PGE abundances in the MVG basalts either fall or remain approximately constant with decreasing Mg number (Figure 7.13b-h), consistent with the PGEs acting as compatible elements. Pd and Ir are commonly analysed in samples which contain low concentrations of PGEs, such as MORB, as they are among the more easily measured of the PGEs (e.g. Barnes et al., 1985). Pd and Ir conveniently fall at opposite ends of the PGE patterns (Figure 7.14), therefore Pd/Ir ratios are used to define the slope of these patterns, in much the same way as La/Sm ratios are used in REE patterns. Pd/Ir ratios are higher in MVG Type C-G basalts compared to Type A basalts (Figure 7.13j), increasing slightly with increasing Mg number, though both groups display Pd/Ir ratios far less than the Pd/Ir \approx 100 values of MORB, the Pd/Ir = 74 of average Leg 115 basalts and the Pd/Ir = 94 of the Canada tholeiites (Table 7.3).

The PGE patterns for the MVG basalts are plotted in Figure 7.14a-c (primitive mantle values used are as shown in Table 7.3; Barnes et al., 1988). Type C-G basalts (Figure 14b) appear to display tighter patterns than Type A samples, though the Type A samples have a more restricted range in most of the PGEs and Pd/Ir ratios compared with Type C-G basalts (Figure 7.13j). This spread is a reflection of the lower abundances of PGEs in Type A samples





Figure 7.14a-c

(and the decreasing accuracy of the data at such low abundances). Type C-G basalts show depletion in IPGEs (Ir group PGEs consisting of Ir, Ru and Os) relative to PPGEs (Pt group PGEs consisting of Rh, Pt, Pd and Au) when normalised to primitive mantle values. Type A lavas show a similar depletion in Ir and Ru, but they also are depleted in Rh relative to Pt, Pd and Au. Type A samples also show elevated Cu and Ni values compared to their PGE abundances, a feature not seen in Type C-G lavas.

It is worth noting that previously published PGE patterns for the OJP lavas (Jain et al., 1996) are very different from the PGE patterns reported in this study, and indeed any hotspot related basalts yet studied. Jain et al. (1996) reported PGE patterns for the OJP showing high peaks in Ir, Pt and Au, which they attributed to core/mantle boundary effects in the OJP's plume source region. The core/mantle boundary indeed may be the origin of the plume responsible for the eruption of the OJP (Section 8.6.3) though why it would give such apparent decoupling between IPGEs and PPGEs, which is not seen in the PGE patterns of any other plume-related basalt is unclear. The new PGE data reported here for the MVG basalts do not show such peaks at Ir, Pt and Au.

The Ir and Ru depletion in the MVG basalts is also seen in the Canada tholeiites, Leg 115 basalts and, though less pronounced, in the Alexo and Gorgona komatiites (Figure 7.14c). These patterns depleted in IPGEs are common in MORB and OIB lavas (e.g. Barnes et al., 1985), though the reason for the fractionation of PGEs is debated (e.g. Barnes et al., 1985); Brügmann et al., 1987; Campbell et al., 1983). PGEs are strongly fractionated into sulphides, and can hence be a useful indicator of sulphur saturation in melts, which is not indicated by lithophile elements (e.g. Barnes et al., 1985). However, there is a lack of agreement to the values of partition coefficients of PGEs between sulphide liquids and silicate melts, though they are generally thought to be high, varying between 10^3 and 10^5 (e.g. Ross and Keays, 1979; Campbell and Barnes, 1984; Stone et al, 1990). If the fractionation of PGEs, as seen in Figure 7.14, is because of the mixing of sulphides into the magma during melting, then the partition coefficient of Ir between sulphide liquids and silicate melts (sul/sil) must be greater than for Pd, i.e. $D^{Ir}_{sul/sil} > D^{Rh}_{sul/sil} > D^{Pt}_{sul/sil} > D^{Pd}_{sul/sil}$ (Barnes et al., 1985). Hence in komatiites, which display the highest extents of partial melting, all sulphides theoretically will have melted and entered the magma, causing a lowering of Pd/Ir ratios.

However, removal of sulphides by fractionation, whether by incorporation as inclusions or solid solution into silicate minerals will cause Pd/Ir ratios to increase. Dunites have been shown to be enriched in IPGEs (Fominykh and Kvostova, 1970; Oshin and Crocket, 1982), suggesting that the removal of olivine and chromite prior to eruption may have depleted the MVG in IPGEs. However, various analyses of PGE in olivines, spinels and silicate minerals (e.g. Ross and Keays, 1979; Naldrett et al., 1979; Crockett, 1979; Mitchell and Keays, 1981) have given conflicting results, leading to opposing interpretations as to whether these elements are in fact concentrated in olivine and chromite crystal structure because of solid-solution substitution, or the result of formation of platinum group minerals (PGMs) or alloys, often as inclusions within olivine and chromite. Analysis of inclusions within chromites from ophiolites and Alpine complexes (e.g. Prichard et al., 1981; Stockman and Hlava, 1984), indicate they are enriched in Os, Ir and Ru metal alloys. It seems these alloys crystallised from the magma, then chromite nucleated on them and settled out, depleting the melt with respect to Ir, Os and Ru (e.g. Barnes et al., 1985, and references therein).

It is impossible to tell whether the fractionation of the MVG's PGE patterns was because of the entrapment of inclusions of Ir and Ru bearing sulphides within chromite and olivine, or solely due to the removal of olivine and chromite. However, we do know that large amounts of olivine and chromite have been removed from the MVG magmas prior to eruption (Section 7.4). The partition coefficients of Ni and Cu into sulphide liquids, though high, are lower than PGEs (both ≈250 (Rajamani and Naldrett, 1978)), hence, removal of only small percentages of sulphides would have little effect on the Cu and Ni but a drastic impact on PGE abundances. Therefore, the removal of slightly larger amounts of sulphides in Type A lavas relative to Type C-G lavas could explain the Cu and Ni anomalies. Similarly, the higher abundances of PGEs in Type C-G lavas may be a direct reflection of higher percentages of melting in the source dissolving more sulphides, and producing lavas with higher PGE abundances in the basalts. The PGEs and their implication for source compositions and percentages of melting in the OJP are discussed further Chapter 8.

Finally, the MVG sample KW13 is circled in Figure 7.14a-j. This basalt contains some of the highest PGE abundances in the MVG basalts analysed. During probing, a couple of sulphide blebs were found encased by an augite crystal and, although no detailed study of the abundance of sulphides in the MVG basalts analysed for PGEs was carried out, the presence of sulphides in this sample is likely to have influenced its PGE levels. The presence of a sulphide within a clinopyroxene, a common groundmass mineral phase within the MVG, rather than within olivine, which is usually altered or chromite (which is absent), implies that fractionation of olivine and chromite alone may not have been solely responsible for the removal of sulphides from evolving MVG magmas.

7.9 Concluding points

• The MVG are low K-tholeiites, with over 95% being quartz normative. They display a restricted range of major and trace elements which overlap with basalts from Site 807 (Unit A and C-G) and from the Nauru Basin. Compared with N-type MORB, they are enriched in highly incompatible elements and are similar to E-type MORB.

• The limited range and evolved composition of the MVG basalts is consistent with the lavas having been well mixed prior to eruption, probably the result of re-equilibration in shallow level magma chambers at pressures of ≥1 atm. However, the CIPW norms of the MVG basalts imply that this re-equilibration may not have always been complete.

The fractionating assemblage of the MVG basalts, as calculated by TRACE3 (Nielsen,
 1988) is: OLIVINE ⇒ PLAGIOCLASE ⇒ CLINOPYROXENE ⇒ CLINOPYROXENE
 + OLIVINE + PLAGIOCLASE + PLAGIOCLASE
 + OLIVINE (± Ti MAGNETITE)
 (± ORTHOPYROXENE)

which is in agreement with petrological studies of the MVG basalts.

• Although the MVG basalts are relatively homogeneous in their major and trace element compositions, different groups (Type A and Type C-G) can be distinguished in their REEs, PGEs and isotopes. Type A basalts typically have La/Sm \geq 1.54, (87 Sr/ 86 Sr)_t > 0.704, ENd < 5 and Pd/Ir \geq 10 and < 20. Type C-G basalts have La/Sm < 1.54, (87 Sr/ 86 Sr)_t < 0.704, ENd > 5 and Pd/Ir >10 and < 40.

• Distinctions between the Pb isotopes of Type A and Type C-G basalts in this study is less clear than those made in previous studies on the OJP basalts. This may be a direct

result of alteration of the Pb isotopes in the unleached samples analysed in previous studies.

• New Hf isotope data on the MVG basalts do not distinguish between Type A and Type C-G samples. Two of the samples plot towards higher ¹⁷⁶Hf/¹⁷⁷Hf values, suggesting the possible involvement in the MVG basalts of a depleted component with high ¹⁷⁶Hf/¹⁷⁷Hf values, and is distinct from DMM.

• The Type C-G MVG basalts typically display enrichment in PGEs, compared to Type A basalts, by as much as a factor of ten. Both groups are depleted in IPGEs, either because of the fractionation of olivine and chrome spinel enriched in these elements or the sulphide inclusions they contain. Similarly, higher extents of melting may have preferentially enriched the Type C-G basalts because of the melting of more sulphides in the source prior to eruption.

• The restricted range of radiogenic isotope ratios, REE, major and trace element abundances displayed by the MVG tholeiites in this study are typical of OIB basalts. They show a complete overlap with the range of basalts recovered from the main body of the OJP, arguing against theories which claim the Pacific Province is unrelated to the OJP, but an isolated fragment of the Indo-Australia Plate (Musgrave, 1990).

• The tight range of Pb-isotope ratios, positive ENd and high ¹⁷⁶Hf/¹⁷⁷Hf argue against speculation that the OJP is a submarine continent, i.e. it has a hidden basement of continental crust through which the OJP basalts were erupted (e.g. Nur and Ben-Avraham, 1982).

• Overall, the MVG basalts in this study appear to show a greater spread in most elements than recorded in the ODP drilled samples. This is not surprising considering the ODP samples were recovered from a maximum depth of 149m into the lava pile, compared with the several km depth we sampled into the lava pile on Malaita. However, perhaps in light of the variation in composition displayed by the ODP drilled samples, especially Site 807, the increase in the OJP data range this study reveals is only very minor compared to what may have been expected.

Chapter 8

Characteristics of the mantle source of the Malaita Volcanic Group

8.1 Introduction

The MVG lavas can be divided into Type A and Type C-G on the basis of their Nd and Sr-isotope and La/Sm ratios. Whether these differences are the result of different extents of partial melting, the depth at which melt segregates or compositional differences in the source region is discussed in this chapter. The depth and extent of partial melting within the OJP plume are also vital considerations when attempting to constrain the eruption environment of the OJP, i.e. was it erupted in an intra-plate or ridge setting?

An impinging plume provides thermal buoyancy and high melt productivity resulting in lithospheric uplift (Crough, 1983; Griffiths et al., 1988); lithospheric uplift is followed by subsidence as the thermal influence of the plume wanes (Coffin and Eldholm, 1994). The MVG, and material recovered by the ODP drill sites on the main body of the plateau, imply a submarine eruption environment for the OJP, possibly as deep as below the Aptian CCD. Why the OJP erupted in a submarine environment at all is unclear, especially as eruption on Iceland and Hawaii, two possible modern day analogues for the OJP, is subaerial. The postemplacement history of the MVG appears to involve a period of **uplift**, as recorded in the plateau's sedimentary sequence. Why this was so remains as much an enigma as the deep submarine eruption environment.

Finally, the composition and mantle boundary-layer origins of the OJP plume is considered. I suggest that the OJP plume was internally heterogeneous, containing both an enriched, highly fusible, component which was preferentially melted in the deeper, mantle melting regimes, i.e. the Sigana Alkali Suite. A second, more refractory component, which had a depleted isotopic signature, dominates the chemistry of the MVG tholeiites, due to the nature of their extensive melting at shallow levels in the mantle. The boundary-layer origin of the OJP plume is more elusive, although the large mantle source volumes required and the composition seem to imply at least some input from the lower mantle.

8.2 Effects of partial melting

8.2.1 Depth of partial melting

In order to study the effects of partial melting the ratios of highly to moderately incompatible elements are used (e.g. Langmuir et al., 1997). In Figure 8.1a the MVG's Nd-isotope ratios, showing the compositional difference between the two groups, are plotted against Zr/Y. Type A and Type C-G MVG basalts both show considerable overlap, although not all of the Type C-G basalts with lower Zr/Y values were analysed for radiogenic isotope ratios and some samples have lower Zr/Y values than shown in Figure 8.1a. Figure 8.1b, c and d show a larger data set and hence a better representation of the Type A and Type C-G MVG basalts.

Zr and Y both behave highly incompatibly in olivine and plagioclase, while Y (and the heavy REEs) is slightly more compatible in clinopyroxene, with D_{cpx} varying between 0.467 (Hart and Dunn, 1993) down to 0.2 (McKenzie and O'Nions, 1995, and references therein). Therefore, large amounts of clinopyroxene fractionation will have an influence on Zr/Y values. Also Y (and the heavy REEs) is compatible in garnet ($D_{gt} \approx 2.1$; McKenzie and O'Nions, 1995, and references therein) and thus Zr/Y values are higher in melts segregated at greater depths as more garnet is present in the source region melted. Sc, like Y, is also compatible in garnet lherzolite and moderately compatible in clinopyroxene, thus, melt segregated at greater depths in the garnet lherzolite field and magmas fractionating large amounts of clinopyroxene will show higher Zr/Sc values. The Type A lavas also show higher (Sm/Yb)n values than Type C-G basalts (Figure 8.1d) probably because of garnet having a greater influence in the source region melted in the production of the Type A basalts. However, the HREE are moderately incompatible in clinopyroxene, and hence large amounts of clinopyroxene fractionation may be responsible for the differences in Sm/Yb between the two groups.

In general the Type A basalts trend towards higher Zr/Y, Zr/Sc and Sm/Yb values (Figure 8.1) which is consistent with either (i) melt segregation occurring at greater average depths, i.e. with garnet having a greater influence in the Type A basalts **or** (ii) Type A lavas



Figure 8.1a-d

precipitated larger amounts of clinopyroxene than Type C-G lavas. In addition to the variation in trace element ratios, such as Zr/Y, the Unit A-like basalts generally have higher Fe_{8.0} and Na_{8.0} values compared to the Type C-G lavas. Higher Fe_{8.0} values implies melting for the Type A basalts occurred at higher average pressures than the Type C-G basalts, while higher Na_{8.0} indicates that they are the result of smaller degrees of melting (Klein and Langmuir, 1987). In addition, Type C-G basalts show lower (La/Sm)n values implying that a greater extent of melting was segregated in the spinel lherzolite field, i.e. at shallower depths, compared to Type A basalts. Type A basalts also appear to be the result of slightly smaller degrees of melting, having slightly higher concentrations of highly incompatible trace elements relative to moderately incompatible trace elements, i.e. higher Nb/Zr and Ba/Zr values (Table 8.1).

It is difficult to determine whether the difference in trace element ratios such as Zr/Y and Sm/Yb are due to differences in the depth of final melt segregation, or differing degrees of clinopyroxene fractionation. But, Type A and Type C-G lavas do have distinct isotope signatures, indicating that the difference between the two groups is unlikely to be the result of fractionation processes in the magma chamber. Therefore, I suggest the Type A lavas are the result of slightly smaller extents of melting, possibly segregating at greater depths than the Type C-G lavas. However, the difference between the two groups is minor, and both groups started melting well within the garnet lherzolite field, but most extensive melting occurred within the spinel stability field.

	εNd	(La/Sm) _n	(Sm/Yb) _n	Zr/Y	Zr/Sc	Na _{8.0}	Fe _{8.0}	Ba/Zr	Nb/Zr	Pd/lr
Type A average	3.8	1.01	1.32	2.9	1.75	1.76	10.9	0.3	0.06	12.8
Type C-G average	5.9	0.86	1.26	2.7	1.35	1.66	10.6	0.2	0.05	26.1

Table 8.1. Average compositions of Type A and Type C-G MVG samples.

8.2.2 Degree of partial melting

McKenzie and O'Nions (1991, 1995) have shown by REE inversion modelling that melt distribution as a function of depth, the total melt fraction and the total thickness of melt (or thickness of basaltic crust) can be estimated. Fractional melting inversion was undertaken by Mahoney et al. (1993) on selected OJP samples, assuming primitive and depleted (MORB) mantle sources, in an attempt to establish more quantitatively the extent of partial melting.

Basalts with MgO > 6 wt% recovered from Leg 130 (Sites 807 and 803) and Site 289 basalts were averaged together. The Unit A and C-G basalt groups were considered suitable for grouping together by Mahoney et al. (1993) because even though they imply different mantle sources, they are still geochemically very similar. The inversion was run using rare-earth abundances in both primitive mantle and depleted MORB-type mantle sources.

In summary, both primitive mantle and MORB mantle sources yielded acceptable REE patterns, both fitting within the upper and lower limits. However, the fits were not perfect with the primitive mantle source inversion and showed slight LREE enrichment and lower middle REE values (Figure 8.2c), while the MORB source inversion showed slightly depleted LREE values but enriched Sm and Nd values (Figure 8.2a). This led Mahoney et al. (1993) to suggest a mixed source between primitive and depleted MORB mantle being most appropriate for the OJP.

The calculated maximum melt fractions were high, estimated to be between 28 and 30% for both sources (Figure 8.2b and d), although the primitive mantle source indicated that melting started deeper (Mahoney et al., 1993). The fractionation correction for the OJP samples was approximately 40%, due to considerable extents of olivine crystallisation (McKenzie and O'Nions, 1991). After correction for fractionation, the total thickness of melt produced by the primitive mantle source was 24.3 km, and 17.2 km for the depleted MORB mantle source.

The fractional melting inversion techniques of McKenzie and O'Nions (1991), with modifications by White et al. (1992), was also carried out on OJP basalts obducted in the Solomon Islands of Malaita (collected from the south of Malaita), Ramos Island and Santa Isabel (Tejada et al., 1996). While no results of the inversion modelling are depicted by Tejada et al. (1996), they report that maximum partial melting was modelled to be approximately 26%, and melt thickness to be 19-22 km using sources ranging from depleted MORB to primitive mantle sources. Therefore, inversion modelling shows the extents of melting and the depth of the melt thickness generated between the OJP basalts recovered from the main body of the plateau and those exposed in several of the Solomon Islands are in good agreement.



a Element concentration ratio with respect to a normal-MORB mantle source (McKenzie and O'Nions, 1991) for average of Leg 130 and Site 289 basalts with MgO>6wt% (average of neutron activation data only). Error bars indicate standard deviation of data for each element. The heavy solid line shows the REE pattern determined by fractionation-melting inversion: the fine continuous and dashed lines represent patterns calculated from the estimated upper and lower bounds (see McKenzie and O'Nions, 1991 for details). Observed and calculated major element values (normalised to 100 wt% without K_2O and P_2O_5) are given below: fractionation correction is calculated from the difference between observed and calculated from the difference between observed and calculated from the difference between observed and calculated MgO and FeO.

	SiO ₂	TiO ₂	Al_2O_3	FeO	MgO	CaO	Na ₂ O	K ₂ O
Obs.	50.37	1.33	14.61	11.63	7.40	12.36	2.30	0.16
±	0.59	0.22	0.44	0.93	0.62	0.32	0.17	0.10
Calc.	50.24	1.24	15.10	8.67	11.33	11.14	2.28	0.26
±	0.81	0.24	0.93	0.97	1.00	1.07	0.29	0.23
<u>(Obs Calc.)</u> S.D.	0.13	0.27	-0.48	2.20	-3.34	1.10	0.07	

b. Melt fraction (by weight) with depth calculated before (heavy continuous line) and after (heavy dashed line) correction for crystal fractionation. Fine dashed and continuous lines again represent the estimated lower and upper limits, respectively. Total melt thickness calculated from the inversion is 17.2 km after fractionation correction. c, d. Same as Figures 8.2a and b except for a primitive mantle source. Here the estimated total melt thickness after the fractionation correction is 24.3 km. Calculated major element values and difference between observed and calculated values are given below.

	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O
Calc. ±	49.82 0.81	1.08 0.24	14.81 0.93	9.00 0.97	12.10 1.00	11.05 1.07	2.15 0.29	0.26 0.23
(Obs Calc.) S.D.	0.55	0.75	-0.19	1.96	-3.99	1.17	0.46	

Figure 8.2. REE inversion modelling of the OJP basalts drilled by Leg 130 (Figure 12 from Mahoney et al., 1993).

The melt thicknesses (or basaltic crust thickness) obtained from the inversion modelling for the OJP give a maximum value of 24.3 km. This is far lower than the 32-42 km crustal thickness of the OJP measured by seismic refraction and gravity studies (Furumoto et al., 1976; Hussong et al., 1979; Gladczenko et al., in press). Admittedly the 42 km thickness for the OJP is a maximum value (Hussong et al., 1979) and recent reinterpretation of refraction profiles combined with gravity modelling yield shallower Moho depths, with a maximum of 32 km (Gladczenko et al., in press). Inversion modelling carried out on basalts from Iceland indicates they resulted from partial melting of \approx 30% and melt thickness was calculated to be from 13.6 km, increasing to 22.3 km after 39% fractionation correction was applied (McKenzie and O'Nions, 1991). These high extents of partial melting and crustal thicknesses are similar to those calculated from similar inversion modelling of the OJP. The similarities between the REE inversion modelling on the OJP lavas and Iceland support suggestions that the OJP is the result of extensive mantle melting predominately at shallow levels (Section 8.3).

The thickness of the OJP crust suggested by inversion modelling is however, too low compared with even the more conservative estimates of the seismic thickness (Gladczenko et al., in press). Similar inversion modelling of Iceland basalts which gave a crustal thickness of \approx 22.3 km (McKenzie and O'Nions, 1991), is not too dissimilar to seismically measured crustal thicknesses of 19 km beneath the Northern Volcanic Zone (Staples et al., 1997). However, the seismically-measured crustal thickness of 19 km is a minimum beneath the neovolcanic zone, thickening to a maximum of 35 km under the older Tertiary lavas of north-east Iceland, due to thickening with accretion and age, and a possible temporal decrease in melt production in the neovolcanic zone (Staples et al., 1997). Large amounts of ultramafic material at shallow depths and higher temperatures may have caused the neovolcanic zone to be seismically thin, i.e. the seismic Moho may be shallower than the petrological Moho.

Another reason for the lack of correlation between inversion melt thicknesses and seismically measured plateau thicknesses may be due to active mantle upwelling. REE modelling assumes passive mantle upwelling but, as is probable when thermally buoyant mantle plumes are located at spreading ridges, the mantle upwelling may be active. If the mantle upwelling is active, more mantle may be driven through the system or melting column, causing larger amounts of melt generation, but REE inversion would give a melt thickness

which was falsely thin. It is also noteworthy that the tholeiites that were sampled by the ODP and the obducted fragments of the plateau in the Solomon Islands are from only the upper few kilometres of the plateau and, hence, they may not be truly representative of the lower, as yet unsampled portions (e.g. Mahoney et al., 1993).

8.3 Eruption Environment of the MVG - intra-plate or ridge-centred?

Regardless of the small differences in Type A and Type C-G lavas, **all** of the MVG basalts show relatively flat to slightly LREE depleted REE patterns and low Zr/Y and Zr/Sc values. This is consistent with large amounts of melt being generated at shallow levels, i.e. mostly within the spinel lherzolite field. However, while melting was segregated largely within the spinel field, melting initiated deeper, well within the garnet lherzolite field, in both groups of MVG basalts (Mahoney et al., 1993).

All the above strongly supports a long melting column for the MVG basalts, starting within the garnet stability field (>80 km) and continuing up to shallow levels well within the spinel field (<< 60 km) consistent with a ridge-centred eruption. High percentages of partial melt modelled by REE inversion for the OJP, and affinities with Iceland (Section 8.2.2), support the idea that the OJP erupted as the result of a plume interacting with a spreading ridge (e.g. Mahoney et al., 1993). Melting started deeper in the source region of the OJP than in the source of MORBs, consistent with a thermal anomaly such as a decompressing mantle plume. However, the flat, near-chondritic REE patterns indicate that the depth of **final** melt segregation was shallow, and not beneath a thick lithosphere as in the case of Hawaii.

The Hawaiian islands consist of both alkali and tholeiitic basalts. Tholeiites make up over 95% of the volume of the Hawaiian Islands, voluminously erupted during the main shield building phases (e.g. Clague, 1987). Figure 8.3 shows a selection of REE patterns from the Hawaiian tholeiites, e.g. BHVO-1 (e.g. Govindaraju, 1994). The Hawaiian tholeiites show LREE-enriched patterns, implying that the Hawaiian plume underwent melt segregation largely within the garnet lherzolite field, as expected when a thick lithosphere truncates the melting column at greater depth in an intra-plate setting (Watson and McKenzie, 1991). A schematic of the MVG's melting column and that of Hawaii are shown in Figure 8.4, the Hawaiian melting column starts at a similar depth to that of the MVG, but is cut short by the thickened

lithosphere because of its intra-plate setting. The MVG melting column reaches shallower depths, resulting in the production of extensive tholeiites with flat, near-chondritic REE patterns.



Figure 8.3

Iceland is the closest present-day analogue for the OJP. When plotted on the MORB global trend, the OJP basalts fall at the high $Fe_{8,0}$ and low $Na_{8,0}$ end, similar to averages from the basalts recovered from the Reykjanes and Kolbeinsey Ridges (Mahoney et al., 1993 and references therein). Klein and Langmuir (1987) showed there is a reasonable correlation between locally averaged $Fe_{8,0}$ and $Na_{8,0}$ values and the axial ridge depth. Hence, Iceland plots at the high $Fe_{8,0}$ and low $Na_{8,0}$ values representing the extreme end member of this trend. However, while the OJP's and Iceland's $Na_{8,0}$ and $Fe_{8,0}$ values are similar, this proposed analogue between the MVG tholeiites and Iceland is **not** consistent with field evidence for the eruption environment of the MVG basalts as seen exposed on Malaita.



Figure 8.4. Schematic melting columns for hotspot and ridge-erupted lavas. MORB lavas, the result of shallow melting of cooler mantle, intersect the solidus at lower pressures, while Hawaii and the OJP were erupted as the result of hotter mantle intersecting the solidus at greater depths than MORB mantle (after McKenzie and Bickle, 1988). The exact depth that melting began in the OJP plume is unclear, but is thought to be well within the garnet lherzolite field. The Hawaiian hotspot has its melting column truncated by the \approx 70 km thick lithosphere of the Pacific oceanic crust (Watson and McKenzie, 1991), hence lavas erupted in Hawaii show HREE depletion consistent with a strong garnet influence (Figure 8.3). x= melt fraction, assuming entropy of melting is 400 JK⁻¹kg⁻¹ (after McKenzie and O'Nions, 1991).

The MVG lavas appear to have been erupted in a submarine environment, but the depth of the submarine eruption of the MVG basalts appears to be deep, possibly below or around the Aptian calcium carbonate compensation depth (CCD) (Section 3.5.5). The Aptian CCD is thought to be about 2.8 - 3 km deep (Arthur et al., 1985), far too deep to correlated with the similarities between the Na_{8.0} and Fe_{8.0} values of the MVG and Iceland, Reykjanes and Kolbeinsey Ridges and the latter's shallow axial depths. Several suggestions can be put forward to explain why the eruption appears to be have been so deep. It could be that the Aptian CCD is far shallower than the estimated 2.8 - 3 km. Alternatively, the MVG basalts exposed on Malaita are the results of distal flows, possibly explaining the lack of any clastic material, and they formed on the deeper flanks of a far shallower plateau (Saunders et al., 1993). The idea that portions of the OJP exposed on Malaita are flows erupted from a distal source is also supported by the low geothermal gradient of the MVG basalts (Section 6.5). However, the rate and ages of eruption indicate that the eruption of the OJP was rapid and voluminous (Section 8.5), consistent with a vigorous plume system, probably the melting of a plume head. But there is no obvious reason why the OJP erupted in a submarine setting at all, let alone an apparently deep setting. The elevation caused by a plume surfacing under a spreading ridge, and the extensive shallow melting in the MVG basalts, surely would imply a subaerial eruption environment, such as seen in Iceland.

The Icelandic plume has captured a segment of the North Atlantic ridge, a slow spreading ridge. One possible explanation for the submarine nature of the MVG's eruption may be the involvement of a super-fast spreading ridge, such as portions of the East Pacific Rise, rather than a slow spreading ridge. If the spreading rate was very fast, there may have lacked the lithospheric support which would be required to provide the buoyancy necessary for the OJP to become subaerial.

8.4 Post-emplacement uplift of the MVG

The melting and thermal buoyancy resulting from plumes predict lithospheric uplift, followed by subsidence either as the plume wanes or as the crust is transported away from the plume's influence (Coffin and Eldholm, 1994, and references therein). However, the Alite and Haruta Limestone Formations overlying the basement record a history of **uplift** after MVG eruption, rather than the expected subsidence. The Alite and Haruta Limestone Formations formed in a gradually shallowing environment, as indicated by their increasing portions of carbonate moving up sequence from the Alite to the Haruta Limestone (Chapter 3).

This apparent shallowing of the sediments overlying the MVG may be because the rate of deposition was greater than the rate of subsidence. However, this seems unlikely when we consider that the absolute maximum deposition rate of the MVG sediments was only 2300 m over \approx 100 m.y. This uppermost value assumes that the maximum stratigraphic thicknesses of the Kwara'ae Mudstone Formation is 270m, the Alite Limestone Formation is 1000m and the Haruta Limestone Formation, which ceased around the Mid Miocene, i.e. \approx 15 m.y., is 1000m thick (Figure 3.2). (The Suafa Limestone Formation , which overlies the Haruta, marks the change into shallow marine sediments due to the deformation of the Malaita Anticlinorium and is therefore not included). Furthermore the sediments would cause subsidence effects of their own because of their weight. Calculated subsidence curves for oceanic plateaus are considered to follow a subsidence curve similar to 'normal' oceanic lithosphere (e.g. Parsons and Sclater, 1977; Coffin and Eldholm, 1994). Such curves predict that after 100 m.y., the subsidence will be approximately 3 km (Hayes, 1988) which is greater than the maximum sedimentation rate estimates for the Malaita sediments.

Alternatively, this apparent shallowing of the MVG's overlying sedimentary sequence may be a direct result of the CCD gradually deepening after the Aptian, possibly in addition to the above estimated sedimentation rates. However, if this was the case the CCD would have had to have deepened at a greater rate than the plateau subsided, i.e. at a minimum of 700 m over 100 m.y.. It has been suggested that the depth of the Pacific CCD did fluctuate to deeper levels after about 85 m.y. (Arthur et al., 1985), but these changes are inconsistent with a gradual shallowing in the deposition environment of the MVG's cover sedimentary sequence.

It is difficult to explain the apparent uplift history the plateau experienced, rather than one of subsidence. However, in light of new, younger ages of OJP-like basalts found on Makira (Section 2.8) it is also difficult to constrain when the thermal influence of the OJP plume was removed, hence subsidence curves may be being oversimplified. Also the eruption of the alkali basalt suites on Malaita (the Southern Younger Volcanics and Northern Malaita Alkali Suite) have been related to the OJP passing over the Samoan or Rarotonga hotspots around 44 m.y. (Tejada et al., 1996). Either hotspot may have caused a buoyancy local to portions of the OJP obducted in the Solomon Islands such as Malaita. None the less, the OJP did not subside after basalt eruption had ceased as expected when the thermal influence of the OJP plume was removed.

8.5 Rapid emplacement of the Ontong Java Plateau

It was originally speculated, on the basis of biostratigraphic and palaeomagnetic data, that the OJP erupted in less than 3 m.y., between 121-124 m.y., at the beginning of the Cretaceous Normal Polarity Superchron (Tarduno et al., 1991). Volume estimates for the OJP, calculated from seismic refraction (Furumoto et al., 1976), satellite geoid data and topography (Sandwell and MacKenzie, 1988) were estimated around 40 x 10^6 km³, and age data for the OJP lead to initial calculations for the rate of eruption ranging between 8 and 22 km³yr⁻¹ (Tarduno et al., 1991), with later, slightly more conservative estimates of 12 and 18 km³yr⁻¹ for emplacement on and off ridge respectively (Figure 8.5; Coffin and Eldholm, 1993, 1994). The higher estimates for the rate of eruption of the OJP are comparable with the contemporaneous global production of the entire mid-ocean ridge system (Larson, 1991; Coffin and Eldholm, 1993). Such rapid eruption rates dwarf estimates of eruption rates for the Deccan Traps, i.e. 8.2-2.1 km³yr⁻¹, and make the eruption rates of the Columbia River Flood Basalts (0.9-0.1 km³yr⁻¹) seem very low (Figure 8.5; Coffin and Eldholm, 1994, and references therein). However, the discovery of a later stage (\approx 90 m.y.) magmatic episode (Mahoney et al., 1993) meant that models which speculated that the OJP erupted in 3 m.y. or less had to be revised.

Therefore it seems that the OJP magmatism was dominated by two widespread, major plateau-building events dated at 122 m.y. and \approx 90 m.y., with no evidence as yet of any obvious age progression across the plateau. (A summary of ⁴⁰Ar-³⁹Ar ages determined on samples recovered from the OJP is shown in Chapter 1). This lack of any age progression between the 90 m.y. and the 122 m.y. events hints that the OJP may have been erupted in two distinct magmatic pulses, with the 122 m.y. event probably being the major episode of magmatism and the latter stage \approx 90 m.y. event possibly being related to a reactivation of the plume source (Bercovici and Mahoney, 1994). However, this apparent bimodal age



Figure 8.5

distribution for the OJP lavas has been complicated by recently obtained dates from OJPtype lavas exposed on the island of Makira where two of the OJP-type basalts give ages of 63 m.y. and 34 m.y. (Petterson et al., 1995c; Section 2.8) . The 63 m.y. and 34 m.y events have also been suggested to have been the result of a similar reactivation of the OJP plume, on a \approx 30 m.y. period, as was suggested for the original bimodal 122 and 90 m.y. lavas (Bercovici and Mahoney, 1995; Petterson et al., 1995c). Do these new, younger Makira ages represent a wide-scale magmatic event, such as the 122 m.y. and 90 m.y. events, with correspondingly aged, but as yet undiscovered, samples on the main body of the plateau? This is unlikely as plate motion would surely have removed the plateau from the vicinity of the hotspot. Many questions remain unanswered and the answers to whether the 63 and 34 m.y. events are local or plateau-wide in scale are beyond the scope of this study and will await further drilling and dating on the main body of the OJP.

If the 63 and 34 m.y. Makira lavas represent widespread magnatism on the OJP, then the plume responsible for the OJP appears to have remained active for some 90 m.y. Whether this activity was continuous or sporadic is unclear, but if the OJP lavas were erupted over a period of 90 m.y., then eruption rate calculations for the emplacement of the OJP are obviously substantially reduced to around 0.4 km³yr⁻¹ (based on a volume of 44x10⁶ km³). Such a reduced rate of eruption of the OJP (Figure 8.5e) is comparable to rates calculated for the Columbia River Basalts which, largely because of their subaerial exposure, have a better collection of well constrained ages. If, however, the OJP was erupted as a series of magmatic pulses, occurring periodically every 30 m.y., and each lasting for a period of ~3 m.y., then the interim periods of guiescence would cause calculations of eruption rates to be falsely low. Allowing for such quiescent periods, it can be speculated that the OJP plume was active for approximately 12 m.y. (i.e. 4 periods of ≈3 m.y. activity) of the 90 m.y. age span of lavas recovered from the plateau. If this was the case, when the OJP plume was periodically active, eruption rates may have been as high as 3.3 km³yr⁻¹, an order of magnitude higher than the 0.4 km³yr⁻¹ estimates calculated for a continuous eruption over 90 m.y. Such eruption rate calculations of 3.3 km³yr⁻¹ for the OJP are similar to the rates suggested by Coffin and Eldholm (1993, 1994) for the Deccan Traps.

It needs to be considered, however, that some of the eruption periods may have been
longer, more voluminous and may, if these more voluminous episodes were erupted over longer distances, cover intermediate aged samples. In addition, it is essential to point out that, although the number of well dated OJP lavas from exposed sections within several of the Solomon Islands has recently increased, there remains a serious lack of sampling from the main body of the plateau. To date **no** samples have been recovered from the OJP's eastern salient, which makes up approximately a third of the area covered by the OJP. Therefore, the seemingly variable collection of ages recorded in the basement of the OJP exposed in the Solomons may be but a hint towards a far more complex story.

8.6 Mantle sources

The Ontong Java Plateau was formed as the result of extensive, shallow melting caused by elevated mantle temperatures and high extension rates, the product of decompressional melting of a mantle plume. The chemical components and boundary-layer origin of the plume responsible for the eruption of the OJP are discussed in Sections 8.6.2 and 8.6.3 respectively. Before any attempts are made to characterise the origin and composition of the OJP's mantle plume, the nature of any possible entrainment as the plume traversed the upper mantle needs to be assessed.

8.6.1 Plume-Upper mantle mixing

Positive ENd values and frequently LREE-depleted patterns of basalts produced by higher degrees of partial melting, i.e. tholeiites, are generally thought to have been the result of the melting of enriched plume material mixed with an entrained upper mantle or depleted component. However, Pb studies on Icelandic basalts (Thirlwall, 1995) and depleted diamond inclusions from lower mantle sources (Kesson and Fitz Gerald, 1991) suggest that plumes are heterogeneous with a depleted component which is not the result of melting an entrained depleted upper mantle source, but a distinctive, depleted lower mantle source (e.g. Kerr et al., 1995). I refer to this as an intrinsic component, as opposed to material that has been entrained during passage through the upper mantle.

A simple but effective way of identify the input of a depleted MORB or DMM-like input in plume generated lavas has been recently proposed by Fitton et al (1997) in their studies of volcanic rocks recovered from Iceland and elsewhere in the North Atlantic igneous province.



By the use of a diagram plotting Nb/Y against Zr/Y Fitton et al (1997) were able to distinguish between lavas with a depleted N-type MORB signature and lavas which had a 'depleted' Icelandic mantle plume signature. Figure 8.6, showing the MVG and other Pacific Plateaus and the Tertiary Basalts from Iceland (Fitton et al., 1997) relies on the depletion of Nb in the mantle source of MORB (because of previous melt extraction events) to distinguish MORB basalts from plume sources. As a result of the depletion of Nb, compared to other incompatible elements, MORB plots below the tram lines outlining the neovolcanic zone of Iceland. This broad range from Iceland, labelled the 'Iceland neovolcanic array' in Figure 8.6, encompasses nearly all OIBs.

The MVG basalts plus several of the Pacific oceanic plateaus are plotted on Figure 8.6. None of the plateaus, with the possible exception of the Nauru Basin, show obvious trends towards the MORB data field, consistent with limited entrainment of MORB-type source material in their melting regimes. The MVG basalts and tholeiites from other Pacific Plateaus typically straddle the Bulk Earth (BE) cross wires, while the Tertiary Icelandic basalts do not sit astride the BE cross wires, but plot higher up the parallel array towards more enriched compositions reflecting smaller degrees of melting.

8.6.2 Depleted isotopes but chondritic implications from trace element ratios

The MVG basalts show relatively flat REE patterns (Figure 8.3). These chondritic trace element ratios and flat REE patterns contradict the depleted nature of the MVG's source region implied by the positive ENd isotope values. The depleted nature of the radiogenic isotopes indicates that the source of the MVG basalts has undergone a melt extraction event in its past and therefore cannot be considered a pristine or chondritic reservoir. Therefore is seems most probable that such flat patterns are attributed to their extensive melting within the spinel field, but also retaining a depleted HREE signature because of melting within the garnet stability field, i.e. the bulk source of the basalts has a light-REE-depleted pattern.

However, in addition to the MVG's near-chondritic REE patterns, they also sit astride the BE cross wires on Figure 8.6, and several of the MVG's incompatible trace element ratios, e.g. Ta/Zr, U/Th, are also chondritic (Figure 8.7). Why are some of the trace element ratios in the MVG basalts chondritic? Admittedly not all the trace element ratios are chondritic in nature and there are slight differences between the Type A and Type C-G basalts, but the



differences between the two groups are small. Do chondritic ratios therefore represent fortuitous mixing between enriched and depleted components in the MVG magmas prior to eruption? In the case of an individual plume this well may have been the case but several oceanic plateaus lie along this chondritic array implying a more fundamental reason.

Homogeneous tholeiites, similar to those which dominate the OJP, make up large volumes of most oceanic plateaus, such as Manihiki (Jackson et al., 1976) and Caribbean Plateaus (e.g. Kerr et al., 1996) and the Nauru Basin (Saunders, 1986; Castillo et al., 1991). Large percentages of these tholeiites remarkably also show chondritic trace element ratios similar to the OJP (Figure 8.8). This implies that their mantle source regions may have undergone similar depletion events and/or similar melting regimes. Why this should be the case in several unrelated plateaus is unclear; surely the different melting regimes within each plume would not produce these similar chondritic ratios.

While several oceanic plateaus appear to have some chondritic trace element ratios in common, each plateau typically has a distinctive radiogenic isotope signature. ENd is generally positive (Figure 7.11c), except in the case of the Kerguelen Plateau which has been contaminated with continental lithosphere (Storey et al., 1988, 1989). The different isotope signatures of each plateau, although there is some overlap, implies distinctive sources for the individual plateaus, even though they show similar chondritic trace element ratios. Therefore it may be fair to speculate that these trace element ratios are more a reflection on the melting processes. Possibly, the ancient melt-depletion event which resulted in the positive ENd signatures of the plateaus did not affect some of the trace element ratios. Whatever the cause, the chondritic nature of such ratios is unlikely to be a result of fortuitous mixing between enriched and depleted components.

8.6.3 Plume Origins-D" or 670 km?

Plumes rise from a boundary layer within the mantle, either the core mantle boundary (D") or the 670 km discontinuity (e.g. Campbell and Griffiths, 1990). Higher ³He/⁴He in some Hawaiian (Kurz et al., 1983; Kurz and Kammer, 1991) and Icelandic basalts (Poreda et al., 1986) indicate an undegassed, primordial component within these plumes which has been suggested to come from the lower mantle (e.g. Kurz et al., 1982). However, the extent to which this undegassed mantle mixes with the upper mantle is unclear. Some authors favour



Figure 8.8. The MVG basalts, plus other oceanic plateaus, show C1 Chondrite Zr/Ta ratios (McDonough and Sun, 1995), contradicting the depleted nature of the source displayed by their positive ENd values (Figure 7.11c). The tight linear array of these plateaus is very different from the spread shown by the EPR MORB lavas (A.D. Saunders, unpublished data). This chondritic ratio may be related to melt extraction, either because of an ancient depletion event or during melting of the source region prior to eruption. Data for MVG from this study; Nauru Basin from Saunders; 1986, Manihiki from A.D. Saunders and G.F. Marriner, unpublished data; Kerguelen Plateau data from Storey et al., 1992.

periods of complete mantle overturn and wholesale mantle mixing (e.g. Stein and Hofmann, 1994), while others suggest the actual quantity of flux between the upper and lower mantle to be very small, and plumes which show a high ³He/⁴He signature only require very small amounts of an undegassed, primordial component (O'Nions and Tolstikhin, 1996).



The volume of source material that is required to generate LIPs such as the OJP are large (Coffin and Eldholm, 1994). These volumes are modelled showing maximum and minimum spheres, assuming 5% and 30% partial melting (Figure 8.9), the latter of which is the more realistic estimate of the extent of partial melting in the OJP plume (Section 8.2.2). Although the time scale over which the plateau erupted is not considered, these volume

estimates imply large mantle volumes were required in the production of the OJP, probably with some input from the lower mantle.

Jain et al. (1995) proposed that the OJP plume originated from the core-mantle boundary, indicated by high W values displayed by the MVG basalts. New W and Mo data for the MVG basalts are presented in this study which show **no** such enrichment in the MVG basalts (Appendix Eii). The new data plot close to the average terrestrial values of W/U \approx 0.4 and Nd/Mo \approx 23.4 calculated by Newsom and Palme (1984), similar to that of other OIB, MORB and continental flood basalts (Figure 8.10). It is difficult to account for the discrepancies between the differing W values in the MVG basalts. However, the lower W values for the MVG basalts reported in this thesis are supported by similarly low Mo values. An indication that the mantle source of the MVG has been depleted relative to chondrite values, similar to other plume-related CFBs and MORB basalts.

The depletion in W and Mo compared to chondritic values in terrestrial magmas is due to the siderophile nature of W and Mo which led them to be strongly partitioned into the Earth's core (e.g. Newsom and Sims, 1991). The argument that Jain et al. (1995) use to advocate a core-mantle boundary origin for the OJP plume was that any material originating from the coremantle boundary would be enriched in siderophile elements such as W. However, I suggest an opposing argument that melts originating from the D" layer may preferentially be most strongly **depleted** in W and Mo because of the partitioning of siderophile elements into the core would surely be strongest at the deepest levels of the lower mantle, adjacent to the core itself.

Whether siderophile elements would be enriched or depleted in material originating from the core-mantle boundary is likely to be a direct reflection on the nature of the boundary itself. Seismological data suggests the D" layer has a highly anomalous vertical and lateral structure (Olson et al., 1987; Garnero et al., 1988). Attenuations in the lowermost regions of the D" (Shoore, 1984) may allow incorporation of small amounts of siderophile-enriched material. If this is the case then material ascending in mantle plumes originating from the D" layer may show enrichment in W and other siderophile elements, as suggested by Jain et al. (1995). However the new data presented here does not support this suggestion.





Figure 8.10. The MVG basalts have W/U and Mo/Nd ratios similar to that of other OIBs, MORBs and CFBs. The depletion in moderately siderophile elements such as Mo and W in the Earth's mantle is due to their strong participation of the Earth's core. Data for the MORB, OIB, CFBs, komatiite and andesite are from Newsom and Palme (1984, and references therein). The average terrestrial W/U and Nd/Mo values are as reported in Newsom and Palme (1984) the C1 chondritie W/U and Nd/Mo Nd/Mo values are from McDonough and Sun (1995).

Figure 8.10a&b

8.7 The OJP plume

The narrow range of radiogenic isotopes and incompatible trace element ratios in the OJP basalts initially suggests that the plume source involved may have been internally relatively homogeneous. The few basalt samples recovered from the ODP Site 317 on the Manihiki Plateau (Mahoney and Spencer, 1991) showed far more isotopic and trace element variation than is seen in the upper several kilometres of OJP obducted on Malaita. However, the discovery of alkali basalts related to OJP on Santa Isabel, with a distinct HIMU isotopic signature (Tejada et al., 1996), implies that the source region was heterogeneous. Thus, the voluminous tholeiites which make up the majority (> 99%) of the OJP lavas sampled to date are in fact the result of extensive melting which has been effectively homogenised, either in the melting column or within magma chambers prior too eruption.

While the Sigana Alkali basalts make up a volumetrically minor portion of the OJP, they are important in identifying the enriched component of the OJP plume. The Sigana Alkali Basalts show a distinctive HIMU signature, resulting from small extents of partial melting, as indicated by their enrichment in highly incompatible elements relative to moderately incompatible elements (e.g. Nb/Zr \approx 0.3); and their final depth of melt segregation was well within the garnet lherzolite field, as shown by their depletion in HREE (Sm/Yb \approx 3.8) (Tejada et al., 1996). The origin of the HIMU material within oceanic basalts is thought to be recycled ancient (1-2 Ga) oceanic crust (Hofmann and White, 1982; Zindler and Hart, 1986; Hofmann, 1997). In the case of the OJP plume the ultimate origin of HIMU material may have been the D" layer, which could represent a 'graveyard' of subducted oceanic material.

While the Sigana Alkali basalts may provide clues to the nature of the enriched plume material, the MVG basalts represent the bulk composition of the OJP plume, melting not only the enriched components but also the more refractory, depleted plume components. Mixing on Sr and Nd isotope diagrams (Figure 7.11c) initially suggests that these depleted tholeiites may have been the result of melting an enriched OIB component (i.e. EM-1) with depleted MORB material (Mahoney et al., 1993). However, this depleted mantle component in the OJP appears to be distinct from DMM, as suggested by trace element ratios (Figure 8.6) and possibly Hf isotopes (Section 7.7), and the enriched component in the OJP plume is HIMU. Griffiths and Campbell (1990) and Hauri (1994) have suggested that plumes rising through

the mantle may entrain lower mantle material on their ascent. This entrained material may be a possible source of this depleted component, but what was the ultimate origin of this depleted lower mantle material? As oceanic material descends into the mantle, it is thought that the oceanic crust separates from the lithosphere as it passes down into the lower mantle (e.g. Hofmann and White, 1982). Oceanic lithosphere, consisting of cumulate rocks and residues from melting would be depleted in trace elements (Saunders et al., 1988). Kerr et al. (1995) suggest this depleted oceanic lithosphere may also descend into the lower mantle and subsequently be the source of this depleted lower mantle material.

A schematic cross section for the OJP plume is shown in Figure 8.11, suggesting that the plume consists of enriched streaks in a more depleted refractory matrix or host (cf. Zindler et al., 1984). The off-axis, smaller and deeper alkali basalts are the result of preferentially melting of an enriched, highly fusible material within the plume below thicker lithosphere. Shallower, more extensive melting resulting in the generation of predominately extrusive tholeiitic magmatism also melts the more refractory depleted material.

8.8 The Fate of the OJP plume - Louisville or R.I.P.?

Plumes typically leave tracks in the form of aseismic ridges or seamount chains. However, in the case of the OJP no obvious plume trace has been found. The Louisville plume has been suggested to have been responsible for the formation of the OJP.

The oldest known seamounts of the Louisville Ridge are dated at \approx 67 m.y. (Watts et al., 1988), with older material having been subducted beneath the Australian Plate at the Tonga Trench (e.g. Hawkins et al., 1986). Initial Nd and Sr isotopic ratios of the Louisville basalts (Cheng et al., 1986) overlap with basalts from the OJP (Figure 7.11c), however, the Louisville lavas have higher ²⁰⁶Pb/²⁰⁴Pb values than OJP basalts (Figure 7.11a and b). This may represent a change in hotspot composition in the >70 m.y. period of the now subducted Louisville ridge, possibly a plume head to tail transition (Mahoney et al., 1993), or the melting process is sampling different components within the plume (Saunders et al., 1995). Similar transitions from low to high ²⁰⁶Pb/²⁰⁴Pb values have been recorded in several other oceanic hotspot sources. For example the early Cretaceous lavas of the Kerguelen Plateau have lower ²⁰⁶Pb/²⁰⁴Pb values (17.47), compared to the later hotspot trail lavas >17.9 (e.g. Storey



Figure 8.11

et al., 1989,1992; Weis et al., 1989; Barling and Goldstein, 1990; Weis and Frey, 1991; Saunders et al., 1991). Such a shift could possibly be due to a reduction in supply of low ²⁰⁶Pb/²⁰⁴Pb in the hotspot source itself. A cooler plume tail may be unable to melt refractory low ²⁰⁶Pb/²⁰⁴Pb mantle material especially in an off-ridge environment. This could be the case in the Louisville Ridge as the lavas do not represent high percentage melts; they are alkali basalts, hawaiites and basanites (e.g. Cheng et al., 1986).

Indeed the Sigana Alkali Suite, which represents smaller extents of melting than the tholeiites, do have a high ²⁰⁶Pb/²⁰⁴Pb isotopic composition, similar to HIMU (Tejada et al., 1996). However, this HIMU signature of the Sigana Alkali basalts shows far higher ²⁰⁶Pb/²⁰⁴Pb and ²⁰⁸Pb/²⁰⁴Pb values, than in any of the Louisville seamounts (Figure 7.11a and b). The Louisville seamounts are mostly alkali basalts and basanites, similar to the Sigana Alkali basalts, and may therefore be assumed to have both resulted from similar, small degrees of melting. Hence it is unlikely that the Sigana and MVG tholeiites form two end members of the OJP plume, with the Louisville basalts representing a mixture of the two.

Directly linking the OJP basalts to a presently active hotspot remains an enigma, especially as not all plate reconstructions for the Cretaceous place the OJP as far south as the Louisville hotspot 120 m.y. ago (Yan and Kroenke, 1993). Therefore, it seems likely that the hotspot which created the OJP is no longer active.

8.9 Concluding Remarks

• Type A basalts trend towards higher Zr/Y, Zr/Sc and Sm/Yb values. This is consistent with melt segregation typically occurring at greater depths (with garnet having a greater influence) in the Type A basalts than in the Type C-G basalts. Alternatively, the Type A magmas precipitated larger amounts of clinopyroxene prior to eruption than the Type C-G magmas. The former of the two is most favoured as the Type A lavas also show a slight enrichment in highly incompatible elements relative to moderately compatible elements, as well as slightly higher Fe_{8.0} and Na_{8.0} values.

 The maximum extent of partial melting which formed the OJP has been estimated by REE inversion modelling (McKenzie and O'Nions, 1991) to be ≈ 30%, similar to that modelled for Icelandic basalts. • Regardless of the small differences between Type A and Type C-G lavas, **all** of the MVG basalts show relatively flat to slightly LREE-depleted REE patterns and low Zr/Y and Zr/Sc values. This is consistent with extensive amounts of melt being generated at shallow levels, i.e. mostly within the spinel Iherzolite field. Extensive melting at shallow levels in the MVG basalts is consistent with a ridge-centred eruption. However, melting in the MVG's source started in the garnet Iherzolite field, deeper than in a typical MORB source, consistent with a thermal anomaly such as a decompressing mantle plume.

• The MVG's and Iceland's Na_{8.0} and Fe_{8.0} values are similar, both plotting at the high Fe_{8.0} and low Na_{8.0} end of the MORB global trend (Klein and Langmuir, 1987). However, the proposed correlation between locally averaged Fe_{8.0} and Na_{8.0} values and the axial ridge depth between the MVG tholeiites and Iceland is **not** consistent with field evidence for the eruption environment of the MVG basalts as seen on Malaita.

• Malaita erupted in a deep submarine environment, possibly below the Aptian CCD. There is no obvious reason as to why the OJP erupted in a submarine setting at all, let alone its apparently deep setting. The elevation caused by a plume surfacing under a spreading ridge, and the extensive shallow melting in the MVG basalts, surely would imply a subaerial eruption environment, such as seen in Iceland.

• The MVG's overlying sediments record a history of **uplift**, after MVG eruption, rather than the expected subsidence. This uplift history for the OJP is as unexpected as its deep submarine eruption environment.

• The OJP magmatism appears to have been dominated by two widespread, major plateau-building events dated at 122 m.y. and ≈90 m.y., with no evidence as yet of any obvious age progression across the plateau. However, sampling on the main body of the plateau remains sparse, and to date **no** samples have been recovered from the OJP's eastern salient, which makes up approximately a third of the area covered by the OJP.

• Trace element ratios and Hf isotopes imply that the entrainment of MORB-type source material into the melting regime of the OJP plume may have been limited, hinting that the depleted component that predominates in the MVG basalts may be intrinsic to the plume itself.

• Several of the MVG's incompatible trace element ratios, e.g. Ta/Zr, U/Th, are the same as chondrite. These chondritic trace element ratios and flat REE patterns contradict the depleted nature of the MVG's source region implied by the positive ENd isotope values. The chondritic nature of such ratios is unlikely to be a result of fortuitous mixing between enriched and depleted components as several other plateaus, i.e. Manihiki, Nauru Basin and Caribbean Plateaus also show similar chondrite ratios. Therefore it is more likely these chondritic trace element ratios are a reflection on the melting processes. Possibly, the ancient melt depletion event that resulted in the positive ENd signatures of the plateaus did not affect some of the trace element ratios.

• The large mantle volumes required to undergo partial melting implies the possible involvement of a lower mantle component in the source of the OJP.

• The OJP plume is thought to have consisted of enriched streaks in a more depleted refractory matrix or host. The off-axis, Sigana Alkali basalts are the result of preferentially melting of an enriched, highly fusible material within the plume below thicker lithosphere. The shallower more extensive melting involves the more refractory depleted plume material, resulting in the generation of predominately extrusive tholeiitic magmatism. The ultimate origin of the enriched HIMU streaks may have been recycled ancient oceanic crust, possibly originating from the D" layer. The depleted plume material may have been entrained as the plume ascended through the lower mantle, possibly recycled ancient oceanic lithospheric material.

• Directly linking the OJP to a currently active hotspot remains an problem. The Louisville plume has been suggested to have been responsible for the formation of the OJP, though differences in ²⁰⁶Pb/²⁰⁴Pb values between OJP lavas and Louisville basalts indicate this is unlikely. Also not all plate reconstructions for the Cretaceous place the OJP as far south as the Louisville hotspot 120 m.y. ago. Therefore, it seems that the hotspot which created the OJP is no longer active.

Appendix A River Traverses (in back pocket)

- Appendix Ai Singgalo River.
- Appendix Aii Kwaleunga River.
- Appendix Aiii Ala'olo River.

Appendix B Analytical techniques

Appendix Bi	-	Sample preparation
Appendix Bii	-	X-ray fluorescence techniques
Appendix Biii	-	Instrumental neutron activation techniques
Appendix Biv	-	Isotope dilution techniques
Appendix Bv	-	X-ray diffraction techniques
Appendix Bvi	-	Electron microprobe techniques
Appendix Bvii	-	ICPMS techniques

Bi - Sample Preparation

Standard preparation techniques were used. Weathered rinds, mineralised veins and amygdales were avoided where possible. The rocks were progressively ground to a fine powder using firstly a fly-press and then an agate-lined barrel on a Tema[®] swing mill.

Bii - X-ray fluorescence (XRF)

Major Elements

Rock powders were dried overnight at 120°C prior to igniting in a muffle furnace at 950°C for 1.5 hours. Samples were cooled for 1.5 hours in a dessicator before re-weighing to determine total ignition loss. The distribution of ignition loss in the total sample suite can be gauged from the histogram (Figure 6.7).

All major elements were measured on fusion beads. 1 gram of ignited rock powder was mixed with 5 grams of dry lithium tetraborate-metaborate flux in a *precise ratio* of 1:5. The mixture was fused in a vertical tube furnace for approximately 15 minutes at 1200°C, in a Pt-Au crucible. Periodic swirling of the crucible was carried out to ensure homogeneity of melt. Previous studies have shown that volatile losses (Na, K) under these conditions are negligible. The

glass beads were formed on a heated aluminium platten prior to annealing at about 250°C for 12 hours. The cooled beads were then sealed in plastic bags to prevent hydration before analysis.

Trace Elements

All XRF trace elements were measured on pressed powder pellets. Approximately fifteen grams of powder were thoroughly mixed with 8-12 drops of the binding agent Mowiol (a 7% solution of Polyvinylpyrrolidine/methyl cellulose) bonded under pressure of 15 ton. in⁻² to form 46 mm diameter pellets, which were subsequently dried overnight. The hydraulic press is faced with polished steel dies to provide a smooth pellet surface.

Machine Conditions

Samples were analysed for major elements and selected trace elements (Ni, Cr, V, Zn, Ga, Rb, Ba, Nb, La, Ce, Sr, Nd, Zr, Ti, Y, Th) at the University of Leicester using wavelengthdispersive x-ray fluorescence (WD-XRF) techniques. Analysis of major and trace elements was made on either a Philips PW1400 or an ARL 8420+ wavelength-dispersive XRF-spectrometer. Both machines use a rhodium-anode X-ray tube for major elements and for the majority of the trace elements, although the Philips also used a W-anode tube for some of the trace elements. Standard tables are given in Appendix Ci, machine running conditions are given in Appendix Cii, and comparisons between XRF, INAA and ICP-MS (ANU) can be found in Appendix Ciii. Fused discs which gave totals out 98.5 to 101.5% were re-fused and re-analysed.

Biii - Instrumental Neutron Activation (INAA)

Rare-earth element (La, Ce, Nd, Sm, Eu, Gd, Tb, Yb, Lu) and trace element (Th, Ta, Hf, and Sc) abundances were determined on selected samples by Instrumental Neutron Activation Analysis (INAA) at the University of Leicester. Full details of the techniques employed are summarised below.

Sample preparation

Rock powders were dried overnight at 120°C, and a 200 mg aliquot of each sample was weighed into a polythene vial, which was heat sealed by soldering. Batches of nine or ten individual vials were stacked into a polythene tube, with standard powders placed at the top, middle, and bottom of the tube. The standards used were irradiation calibration standards JB-1A

and an internal standard Whin Sill basalts. The neutron flux gradient across the tube during irradiation was monitored via weighed discs of iron foil placed between the vials in the tube. Each foil was coated with plastic in order to prevent corrosion during irradiation.

Irradiation was carried out in the 270°C core tube facility at Silwood Park, Ascot (University of London Reactor Centre). Samples and iron foil flux monitors were given a thermal neutron flux of approximately 9.4×10^{16} n. cm⁻². s⁻¹ by irradiating in tube 4, positions 3 and 4, for 3 days (equivalent to 3 x 7.5 hrs of reactor time). After six days of 'cooling' to allow for the decay of short-lived isotopes (predominantly ²⁴Na), the samples were counted at the University of Leicester.

Detectors and peaks of interest

Two detectors are used: a high-energy GEM (Li) coaxial detector, with a resolution of 1.67 keV at 1.33 MeV (FWHM), for measuring La at 1596.2 keV and Sc at 889.3 keV; and a LOAX (planar low-energy photon) detector with a resolution of 299 eV at 5.9 keV and 563 eV at 122 keV (FWHM) for Sm (103.2 keV), Nd (91.0 keV), Eu (123.1 keV), Tb (86.8 keV), Tm (84.3 keV), Yb (63.1 keV), Lu (113.0 or 208.4 keV), Ta (67.7 or 100.1 keV), Hf (133.0 keV), and Th (98.4 keV).

Data collection and processing

The equipment used is EG&G hardware, with an MCA buffer emulator installed on an IBM 30 computer. Data from the two detectors are channelled through a multiplexer prior to storage in the MCA buffer. Spectra are transferred to the VAX for peak stripping and peak-area computation using SAMPO software. Peak intensities are compared to those of known reference standards (AC-2 for most elements, JB-1A for Sc) for calculation of element abundances. Inter-element corrections are made where necessary (see Potts *et al.*, 1985), although it was found that Tm data were excessively high and were ignored.

Monitoring of data quality

Standards JB-1A and an internal standard Whin Sill basalts are used to check the accuracy and precision of the data. These are given in Appendix Ciii, and the REE patterns in Appendix Civ (Page C.20). Comparisons with ICP-MS data run at ANU is good, though separate rock splits were used (Appendix Civ; Page C.21).

Biv - Isotope dilution techniques

Samples were prepared for isotope analysis by washing in de-ionised water and then crushing in a WC mill. Agate milled powders were not used due to possible Pb contamination. All radiogenic isotope geochemistry on the Malaita Volcanic Group samples was carried out at the NERC Isotope Geosciences Laboratory (NIGL) complex, Keyworth, Nottingham.

Acid leaching procedures

The samples were acid leached in order to remove possible secondary phases. Approximately 300 mg of rock powder was weighed into a Teflon beaker previously cleared of static by an anti-static pistol. Each sample was mixed with approximately 6 ml of 6M HCl (except MB8A to which only 2 ml was added). The beakers were then left on a hotplate at \approx 120 °C for 1 hour (except for MB8A which was removed after 30 minutes). The leachate was removed by automatic pipette using a clean tip each time; The leachates were analysed for samples from batches P66 and P67, but discard for batches P81 and P82. After leaching the sample was washed by adding MQ H₂O and allowed to stand to allow any sample in suspension to settle out before being pipetted off and dried overnight on a low temperature hotplate. The dried samples were weighed (about 0.1 g of leached sample used), and approximately 0.01 and 0.02 g of Sr spike (Oak ridge dilute or NBS988), 0.03 g of mixed Sm/Nd spike and 0.1 g of U/Pb spike was added. International reference standard, BHVO, and a blank were subjected to the same procedure as each of the samples.

Rock Digestion

Rock digestion was performed by adding ~0.5-1 ml of 16M HNO₃ and 2-3 ml of TD 29M HF. The samples were shaken and left overnight. The HNO₃-HF mixture was evaporated on a hotplate at \approx 100 °C, before a further 2-3 ml of 16M HNO₃ was added; the samples were shaken again and left to evaporate slowly overnight at < 80 °C. After the HNO₃ was evaporated, 2-3 ml 6M HCl was added, and left on the hotplate with the lid on to reflux for at least an hour. If the solution was clear the sample was dried down, if not it was placed in an ultrasonic bath for half an hour before another round of 16M HNO₃ (left overnight, sealed on the hotplate) and 6M HCl was added (again the beakers were sealed and placed on the hotplate to reflux). The maximum any sample needed was two rounds of acid. When the solutions are clear, the HCl is evaporated off, ~1 ml of 1M HBr is added, and allowed to stand cold overnight.

Pb chemistry

The Pb columns are made of 1 ml polypropylene pipette tips with a small polyethylene or polypropylene frit (35μ pore size: 1/8" thick) in the tip. Column procedure was as follows: (i) Pipette tip and frit assembly were placed in a column rack in a fume cupboard with a wash beaker underneath. One column full (1 CF) of QD 6M HCl was added to the column. One CF of MQ was added and left to drain. Following this, 4-5 drops ($15-25 \mu$ l) of resin were added from a dropper bottle. This was washed in with 1 CF TD 6M HCl, followed by 1 CF MQ; the HCl and MQ washes were then repeated. The columns were preconditioned with 0.5 CF 1M HBr; (ii) a clean beaker was placed under each column in order to collect Sr and Nd.

(iii) each HBr-sample solution was sucked up with a clean pipette, leaving any residue in the beaker. The solution was loaded onto the column and left to drain through. Following this, the column was washed with 0.5 ml of 1M HBr and a repeat wash of 1 ml of 1M HBr was made;

(iv) once the HBr washes had drained through, the beaker with the Sm/Nd and Sr fractions were removed, 2 ml of TD 16M HNO3 was added and the samples placed on a hotplate and evaporated to dryness to remove excess bromine (see below). A clean beaker was placed under the column, and Pb collected by adding 1 CF of TD 6M HCl to the column. To the collected HCl were added 3-4 drops of HNO3 from the squeeze bottle (to eliminate excess bromine). The collected sample was then evaporated to dryness;

(v) 0.5 ml of 1M HBr were added to the sample beaker, and the column procedure repeated (without collecting the HBr this time). After the second pass through the columns, 1 drop of phosphoric acid (H_3PO_4) was added to the collected HCl-sample solution and HNO₃. The solution was evaporated to near-dryness on a hotplate, leaving a small drop of liquid in the bottom of the beaker.

Pb chemical methods, as outlined above, were run following the procedures reported in *Common Pb chemical procedures for silicate rocks and minerals, methods of data correction and an assessment of data quality at the NERC Isotope Geosciences Laboratory*, NIGL report series No. 78 by P.D. Kempton (1995).

U chemistry

The U column procedure is carried out once the samples have been through the first set of Pb columns. The columns are set up using ≈ 0.5 ml of pre-cleaned AG 1 X 8 resin.

(i) Each of the columns was washed with 1 ml H_2O , and this washing procedure was repeated another 5 times. The H_2O was allowed to elute into plastic waste beakers.

(ii) 1 column full (CF) of 6M HCI was added to precondition the columns and allowed to elute.

(iii) the U sample was taken up in 6M HCl and added to the columns. A clean pipette was used for each sample.

(iv) 0.5 ml of 6M HCl was added to the columns and allowed to elute.

(v) 1 ml of 8M HNO₃ was added to the columns and allowed to elute.

(vi) 1 ml of 8M HNO₃ was added to the columns and allowed to elute.

(vii) 1 ml of 6M HCl was added to the columns and allowed to elute.

(viii) 1 ml of 6M HCl was added to the columns and allowed to elute.

(ix) the wash beakers were replaced with pre-cleaned U collection beakers.

(x) 1 ml of H_2O was added to the column and collected.

(xi) a second 1 ml of H_2O was added to the columns and collected.

This 2 ml of H₂O containes the U fraction. The U fraction was dried down and a second stage of U column chemistry was carried out. Again the columns were set up using \approx 0.5 ml of pre-cleaned AG 1 X 8 resin.

(i) The columns were washed with 1 ml of H_2O and allowed to elute into wash beakers placed beneath the columns.

(ii) 1 ml of 1M HCl was added and allowed to elute.

(ii) 1 ml of 1M HCl was added and allowed to elute.

(iii) 1 ml of 1M HCl was added and allowed to elute.

(iv) 1 ml of H₂O was added and allowed to elute.

(v) 1 ml of H₂O was added and allowed to elute.

(vi) 1 ml of 8M HNO₃ was added and allowed to elute.

(vii) 1 ml of 8M HNO₃ was added and allowed to elute.

(viii) 1 ml of 8M HNO3 was added and allowed to elute.

(ix) the sample was taken up in 1 ml of 8M HNO₃, added to the column using clean pipettes for each sample.

(x) 1 ml of 8M HNO₃ was added to the column and allowed to elute.

(xi) 1 ml of 8M HNO₃ was added and allowed to elute.

(xii) 1 ml of 8M HNO₃ was added and allowed to elute into the wash beakers. The wash beakers were then replaced with pre-cleaned U beakers

(xiv) 1 ml of H₂O is added to the column and collected.

(xv) 1 ml of H₂O is added to the column and collected.

(xvi) 1 drop of H_3PO_4 was added to the collected U fraction. The sample was placed on a hotplate and evaporated to near-dryness.

U chemistry procedures, outlined above, were run following the methods reported in *Procedures for the chemical analysis of U in silicate rocks at the NERC Isotope Geosciences Laboratory*, NIGL report series No.117 by P.D. Kempton (1997).

Sm, Nd, Sr and Rb column chemistry

The dried residue (part iv above) was taken up with 2 ml 2.5M HCl, left for 15-20 minutes, then centrifuged for 5 minutes. Samples were added to the Sr columns, taking care to leave behind any residue. The following column procedure was carried out;

(i) After the sample had been added, 1 ml of 2.5M HCl was added to the columns, allowed to elute though into waste beakers, and repeated.

(ii) 16 ml of 2.3M HCl was added, allowed to elute.

(iii) When Rb was collected (batch P81) the waste beakers were removed and replaced with clean beakers, 5 ml of 2.5M HCl was added to the columns and collected in the clean beakers. The beakers were then removed from the columns, and evaporated on a hotplate. The waste beakers were then returned to the columns.

(iv) 19 ml of 2.5M HCl was added to the columns and allowed to elute.

(v) The waste beakers were removed and replaced with clean beakers, 10 ml of 2.5M HCl was added to the columns and collected, this contained the Sr fraction. The beakers containing the Sr fraction was removed from the column and evaporated to dryness.

(vi) The waste beakers were replaced, and 9 ml of 6M HCl was added to the columns and eluted.

(vii) The waste beakers were replaced by clean beakers, 11 ml of 6M HCl was added to the columns and collected, this contains the Sm and Nd fractions. The collected fraction were removed from the columns and dried down.

The Sm and Nd fractions were taken up by 1 ml of 0.2M HCl and pipetted onto the REE columns using the following procedure.

(i) After the sample was added to the column, 1 ml of 0.2M HCl was added to the columns and eluted into a waste beaker, and repeated.

(ii) 9 ml of 0.2M HCl was added to the columns and eluted.

(iii) The waste beakers were replaced with clean beakers, 3 ml of 0.2M HCl was added to the columns and collected, this column wash contained the Nd fraction. The waste beakers were replaced.

(iv) 2 ml of 0.2M HCl was added to the columns and eluted.

(v) 2 ml of 0.2M HCl was added to the columns and collected, this contained the Sm fraction.

(vi) Both the Sm and Nd fractions were dried down.

Measurement of isotope ratios

Sr, Nd, and Pb isotopic ratios were analysed using two different multi-collector mass spectrometers in static mode: a Finnigan MAT 262 and a VG 354.

Blank chemistry

The total procedure blanks for Sr and Nd analyses were negligible, a maximum of 625 pg for Sr, 263 pg for Nd and 75 pg for Sm. Pb blanks were less than 217 pg and U blanks less than 7 pg.

Reference standards

Each set of samples was run with the international reference standard, BHVO-1, in order to monitor variations introduced during chemistry procedures. External precision for the Sr standard NBS 987 was 0.710180 ± 14 (1 sigma error) for batches P66 and P67, and for batches P81 the value was 0.710193 ± 14 (1s). During the same period, runs of the Nd standard J &M gave an average value of 0.511129 ± 6 (1s) for batches P66 and P67, and for batches P81 the value was 0.511125 ± 7 (1s) on the VG 354 mass spectrometer. In addition Nd was run on the MAT 262 mass spectrometer for batch P82, which gave values of 0.511101 ± 20 (1s).

The Pb standard NBS 981 gave values of 19.904 ± 14 ($^{206}Pb/^{204}Pb$), 15.439 ± 14 ($^{207}Pb/^{204}Pb$), and 36.532 ± 38 ($^{208}Pb/^{204}Pb$). This amounts to an errors of 0.1% on Pb analyses. Based on 8 replicate analyses of BHVO-1, the U values are 0.417±1, and Pb concentration values are 1.93 ± 5 . The U/Pb ratio is 0.217 ± 5 , which is accurate to $\approx2\%$.

Hf isotopes

Hf isotopes were run using the procedure outlined in *Procedure for high precision isotope analysis of Hafnium by static multicollection thermal ionisation mass spectrometry*, NIGL report series No. 99 by G.M. Nowell (1996). The Hf standard, JMC 475, gave a value of 0.282155±18 (2 sigma error, n=65) when samples KW6, KW13, TLB25, and MB2 were run and a value of 0.282155±6 (2s, n=26) when MB6 and ML219 were run. All Hf data (shown in Appendix F) were normalised to a JMC 475 value of 0.28216.

Bv - X-ray diffraction

X-ray diffraction (XRD) was carried out at the University of Leicester. XRD analyses were on clay fractions (< $2 \mu m$) from selected samples, the results can be found in Section 6.2.

Sample preparation

Whole rock powders crushed by agate mill were used in clay separation (< 2 μ m) techniques. The powder was first mixed with a de-flocculation agent (dispex or sodium hexametaphosphate) and then allowed to settle over night. This allows the sample and the clays to become wetted, and thus allow them to break down into single clay platelets. The next day the samples were re-shaken and the clay fractions were subsequently separated from the coarser fractions by centrifuging for 3.5 minutes at 730 rpm. After centrifuging the liquid containing the < μ m fraction is poured into another tube and then centrifuged for a further 30 mins at 3000 rpm or until the clay particles are concentrated into a slurry at the bottom of the tube. The clay fractions were then pipetted onto a glass plate and allowed to dry in air. The glass plates were placed in the Al holders, and XRD analyses were carried out on these air dried samples (traces labelled as air-dried in Figure 6.1).

After XRD analysis was ran on the air-dried samples, the glass plates were placed in a dessicator containing ethylene glycol at 60°C overnight. These glycolated samples were re-run using the same XRD operating conditions as for the air dried run (Glycolated traces in Figure 6.1). The next stage was to heat the samples in a furnace at 330°C for one hour, before analysing again (Heated 330 traces in Figure 6.1). Finally, the samples were heated to 550°C for one hour and re-run (Heated 550 traces in Figure 6.1). Hence, four XRD scans were generated from the clay fractions; air dried, glycolated, 330°C, 550°C.

Analytical conditions

The XRD analyses were carried out using a Philips PW 1729 X-ray generator equipped with a PW 1710 computerised control unit. The machine was operated at 40 kV and 30 mA, using Cu-K alpha radiation and a scanning speed of 1^o 20/min.

Bvi - Electron microprobe

Electron microprobe analyses were carried out at the University of Leicester on a JEOL JXA-8600 Superprobe running a wavelength dispersive system. The samples were analysed using a general silicate programme which was calibrated to the following standards; natural wollastonite for Si and Ca, natural rhodonite for Mn, natural jadeite for Na, natural rutile for Ti, and synthetic standards for Al, Fe, Mg, Cr, Ni and K. The operating voltage was 15 kv, the probe current was 3×10^{-8} A. Beam diameters used were < 1 to 10 µm (for clay minerals) and 10 µm (for feldspar, clinopyroxene, olivine and oxides). The analysed element was measured with a peak count of 20 seconds and a background count of 20 seconds. Estimated precision at 2s for the common petrogenetic minerals is ± 0.25 % at 50 wt % oxide (calculated from standard deviations of raw intensity data).

Bvii - Platinum Group Elements

ICP-MS was carried out on several samples from Malaita, for trace and rare earth elements, including platinum group elements and Au, at the Australian National University (ANU) at Canberra as outlined below.

TRACE ANALYSES OF ONTONG-JAVA BASALTS

Trace element analyses by solution nebulization-ICPMS were carried out using the procedures described in Eggins et al. (1997). Dissolution methods, ICPMS running conditions and the results for several standard rock reference materials are given in that work.

RE, AU, PGE ANALYSES OF ONTONG-JAVA BASALTS

Sample dissolution and preconcentration

Analyses for Re, Au and the Pt-group elements (PGEs) were made by solution nebulization-ICPMS following their separation from matrix elements and preconcentration using a noble-metal-specific chelating resin (Koster and Schmuckler 1967, Nadkarni and Morrison 1977).

The resin consists of isothiouronium functionalized chloromethylated copolymer of styrene and divinylbenzene in chloride form and is marketed in the form of spherical beads as "lonac SR-3" by Sybron Chemicals Inc. of Birmingham, New Jersey, USA. Isotope dilution procedures were employed for Re and the polyisotopic PGE. Concentrations for Au and Rh, which are monoisotopic, were measured by reference to an external calibration standard. Gold is almost completely (>99%) adsorbed by the resin but Rh is not (Nadkarni and Morrison 1977), requiring a yield correction of 25%. The detailed analytical methodology is described in Sylvester (in prep.).

For the Ontong-Java samples, five to six grams of rock powder were dissolved using a Milestone (mls 1200 mega) 6-pressure-vessel laboratory microwave. Each sample was weighed on a microbalance and placed in a microwave vessel along with a small amount of a mixed spike containing enriched isotopes of ⁹⁹Ru (96.63% ⁹⁹Ru, 0.69% ¹⁰¹Ru), ¹⁰⁵Pd (96.58% ¹⁰⁵Pd, 0.46% ¹⁰⁸Pd), ¹⁸⁵Re (97.40% ¹⁸⁵Re, 2.60% ¹⁸⁷Re), ¹⁹¹Ir (96.188% ¹⁹¹Ir, 3.812% ¹⁹³Ir) and ¹⁹⁵Pt (94.91% ¹⁹⁵Pt, 2.24% ¹⁹⁴Pt). The enriched isotope spike was prepared from metals obtained from Oak Ridge National Laboratory, Tennessee, USA. Three acid dissolution/ evaporation steps were carried out on each sample; first, with a mixture of concentrated HNO₃ and HF acids, then with a concentrated HNO₃-HCL acid mix, and finally with 4N HCL. The 4N HCL solutions were centrifuged to remove any gelatinous residue and evaporated to incipient dryness in clean 25 ml Savillex beakers. A procedural blank was prepared in the same way as the samples.

Approximately 10-20 mg of the chelating resin and 3 ml of 0.05 N HCL were added to the beakers containing the dried residues of the digests. The beakers were ultrasonicated and then set aside for 1 week to allow sufficient time for wholesale adsorption of Re, Au and the PGEs onto the resin. Each day, the contents of the beakers were swirled gently to promote contact between the entire solution and resin. At the end of the week, the solution and any sample precipitate present were removed from the beakers with a pipette. The resin remaining in the beakers was rinsed with Milli-Q water and then a dilute HCL-HNO₃ acid mixture. The rinses remove from the resin a large fraction of adhering base metals and high-field strength elements, which can cause particularly severe molecular interferences on the noble metals during ICPMS analysis.

To prepare the resins for introduction into the ICPMS as a solution, each was dissolved in 3 ml of concentrated HNO₃ acid using the laboratory microwave. The digests were then transferred back to Savillex beakers along with 4 ml of 4N HCL used to rinse out each microwave vessel. After evaporation, five to six ml of 1N HCL containing 10 ng/g of each of Be, In and Bi was added to each sample. The Be, In and Bi served as internal standards during ICPMS analysis to correct for variations in the sensitivity of the instrument over the course of the measurement run. Contributions of Be, In and Bi from the samples are negligible as the resin does not quantitatively adsorb these elements. For external calibration of Au and Rh concentrations, and measurement of instrument mass bias factors for isotope ratios of Re, Pd, Pt, Ru and Ir, a 1N HCL standard solution containing known concentrations (~20 ng/g) of each element was prepared from concentrated (100 ug/ml), ultra-high purity, single-element, 10% HCL solution standards obtained from Plasma Chem Corp. of Farmingdale, New Jersey, USA. For calibration of the concentrations of enriched ⁹⁹Ru, ¹⁰⁵Pd, ¹⁸⁵Re, ¹⁹¹Ir and ¹⁹⁵Pt in the mixed spike, needed for the isotope dilution calculations, a dilution of the concentrated mixed spike was prepared in 1N HCL with a carefully weighed aliquot of the Au-Re-PGE standard solution.

ICPMS Analysis

The ICPMS used in this work is a Fisons VG PlasmaQuad PQ2-XR. Operating conditions are shown in Table B.1. All solutions to be analyzed were poured into clean, 12 ml plastic tubes and introduced into the ICPMS using a Gilson autosampler. Solutions were introduced into the ICPMS at a rate of ~0.75 ml/min. To ensure that analyte signals had become stable prior to the start of analysis, counts-per-sec data collection was delayed until 2 min after the start of solution uptake. Data were collected over 4 separate, 60-sec acquisitions. Between analyses of different sample solutions, the ICP-MS was washed out by the introduction of a dilute surfactant (0.5% Triton X-100) for 5 mins, 1N HCL for 4 mins, and 2% HNO₃ for 10 mins. The long washout times were necessary to return instrument blank levels for the elements of interest to those present at the start of the experiment.

Analyte isotopes, dwell times and corrected and monitored isobaric interferences are listed in Table B.2. To minimize counting statistic errors on Re, Au and the PGEs, dwell times employed for them were much longer than those used for potentially interfering elements. Corrections for the blank, isobaric interferences, and drift in instrument sensitivity over the course of the experiment were applied to raw measured cps data on-line using the manufacturer's software. Blank subtractions for each of the Re, Au and PGE isotopes measured amounted to <10% except for ¹⁰³Rh in TLB13, TLB14, TLB25 (20-30%), ¹⁹³Ir in ML543, TLB13, TLB14, TLB25 (15-30%) and the Ru isotopes in all samples (20-70%). Molecular interference subtractions were made using data for oxide, chloride and argide production collected on solutions containing potential interfering elements but not Re, Au or the PGEs. These solutions were analyzed prior to measurement of the unknowns. The subtractions were less than 10% except for ZrO on ¹⁰⁸Pd in ML543, TLB13, TLB14, TLB25 (15-25%). Drift corrections were made by linear interpolation between the intensities of the measured masses of Be, In and Bi. These corrections were less than 30%. After application of the corrections, concentration calculations for Au and Rh were carried out on-line by reference to a linear relationship between concentration and analyte intensity defined by the external calibration standard and the origin.

Isotope dilution calculations for Re, Pd, Pt, Ru and Ir were carried out off-line using the corrected cps data for the unknowns and the enriched spike calibration solution. Instrument mass bias corrections of 3.7% for ⁹⁹Ru/¹⁰¹Ru, 5.1% for ¹⁰⁵Pd/¹⁰⁸Pd, 1.6% for ¹⁸⁵Re/¹⁸⁷Re, 2.1% for ¹⁹¹Ir/¹⁹³Ir and 1.6% for ¹⁹⁵Pt/¹⁹⁴Pt were applied to the data. The bias factors were determined by the difference in the measurements of those ratios in the Au-Re-PGE standard solution and their accepted natural values (⁹⁹Ru/¹⁰¹Ru=0.743, ¹⁰⁵Pd/¹⁰⁸Pd=0.8439, ¹⁸⁵Re/¹⁸⁷Re=0.5974, ¹⁹¹Ir/¹⁹³Ir= 0.595, ¹⁹⁵Pt/¹⁹⁴Pt=1.027) (Handbook of Chemistry and Physics 1992-1993). For both the external calibration and isotope dilution calculations, a concentration was calculated for each of the 4 acquisitions made on each unknown. The mean concentrations of each 4 acquisitions are reported and their standard deviations (1s) are used as a measure of the precision on the measurements. Results for two standard reference materials, analyzed in the same way as the unknowns, are shown in Table B.3.

Forward power	1350 W
Reflected power	2 W
Ar gas flow rates	
nebulizer	0.93 l/min
coolant	14 l/min
auxillary	0.4 l/min
Nebulizer	Quartz V-groove
Spray chamber	Scott-type, double-pass
Cones	Ni
Chamber pressure	
expansion	2.0 mbar
intermediate	1 x 10 ⁻⁴ mbar
analyzer	5 x 10 ⁻⁶ mbar
Ion lens settings	
extraction	3.5
collector	7.4
L1	8.5
L2	1.3
L3	6.3
L4	1.0
pole bias	5.0
Data acquisition	
mode	peak hopping
points per peak	3
dwell time	2.56-20.48 msec
quad settle time	5 msec
analysis time	4 replicates of 60 secs each

 Table B.1. Operating conditions for Fisons VG PlasmaQuad PQ2-XR ICPMS.

Appendix B: Analytical techniques

Element	Isotope	Dwell Time	Corrected Interferences	Monitored Interences
	Measured	(msec)		
Be	9	10.24		
Ti	49	2.56		
Ni	60	2.56		
Cu	65	2.56	⁴⁹ Ti ¹⁶ O	
Zn	66	2.56		
Sr	86	2.56		
Zr	90	2.56		
Mo	95	5.12	60Ni35Cl	
Ru	99	20.48	64Zn35Cl	62Ni37Cl
Ru	101	20.48	66Zn35Cl	40Ar61Ni
Rh	103	20.48	40Ar63Cu, 68Zn35Cl	87Sr16O
Pd	105	20.48	40Ar65Cu, 68Zn37Cl	88Sr16OH
Pd	108	20.48	92Zr16O, 92Mo16O	108Cd
Cd	111	5.12		
In	115	10.24		
Hf	178	5.12		
Та	181	5.12		
Re	185	20.48		
Re	187	20.48		187Os
Os	189	20.48		
lr	191	20.48		174Hf16OH
lr	193	20.48	177Hf16O	
Pt	194	20.48	178Hf16O	
Pt	195	20.48	179Hf16O	
Au	197	20.48	180Hf16OH, 181Ta16O	
Bi	209	10.24		

Note: Be, In, Bi are internal standard elements

 Table B.2. Analyte isotopes, dwell times, and polyatomic interferences for ICPMS

CANMET WGB-1 Gabbro										
	ICPMS mean	SD	recommended	95% C.I.						
ng/g	n=6									
Re	1.1	0.1	-	-						
Au	1.8	0.3	2.9	1.1						
Pd	12.8	1.5	13.9	2.1						
Pt	5.5	1.2	6.1	1.6						
Rh	0.23	0.05	0.3	-						
Ru	0.20	0.04	0.32	0.21						
Ir	0.22	0.07	0.33	0.17						
USGS W-1 Diabase										
	ICPMS	SD	recommended	SD						
ng/g	n=1									
Re	0.53	0.02	0.49	0.16						
Au	4.27	0.09	4.3	1.2						
Pd	11.56	0.05	14	3						
Pt	13.18	0.02	13	4						
Rh	0.79	0.01	<3							
		0.01	0.0							
Ru	0.11	0.01	0.2	-						
Ru Ir	0.11 0.172	0.004	0.2	0.05						

 Table B.3. ICPMS analyses of standard reference materials.

Appendix C Data accuracy and precision

Appendix Ci -	XRF Standard tables.
Appendix Cii -	XRF operating conditions.
Appendix Ciii -	INAA standard tables.
Appendix Civ -	XRF and INAA standard graphs.

Lower detection limits for XRF data are as follows:

SiO ₂ , Al ₂ O ₃ , Fe ₂ O _{3(t)} , MnO, MgO, CaO and Na ₂ O	0.01 wt%
TiO ₂ , K ₂ O and P ₂ O ₅	0.002 wt%
Nb and Zr	0.6 ppm
Rb	0.5 ppm
Y, Sr, Cr, Zn, Cu and V	1 ppm
Ni , La, Ga, Ba, Th and Nd	2 ppm
Co and Ce	3 ppm
Sc	4 ppm

Appendix Ci: Data accuracy and precision - XRF major element standard tables.

Standard	Run (ARFL)	SiO2	TiO2	AI2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	Total
:NIM-N	278	53.59	0.19	16 35	8 84	0 17	7 58	10.99	2 16	0 229	0.021	100 1
:NIM-N	279	53.36	0.19	16.29	8.85	0.17	7.56	11	2.16	0.23	0.021	99.82
:NIM-N	280	53.4	0.19	16.3	8.84	0.17	7.58	11.08	2.16	0.227	0.021	99.97
:NIM-N	297	52.91	0.19	16.26	8.89	0.17	7.58	11.07	2.15	0.225	0.018	99.47
:NIM-N	317	52.56	0.19	16.15	8.86	0.17	7.55	11.05	2.14	0.228	0.022	98.92
:NIM-N	318	52.47	0.19	16.11	8.86	0.17	7.55	11.07	2.15	0.226	0.02	98.21
:NIM-N	319	52.64	0.19	16.11	8.87	0.171	7.54	11.03	2.13	0.226	0.022	99.98
:NIM-N	320	52.71	0.19	16.16	8.87	0.17	7.56	11.06	2.16	0.228	0.021	99.13
:NIM-N	322	52.63	0.19	16.18	8.9	0.17	7.57	11.1	2.14	0.228	0.02	99.13
Mean		52.92	0.19	16.21	8.86	0.17	7.56	11.05	2.15	0.23	0.02	
Standard o	leviation	0.42	0	0.09	0.02	0	0.02	0.04	0.01	0	0	
Recomme	nded	52.6	0.2	16.5	8.91	0.18	7.5	11.5	2.46	0.25	0.03	100
		-				••••		• • • •				
:MRG-1	317	39.89	3.79	8.34	17.8	0.165	14.04	14.55	0.55	0.174	0.069	99.37
:MRG-1	318	39.78	3.78	8.33	17.8	0.165	14.05	14.55	0.54	0.174	0.066	100.5
:MRG-1	319	40.07	3.78	8.35	17.8	0.165	14	14.51	0.55	0.177	0.069	99.89
:MRG-1	319a	40.17	3.8	8.42	17.84	0.166	14.14	14.53	0.56	0.179	0.068	97.88
:MRG-1	320	40.48	3.79	8.44	17.83	0.166	14.19	14.56	0.57	0.177	0.068	100.3
:MRG-1	322	40.13	3.78	8.47	17.69	0.164	14.16	14.51	0.55	0.176	0.067	99.7
:MRG-1	323	40.1	3.8	8.4	17.81	0.163	14.08	14.54	0.55	0.176	0.067	99.69
:MRG-1	340	40.16	3.78	8.36	17.75	0.165	14.1	14.52	0.56	0.176	0.067	99.64
:MRG-1	370	40.57	3.83	8.47	17.19	0.166	14.29	14.63	0.61	0.18	0.069	99.99
:MRG-1	371	40.47	3.84	8.44	17.23	0.165	14.29	14.63	0.61	0.179	0.069	99.93
:MRG-1	381	40.01	3.8	8.3	17.13	0.153	14.11	14.57	0.7	0.186	0.07	99.03
Mean		40.17	3.80	8.39	17.62	0.16	14.13	14.55	0.58	0.18	0.07	
Standard o	deviation	0.25	0.02	0.06	0.29	0	0.1	0.04	0.05	0	0	
Recomme	nded	<i>39.1</i>	3.77	8.47	17.9	0.17	13.6	14.7	0.74	0.18	0.08	98.7
· ID_1	278	75 12	0 12	12 31	0 9	0.1	0.05	0.67	3 79	4 4 1 6	0 025	97 49
.JR-1	270	75.12	0.12	12.01	0.9	0.1	0.00	0.66	3 79	4 421	0.023	97 93
	219	74.96	0.11	12.30	0.03	0.1	0.04	0.00	3 77	4.421	0.020	97.35
	200	75 16	0.11	12.04	0.03	0.1	0.00	0.67	3 78	4.400	0.020	07.20
Standard	doviation	10.10	0.11	0.04	0.03	0.10	0.00	0.07	0.01	0.01	0.02	
Boommo	ndod	75 41	0.01	12 80	0.01	0 10	0.01 n an	0.67	<i>4</i> 10	4 41	0 02	90 52
necomme	nueu	75.41	0.10	12.03	0.90	0.10	0.50	0.00	4.70	7.71	0.02	00.0Z
:AC-E	278	72.06	0.11	14.66	2.54	0.05	0	0.33	6.17	4.602	0.017	100.5
:AC-E	279	72.35	0.11	14.71	2.54	0.05	0	0.33	6.11	4.615	0.018	100.8
:AC-E	278	72.06	0.11	14.66	2.54	0.05	0	0.33	6.17	4.602	0.017	100.5
:AC-E	280	72.11	0.11	14.78	2.55	0.05	0	0.34	6.13	4.647	0.017	100.7
:AC-E	282	71.7	0.11	14.64	2.54	0.05	0	0.34	6.13	4.632	0.017	100.1
:AC-E	297	69.73	0.11	14.38	2.79	0.06	0	0.33	6.07	4.604	0.017	98.02
Mean		71.67	0.11	14.64	2.58	0.05	0.00	0.33	6.13	4.62	0.02	
Standard	deviation	0.97	0	0.14	0.1	0	0	0.01	0.04	0.02	0	
Recomme	ended	70.4	0.11	14.7	2.53	0.06	0.03	0.34	6.54	4.49	0.01	99.2
:JA-3	279	63.27	0.7	15.39	6.47	0.1	3.7	6.04	2.91	1.352	0.121	100.1
:JA-3	278	63.02	0.7	15.32	6.46	0.1	3.71	6.01	2.92	1.343	0.12	99.69
:JA-3	280	63.12	0.7	15.5	6.49	0.1	3.7	6.07	2.91	1.359	0.117	100.1
:JA-3	282	63.01	0.7	15.42	6.48	0.1	3.69	6.08	2.93	1.351	0.12	99.88
:JA-3	317	62.1	0.7	15.24	6.49	0.099	3.68	6.06	2.89	1.346	0.119	98.72
:JA-3	318	61.85	0.7	15.25	6.5	0.101	3.69	6.06	2.89	1.351	0.118	99.6
:JA-3	319a	62.54	0.7	15.33	6.54	0.099	3.72	6.11	2.9	1.361	0.121	100
:JA-3	320	63.13	0.7	15.35	6.53	0.1	3.73	6.08	2.91	1.359	0.122	100
:JA-3	322	61.83	0.69	15.32	6.4	0.095	3.73	6.09	2.89	1.35	0.121	98.52
·JA-3	323	62.32	0.71	15.27	6.51	0.099	3.7	6.08	2.89	1.353	0.121	100.1
.JA-3	325	62 19	0.7	15.19	6.5	0.096	3.71	6.03	2.92	1.341	0.122	98.81
Mean	520	62.58	0.70	15.33	6.49	0.10	3.71	6.06	2.91	1.35	0.12	
Standard	deviation	0.55	0	0.09	0.04	0	0.02	0.03	0.01	0.01	0	
Recomme	ended	62.3	0.68	15.6	6.59	0.11	3.65	6.28	3.17	1.41	0.11	99.8
											-	

Appendix Ci: Data accuracy and precision - XRF major element standard tables.

Standard	Run (ARFL)	SiO2	TiO2	AI2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	Total
:SO-2	279	61.88	1.67	17	8.96	0.1	0.97	2 96	2 58	3 377	0 843	100.3
:SO-2	278	61.51	1.66	16.91	8.93	0.1	0.99	2.94	2.58	3 362	0.839	99 82
:SO-2	280	61.62	1.67	17.04	8.95	0.1	0.99	2.98	2 59	3 373	0.849	100.2
:SO-2	282	61.63	1.66	17	8.96	0.1	0.98	2.98	2.58	3 375	0.851	100.1
:SO-2	297	60.99	1.67	16.94	9.02	0.1	0.99	2.98	2.58	3.394	0.842	99.51
Mean		61.53	1.67	16.98	8.96	0.10	0.98	2.97	2.58	3.38	0.84	
Standard d	eviation	0.33	0.01	0.05	0.03	0	0.01	0.02	0	0.01	0.01	
Recommer	nded	53.4	1.43	15.1	7.89	0.09	0.89	2.77	2.48	2.94	0.69	87.7
Recalculate	ed* (11.7	60.5	1.62	17.1	8.94	0.1	1.01	3.14	2.81	3.33	0.78	99.4
:UB-69	279	58.38	0.78	15.92	9.53	0.22	3.16	6.54	2.18	2.237	0.268	99.22
:UB-69	278	58.06	0.77	15.83	9.5	0.22	3.16	6.53	2.2	2.227	0.264	98.76
:UB-69	280	58.24	0.78	15.97	9.54	0.22	3.17	6.6	2.2	2.245	0.272	99.24
:UB-69	282	58.06	0.77	15.91	9.52	0.22	3.18	6.58	2.19	2.235	0.271	98.95
:UB-69	317	57.33	0.77	15.84	9.56	0.224	3.17	6.58	2.18	2.244	0.266	98.16
:UB-69	318	57.21	0.77	15.83	9.57	0.224	3.18	6.59	2.2	2.244	0.269	99.81
:UB-69	319a	57.6	0.78	15.89	9.6	0.223	3.18	6.61	2.22	2.246	0.268	99.94
:UB-69	322	57.08	0.77	15.76	9.4	0.215	3.19	6.62	2.22	2.226	0.27	97.75
Mean		57.75	0.77	15.87	9.53	0.22	3.17	6.58	2.20	2.24	0.27	
Standard d	leviation	0.5	0.01	0.07	0.06	0	0.01	0.03	0.02	0.01	0	
:PCC-1	317	44.57	0.01	0.6	8.67	0.122	45.72	0.5	0	0	0.012	100.1
:PCC-1	318	44.41	0.01	0.59	8.67	0.12	45.74	0.5	0	0	0.01	99.27
:PCC-1	319a	44.73	0.01	0.6	8.69	0.12	45.85	0.51	0	0	0.011	100.2
:PCC-1	320	45.4	0.01	0.6	8.7	0.12	46.13	0.51	0	0	0.012	101.3
:PCC-1	322	44.56	0.01	0.6	8.57	0.117	45.91	0.51	0	0	0.011	100.1
Mean		44.73	0.01	0.60	8.66	0.12	45.87	0.51	0.00	0.00	0.01	
Standard d	leviation	0.39	0	0	0.05	0	0.17	0.01	0	0	0	
Recommer	nded	41.7	0.01	0.68	8.25	0.12	43.4	0.52	0.03	0.01	0	94.8
Recalculate	ed* (5.12	44	0.01	0.71	8.7	0.13	45.8	0.55	0.03	0.01	0	99.9
:NIM-G	340	76.91	0.1	11.96	1.99	0.012	0	0.74	3.18	4.865	0.017	99.76
:NIM-G	370	76.42	0.11	11.89	1.92	0.013	0.03	0.75	3.19	4.888	0.02	99.23
:NIM-G	371	76.28	0.11	11.85	1.91	0.012	0.03	0.75	3.21	4.901	0.02	99.07
Mean		76.54	0.11	11.90	1.94	0.01	0.01	0.75	3.19	4.88	0.02	
Standard d	leviation	0.33	0.01	0.06	0.04	0	0.03	0.01	0.02	0.02	0	
Recommer	nded	75.7	0.09	12.1	2.02	0.02	0.06	0.78	3.36	4.99	0.01	99.1
:JB-1a	370	53.31	1.31	14.22	8.55	0.139	8	9.03	2.45	1.353	0.277	98.63
:JB-1a	371	53.29	1.31	14.21	8.56	0.14	7.98	9.03	2.47	1.361	0.276	98.63
Mean		53.30	1.31	14.22	8.56	0.14	7.99	9.03	2.46	1.36	0.28	
Standard d	leviation	0.01	0	0.01	0.01	0	0.01	0	0.01	0.01	0	
Recommer	nded	52.2	1.3	14.5	9.1	0.15	7.75	9.23	2.74	1.42	0.26	98.6
:JG-3	370	67.74	0.5	14.99	3.53	0.068	1.74	3.62	3.66	2.55	0.142	98.53
:JG-3	371	67.53	0.5	14.94	3.53	0.067	1.74	3.62	3.66	2.546	0.14	98.26
Mean		67.64	0.50	14.97	3.53	0.07	1.74	3.62	3.66	2.55	0.14	
Standard d	leviation	0.15	0	0.04	0	0	0	0	0	0	0	
Recommer	nded	67.1	3.73	15.5	3.73	0.07	1.79	3.76	4.03	2.63	0.12	102
:JA-2	317	58.19	0.71	15.85	6.29	0.1	7.46	6.09	2.74	1.801	0.169	99.41
:JA-2	318	58.09	0.71	15.86	6.29	0.103	7.46	6.12	2.74	1.802	0.165	99.38
:JA-2	319a	58.53	0.71	15.88	6.3	0.104	7.48	6.12	2.76	1.805	0.17	100.2
:JA-2	320	59.24	0.72	15.97	6.32	0.102	7.52	6.13	2.76	1.805	0.166	100.7
:JA-2	322	58.18	0.7	15.93	6.2	0.099	7.54	6.12	2.74	1.801	0.153	99.46
:JA-2	325	58.4	0.71	15.78	6.29	0.102	7.47	6.09	2.76	1.791	0.166	99.56
Mean		58.44	0.71	15.88	6.28	0.10	7.49	6.11	2.75	1.80	0.16	
Standard o	leviation	0.42	0.01	0.07	0.04	0	0.03	0.02	0.01	0.01	0.01	
Recomme	nded	56.2	0.67	15.3	6.14	0.11	7.68	6.48	3.08	1.8	0.15	97.6
Recalculat	ed* (2.07	57.4	0.68	15.6	6.27	0.11	7.84	6.62	3.15	1.84	0.15	99.7

Appendix C.3

Appendix Ci: Data accuracy	and precision - XRI	F major element	standard tables.

Standard	Run (ARFL)	SiO2	TiO2	AI2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	Total
:NIM-D	370	40.04	0.02	0.21	16.18	0 2 1 9	44 57	0.26	0	0.009	0.014	101 4
:NIM-D	371	40.06	0.02	0.21	16.10	0.22	44.6	0.26	õ	0.008	0.014	101.5
Mean	011	40.05	0.02	0.21	16.20	0.22	44 59	0.20	0 00	0.000	0.014	101.0
Standard d	eviation	0.01	0	0	0.02	0.22	0.02	0.20	0.00	0.07	0.07	
Recommer	nded	39	0.02	0.3	17	0.22	43.5	0.28	0.04	0.01	0.01	100
:BHVO-1	370	50.48	2.77	13.39	11.77	0.166	7.4	11.11	2.06	0.512	0.279	99.94
:BHVO-1	371	50.62	2.77	13.35	11.77	0.162	7.38	11.1	2.06	0.509	0.276	100
Mean		50.55	2.77	13.37	11.77	0.16	7.39	11.11	2.06	0.51	0.28	
Standard d	leviation	0.1	0	0.03	0	0	0.01	0.01	0	0	0	
Recommer	nded	49.9	2.71	13.8	12.2	0.17	7.23	11.4	2.26	0.52	0.27	101
:BE-N	381	40.14	2.66	10.1	12.65	0.193	13.98	14.06	2.86	1.331	1.1	99.08
Recommer	nded	38.2	2.61	10.1	12.8	0.2	13.2	13.9	3.18	1.39	1.05	96.6
Recalculat	ed* (2.4£	<i>39.2</i>	2.68	10.3	13.2	0.21	13.5	14.2	3.26	1.43	1.08	<i>99</i>
:JB-2	323	53.18	1.17	14.33	13.94	0.203	4.62	9.51	1.83	0.401	0.101	99.3
Recomme	nded	53.2	1.19	14.7	14.3	0.2	4.66	9.89	2.03	0.42	0.1	101

NB: Recalculated* totals for BE-N, SO-2, JA-2 and PCC-1 are to a volatile free basis. (L.O.I. values shown in brackets).
Appendix Ci: Data accuracy and precision - XRF trace element (RHT) standard tables.

Standard	Run (RHT)	Nb	Zr	Y	Sr	Rb	Th	Ga	Zn	Ni
:GH	501	91.2	155.0	80.8	9.4	382.5	71.9	22.8	55.2	4.2
:GH	504	90.5	153.6	78.2	9.8	384.5	72.7	22.9	55.3	4.0
:GH	505	90.8	155.1	79.3	9.8	384.1	72.2	22.4	55.7	3.6
:GH	509	91.1	156.3	80.6	9.2	386.9	74.3	21.7	56.0	3.9
:GH	510	89.6	154.5	79.8	9.5	384.3	72.6	22.4	55.3	4.0
:GH	523	90.8	154.2	81.1	9.7	384.9	72.3	22.2	55.3	5.1
:GH	541	92.2	159.9	80.2	9.7	391.0	73.4	23.1	52.9	6.4
:GH	547	92.3	157.0	81.6	10.4	388.9	76.0	22.4	52.8	6.0
:GH	550	93.3	156.6	81.2	10.2	390.4	76.1	22.7	53.1	6.2
Mean		91.3	1 <i>55.8</i>	80.3	9.7	386.4	73.5	22.5	54.6	4.8
Standard deviation		1.1	1.9	1.1	0.4	3.1	1.6	0.4	1.3	1.1
Recommended		90.0	155.0	75.0	10.0	390.0	87.0	23.0	55.0	3.0
:SO-1	501	11.5	88.1	24.2	318.0	125.7	10.8	25.2	124.9	86.8
:SO-1	550	11.7	87.7	25.1	321.6	126.4	12.1	24.7	116.8	86.0
Mean		11.6	87.9	24.7	319.8	126.1	11.5	25.0	120.9	86.4
Standard deviation		0.1	0.3	0.6	2.5	0.5	0.9	0.4	5.7	0.6
Recommended		11.7	84.0	24.5	331.0	141.0	12.4	24.1	144.0	92 .0
:JA-1	501	2.1	84.5	31.0	249.3	11.7	0.5	18.0	87.9	1.4
:JA-1	502	1.7	83.5	30.0	248.6	12.1	0.7	18.4	88.2	0.6
:JA-1	509	1.0	85.0	31.1	254.0	12.0	0.2	18.2	87.8	0.0
:JA-1	523	1.8	84.8	30.1	254.3	12.1	0.9	17.9	86. 9	0.9
:JA-1	541	2.5	85.1	31.1	256.2	11.1	0.0	18.1	82.8	2.1
:JA-1	547	1.7	83.2	30.5	256.0	11.4	0.6	17.7	79.9	2.7
:JA-1	550	0.9	84.4	30.7	254.9	11.5	0.4	17.5	82.7	2.8
Mean		1.7	84.4	30.6	253.3	11.7	0.4	18.0	<i>85.2</i>	1.5
Standard deviation		0.6	0.7	0.5	3.1	0.4	0.6	0.3	3.3	1.1
Recommended		1.7	88.3	30.6	266.0	11.8	0.8	17.3	90.6	1.8
:BE-N	501	108.3	267.0	30.4	1368.0	49.4	11.8	17.9	109.0	274.5
:BE-N	502	109.3	266.8	29.9	1373.7	48.7	9.3	18.5	111.3	271.6
:BE-N	509	108.2	266.0	30.7	1374.3	49.1	13.5	18.0	112.2	276.0
:BE-N	510	108.9	268.8	29.9	1375.0	48.4	11.8	18.8	112.9	277.4
:BE-N	523	108.6	269.2	29.4	1371.0	50.1	10.6	18.9	112.9	273.6
:BE-N	541	107.0	265.7	31.2	1328.2	47.1	10.7	17.8	105.2	265.1
:BE-N	544	106.0	265.9	30.1	1331.2	46.3	9.4	17.7	106.9	268.9
:BE-N	547	107.2	264.9	29.4	1331.6	46.6	9.9	17.4	103.7	265.4
:BE-N	550	106.9	266.2	30.4	1324.8	46.3	9.4	17.2	104.1	270.3
Mean		107.8	266.7	30.2	1353.1	48 .0	10.7	18.0	108.7	271.4
Standard deviation		1.1	1.4	0.6	23.1	1.4	1.4	0.6	3.8	4.4
Recommended		110.0	270.0	30.0	1370.0	47.0	10.4	17.0	120.0	267.0
:AN-G	501	1.3	10.2	8.6	75.5	2.6	0.0	20.0	24.1	31.3
:AN-G	504	0. 9	8.8	8.4	74.0	2.3	0.0	19.1	23.8	31.5
:AN-G	505	1.4	9.5	8.1	74.4	2.7	0.2	19.1	24.8	31.8
:AN-G	509	1.4	8.8	8.6	74.6	1.3	0.6	19.0	23.3	31.2
:AN-G	510	1.3	9.9	8.0	74.6	2.8	0.3	19.3	24.6	31.4
:AN-G	523	1.2	9.5	8.2	75.3	2.2	0.9	19.3	24.5	33.1
Mean		1.3	9.5	8.3	74.7	2.3	0.1	19.3	24.2	31.7
Standard deviation		0.2	0.6	0.3	0.6	0.5	0.7	0.4	0.6	0.7
Recommended		0.7	11.0	7.5	76.0	1.0	0.0	18.0	20.0	35.0
:JP-1	501	0.6	6.4	0.1	0.7	1.2	0.0	0.4	41.3	2411.5
:JP-1	504	0.3	6.4	0.2	1.0	1.7	0.0	1.6	41.9	2408.3
:JP-1	505	0.2	6.5	0.4	0.9	1.2	0.0	1.3	43.0	2402.8
:JP-1	509	0.3	6.8	0.5	0.6	1.4	0.1	1.3	42.1	2493.2
:JP-1	510	0.4	6.2	0.7	1.3	1.5	0.0	1.5	41.5	2483.8
:JP-1	523	0.6	6.4	0.1	1.2	2.1	1.1	0.9	42.6	2355.8

Appendix Ci: Data accuracy and precision - XRF trace element (RHT) standard tables.

Standard	Run (RHT)	Nb	Zr	Y	Sr	Rb	Th	Ga	Zn	Ni
:JP-1	541	0.8	5.0	0.8	1.7	1.2	0.2	0.7	40.5	2372.3
:JP-1	547	0.0	5.4	0.7	1.8	1.3	0.0	0.7	39.3	2333.0
:JP-1	550	0.3	5.7	0.2	2.0	12	0.7	0.0	39.6	2322.4
Mean		0.4	6.1	0.4	1.2	14	0.0	0.9	41.3	2398.1
Standard deviation		02	0.6	0.3	0.5	0.3	0.8	0.6	1.3	60.3
Recommended		1.2	6.3	0.0	0.0	0.0	0.2	0.5	29.5	2460.0
:JA-3	501	6.5	128.5	22.2	292.8	38.8	3.0	19.0	65.6	27.7
:JA-3	510	1.8	84.8	31.0	251.8	11.7	1.0	18.1	86.8	1.7
Mean		4.2	106.7	26.6	272.3	25.3	2.0	18.6	76.2	14.7
Standard deviation		3.3	30.9	6.2	29.0	19.2	1.4	0.6	15.0	18.4
Standard		3.0	119.0	21.3	<i>294.0</i>	36.0	3.1	17.0	67.5	35.5
:ZW-C	501	194.9	80.8	11.0	18.4	8548.4	38.1	103.7	957.0	21.6
Recommended		198.0	82.0	33.0	17.0	8500.0	43.0	99.0	1050.0	11.0
:SO-2	501	22.4	784.1	43.5	333.7	71.7	3.7	26.5	107.8	2.7
:SO-2	550	22.5	786.6	43.2	340.2	72.1	5.8	27.4	99.4	2.2
Mean		22.5	785.4	43.4	337.0	71.9	4.8	27.0	103.6	2.5
Standard deviation		0.1	1.8	0.2	4.6	0.3	1.5	0.6	5.9	0.4
Recommended		22.0	760.0	40.0	331.0	77.0	3.8	24.3	115.0	8.0
	504		100.0	10.4	007.0	05.4		7.0	44.0	
:SU-3	501	6.2	160.2	10.4	207.0	35.1	2.3	7.0 6.4	44.0	0.0 14.0
Recommended		0.4	150.0	10.4	222.0	39.0	3.9	0.4	40.3	14.0
:SO-4	501	10.0	288.3	23.1	152.9	64.1	6.6	11.3	80.9	18.0
Recommended		10.0	270.0	22.0	168.0	69.0	8.6	10.7	94.0	24.0
:BHVO-1	501	18.3	182.2	27.9	385.3	10.2	1.6	23.0	92.7	115.5
:BHVO-1	504	17.9	180.8	27.0	383.7	10.1	1.3	23.1	92.7	115.5
:BHVO-1	505	18.3	180.2	28.2	384.5	10.1	0.6	22.1	92.7	115.2
:BHVO-1	550	18.3	178.6	28.5	387.8	8.5	0.1	22.2	83.1	110.9
Mean		18.2	180.5	27.9	385.3	9.7	0.9	22.6	90.3	114.3
Standard deviation		0.2	1.5	0.6	1.8	0.8	0.7	0.5	4.8	2.3
Recommended		19.8	182.0	27.6	403.0	11.0	1.1	21.0	105.0	121.0
	501	50	107.0	28.2	102 1	67	0 9	16 1	67.2	107.6
	501	5.Z 4 G	107.9	20.3	102.1	71	0.5	15.7	66.7	107.0
BOB-1	502	4.0	105.4	27.7	101 5	6.6	0.5	15.7	66.6	109.1
BOB-1	504	4.0	106.1	27.0	191.5	6.0	0.0	10.0	67.6	100.7
BOB-1	505	4.0	106.0	20.2	100.0	0.5	0.0	15.6	67.5	109.0
BOB-1	509	4.0	100.2	29.0	192.0	5.0	0.0	10.0	67.0	109.0
BOB-1	510	4.0	107.7	20.0	102.1	6.0	0.0	10.2	67.0	109.7
BOB-1	523	4.7	100.0	20.0	193.1	0.1	1.0	15.0	65.0	110.4
:BOB-1	541	5.1	107.3	29.0	194.9	0.1	1.2	15.0	00.2 64.5	100.5
:BOB-1	547	4.8	104.7	28.1	193.4	0.5 5 7	0.0	15.4	04.5	110.0
:BOB-1	550	4.9	106.6	30.2	194.7	5.7	0.0	10.3	05.4 66 5	100.0
Mean		4.9	106.4	28.6	192.7	0.3	0.0	15.8	00.5	109.3
Standard deviation		0.2	1.0	0.7	1.3	0.5	0.7	0.3	1.1	1.0
RH average*		4.8	104.9	28.3	196.7	5.5	0.5	15.1	65.4	110.9
:SY-2	504	31.7	306.7	138.2	269.2	232.4	382.1	29.6	229.1	6.8
:SY-2	505	31.6	307.6	136.9	269.8	231.6	383.5	30.1	228.0	7.8
:SY-2B	504	31.7	303.9	137.3	267.8	230.8	376.2	30.0	222.0	6.7
:SY-2B	505	32.8	303.2	137.4	269.5	228. 9	377.9	28.8	222.3	7.2
Mean		32.0	305.4	137.5	269.1	230.9	379.9	29.6	225.4	7.1
Standard deviation		0. 6	2.1	0.5	0.9	1.5	3.4	0.6	3.7	0.5
Recommended		<i>29</i> .0	280.0	128.0	271.0	217.0	379.0	29.0	248.0	9.9
·W-1	501	7.7	98.0	22.9	186.7	22.9	1.4	19.3	80.2	71.7
:W-1	502	7.3	97.1	22.5	185.4	23.2	2.4	19.5	80.4	73.6

Appendix Ci: Data accuracy and precision - XRF trace element (RHT) standard tables.

Standard	Run (RHT)	Nb	Zr	Y	Sr	Rb	Th	Ga	Zn	Ni
:W-1	504	7.0	97.1	22.7	186.0	23.1	2.7	19.5	81.0	73.6
:W-1	505	7.1	96.2	22.9	184.8	22.2	0.7	19.0	81.0	71.6
:W-1	509	6.5	97.9	23.4	187.7	22.2	3.5	18.1	81.6	73.3
:W-1	510	7.1	97.3	23.7	186.1	22.7	0.2	18.5	80.0	72.7
:W-1	523	6.8	97.0	23.5	186.2	22.5	2.8	18.6	82.4	71.2
:W-1	541	7.5	97.6	23.1	187.7	22.4	0.6	18.7	75.6	72.6
:W-1	547	6.8	95.6	22.9	188.8	21.7	1.8	18.9	75.6	70.5
:W-1	550	6.7	94.9	22.5	188.6	22.0	1.5	19.0	75.0	72.6
Mean		7.1	96.9	23.0	186.8	22.5	1.8	18.9	79.3	72.3
Standard deviation		0.4	1.0	0.4	1.3	0.5	1.1	0.5	2.8	1.1
Recommended		7.7	98.6	26.0	186.0	21.4	2.4	17.4	84.0	75.0
:NIM-G	501	54.9	304.7	142.4	10.3	326.2	52.1	28.5	52.2	7.3
:NIM-G	504	54.6	305.1	140.6	11.3	332.1	52.5	29.4	52.4	6. 9
:NIM-G	505	54.8	303.8	139.6	11.0	328.0	54.2	29.4	52.3	6.8
:NIM-G	509	54.8	305.6	142.5	11.1	326.9	53.9	28.4	51.6	7.6
:NIM-G	510	55.0	305.8	139.9	11.6	330.8	52.6	29.2	52.5	7.8
:NIM-G	523	55.1	301.6	142.6	11.3	326.7	52.4	27.9	51.7	8.2
:NIM-G	541	54.7	301.0	139.3	13.5	321.0	51.5	28.6	44.8	5.2
:NIM-G	547	53.7	294.4	138.1	12.7	318.3	51.1	28.3	44.1	5.5
:NIM-G	550	53. 9	297.8	139.2	13.1	320.4	51.9	28.9	44.5	4.5
Mean		54.6	302.2	140.5	11.8	325.6	52.5	28.7	49.6	6.6
Standard deviation		0.5	3.9	1.7	1.1	4.7	1.0	0.5	3.8	1.3
Recommended		55.0	300.0	143.0	10.0	320.0	51.0	27.0	50.0	8.0
:GA	501	11.9	135.9	20.7	292.4	173.1	15.7	16.0	66.4	5.4
:GA	504	11.8	134.5	19.7	291.6	174.6	15.9	16.8	67.2	5.2
:GA	505	12.4	134.1	18.6	289.2	172.3	14.8	16.5	66.5	4.7
:GA	509	12.1	136.0	21.4	293.5	1/2.8	15.5	15.7	67.8	5.2
:GA	510	12.2	135.4	21.2	291.4	170.4	14.6	15.9	66./	5.0
:GA	523	11.9	134.8	19.9	295.0	1/4.1	16.3	15.8	67.8	5.4
:GA	541	12.1	135.7	20.2	296.0	172.0	16.7	15.9	63.1	8.2
:GA	550	12.7	134.9	20.1	297.3	174.8	15.8	15.1	62.2	6.8
Mean		12.1	135.2	20.2	293.3	1/3.0	15.7	16.0	66.0	5.7
Standard deviation		0.3	0.7	0.9	2.7	1.5	0.7	0.5	2.1	1.2
	501	12.0	135.0	21.0	310.0	7.6	0.5	16.0 20.3	80.0	7.0
IMRG-1	501	19.5	100.4	14.5	207.0	8.6	0.5	20.5	188 1	102.0
IMRG-1	504	19.2	109.0	14.0	200.9	8.6	10	10.6	185 /	193.3
	505	10.7	109.9	121	207.5	77	2.0	19.0	186 1	10/ 6
MRG-1	509	19.2	109.4	1//	207.3	7.7	0.4	20.2	185.8	105.0
	510	10.9	109.4	13.6	267.9	84	12	10.1	183.5	193.9
	525	20.2	110.2	15.0	270.1	63	1.2	20.2	173.6	187 5
	544	20.2	110.0	14.9	270.5	59	14	20.8	177 1	189 1
-MRG-1	547	20.0	109.1	14.7	272.0	6.2	0.0	19.0	170.8	189.0
MPG-1	550	19.9	109.7	14.7	268.6	5.8	14	19.4	172.8	189.2
.Mind-1 Moan	550	19.5	109.6	14.3	269 1	7.3	0.9	19.8	180.5	192.0
Standard deviation		06	07	0.5	16	1.1	1.0	0.7	6.4	3.0
Recommended		20.0	108.0	14.0	266.0	8.5	0.9	17.0	191.0	193.0
:JR-1	523	15.4	104.9	45.1	27.7	253.7	27.8	16.4	27.1	-0.7
:JR-1	541	15.8	106.4	44.7	28.5	253.6	27.0	16.5	28.3	0.2
:JR-1	547	16.1	104.0	45.0	28.5	253.9	25.8	16.7	27.6	0.6
:JR-1	550	15.3	105.3	45.5	27.7	253.1	28.5	16.6	27.9	0.2
Mean		15.7	105.2	45.1	28.1	253.6	27.3	16.6	27.7	0.1
Standard deviation		0.4	1.0	0.3	0.5	0.3	1.2	0.1	0.5	0.6
Recommended		15.5	101.0	45.4	30.0	257.0	26.5	17.6	30.0	0.7

Appendix Ci: Data accuracy and precision	- XRF trace element	(RHT) standard tables.
		and the second

Standard	Run (RHT)	Nb	Zr	Y	Sr	Rb	Th	Ga	Zn	Ni
:STM-1	509	264.3	1466.3	47.2	718.2	122.1	33.4	37.4	213.2	0.2
:STM-1	510	264.8	1458.6	48.7	712.5	121.6	33.1	38.4	210.0	0.0
:STM-1	523	265.1	1442.7	47.7	712.3	122.3	33.3	37.5	212.6	0.0
:STM-1	541	263.3	1462.0	48.3	718.1	122.4	34.3	37.5	198.1	1.6
:STM-1	547	267.3	1432.5	46.9	714.8	121.9	33.0	37.5	194.4	2.4
:STM-1	550	269.9	1453.3	47.9	718.3	122.7	35.3	37.0	196.4	2.0
Mean		265.8	1452.6	47.8	715.7	122.2	33.7	37.6	204.1	1.0
Standard deviation		2.4	12.8	0.7	2.9	0.4	0.9	0.5	8.7	1.2
Recommended		266.0	1440.0	46.0	700.0	118.0	31.0	36.0	2 3 5.0	3.0
:BCR-1	510	12.5	199.6	37.7	325.5	50.0	6.5	24.3	133.4	13.7
:BCR-1	550	12.3	200.4	38.6	324.2	46.9	6.4	23.8	119.8	11.6
Mean		12.4	200.0	38.2	324.9	48.5	6.5	24.1	126.6	12.7
Standard deviation		0.1	0.6	0.6	0.9	2.2	0.1	0.4	9.6	1.5
Recommended		14.0	190.0	38.0	330.0	47.2	6.0	22.0	129.5	13.0
:AGV-1	510	14.0	243.0	19.9	650.5	68.9	6.8	21.3	77.0	11.0
:AGV-1	550	13.9	242.0	21.1	653.0	68.9	8.2	21.3	74.4	11.2
Mean		14.0	242.5	20.5	651.8	68.9	7.5	21.3	75.7	11.1
Standard deviation		0.1	0.7	0.8	1.8	0.0	1.0	0.0	1.8	0.1
Recommended		15.0	227.0	20.0	662.0	67.3	6.5	20.0	88.0	16.0
:AC-E	510	109.4	879.3	187.4	2.9	150.1	19.2	42.7	181.9	0.4
Recommended		110.0	7 8 0.0	184.0	3.0	152.0	18.5	39.0	224.0	1.5
:DNC-1	550	1.7	35.4	19.3	143.0	4.3	0.0	15.2	53.3	234.5
Recommended		3.0	41.0	18.0	145.0	4.5	0.2	15.0	6 6 .0	247.0
:JA-2	550	8.7	123.0	17.9	243.7	73.4	5.3	17.7	55.1	120.6
Recommended		9.8	119.0	18.1	252.0	68.0	4.7	16.4	62.7	142.0
:G2	550	11.5	322.5	10.0	467.5	170.1	25.6	22.5	70.0	0.4
Recommended		12.0	30 9 .0	11.0	478.0	170.0	24.7	23.0	86.0	5.0
:BIR-1	550	0.4	12.3	17.0	109.3	0.4	0.0	17.6	56.1	151.7
Recommended		0. 6	15.5	16.0	108.0	0.3	0.0	16.0	71.0	166.0
:QLO-1	550	10.2	200.6	25.0	327.0	73.2	6.2	18.5	56.1	0.2
Recommended		10.3	185.0	24.0	336.0	74.0	4.5	17.0	61.0	5.8
:RGM-1	550	8.9	248.9	23.9	102.3	151.9	15.6	15.6	28.5	0.0
Recommended		8.9	219.0	25.0	108.0	149.0	15.1	15.0	32.0	4.4
:BR	550	110.0	270.5	30.8	1343.8	49.2	11.9	18.6	133.8	268.1
Recommended		98.0	260.0	30.0	1320.0	47.0	11.0	19.0	160.0	260.0

NB: RH average* for BOB-1 is taken from an average of 24 runs at Royal Holloway, London.

Appendix Ci: Data accuracy and precision - XRF trace element (Tranmet) standard tables.

Standard	Run (T)	Sc	V	Cr	Co	Cu	Ba	La	Ce	Nd
:AC-E	202	2.9	9.3	0.0	28.8	1.4	51.3	61.5	162.0	96.2
:AC-E	201	2.8	12.7	0.0	30.2	1.5	52.6	61.1	163.4	94.9
:AC-E	200	-0.9	4.6	0.6	29.8	0.9	52.2	60.7	168.6	92.3
:AC-E	198	0.2	12.8	0.4	30.5	1.1	49.5	62.9	171.5	95.4
:AC-E	197	0.5	6.4	0.0	31.8	1.6	52.4	61.1	169.9	95.5
Mean		1.1	9.2	0.2	30.2	1.3	51.6	61.5	167.1	94.9
Standard (deviation	1.7	3.7	0.3	1.1	0.3	1.3	0.9	4.2	1.5
Recomme	nded	0.1	3.0	3.4	0.2	4.0	55.0	180.0	154.0	92.0
:BR	202	24.6	245.8	375.5	59.5	77.2	1199.3	84.5	145.2	64.7
:BR	201	26.5	246.5	376.5	56.4	79.8	1211.1	84.0	150.4	66.5
:BR	200	19.1	239.7	369.0	60.8	78.0	1198.0	84.0	154.7	62.0
:BR	198	27.9	250.0	373.9	56.9	79.7	1199.6	85.3	151.2	62.8
:BR	197	24.4	243.3	371.4	60.5	77.2	1202.7	81.4	155.4	63.6
:BR	196	20.9	241.1	364.7	56.2	76.3	1180.6	83.5	150.5	60.6
:BR	190	22.4	251.2	376.3	55.1	76.4	1195.1	83.0	153.1	59.1
·BR	189	22.8	252.0	371.0	57.0	78.7	1201.6	84.8	160.0	67.0
Mean	100	23.6	246.2	372.3	57.8	77.9	1198 5	83.8	152.6	63.3
Standard	doviation	20.0	16	A 1	22	1 1	86	12	102.0 A A	27
Recomme	ended	2.1	235.0	380.0	52.0	72.0	1050.0	82.0	151.0	65.0
:MRG-1	202	57.2	571.4	535.4	88.1	130.2	51.2	9.9	31.9	18.2
:MRG-1	201	59.7	573.3	541.2	85.2	132.1	54.2	8.2	32.7	20.5
:MRG-1	200	46.9	577.3	530.5	89.7	132.7	54.8	11.5	32.6	18.9
·MRG-1	198	55.6	578.3	530.5	85.2	127.6	57.9	9.7	32.2	22.5
MRG-1	197	52.4	571 7	538.4	85.5	128.1	55.9	7.5	31.9	20.6
MRG-1	196	53 1	572.4	537.3	88.2	130.8	50.7	64	25.4	23.1
MRG-1	190	18 9	580 3	534 1	84.6	130.4	51 7	6.5	33.4	18.6
	70		506.0	517 5	01.0	100.4	61.6	0.0	22 A	10.0
	12	53. 4	530.3	522.1	91.5	120 2	54.8	10.2	30.0	20.3
Standard	doviation	10	011.1	72	25	1 0	37	51	10	10
Decommo	uevialion	4.2 55 0	0.4 526 0	1.0	2.5	124.0	61.0	0.1	7.3 26 0	10.2
necomme	anueu	55.0	520.0	430.0	07.0	134.0	07.0	3.0	20.0	13.2
:NIM-G	189	3.1	4.5	28.3	0.2	3.8	105.9	111.8	204.6	73.3
:NIM-G	201	0.0	12.7	26.0	1.8	5.0	109.2	108.2	201.2	71.0
:NIM-G	200	0.7	6.7	25.7	0.0	4.9	107.2	109.7	201.7	72.2
:NIM-G	202	3.0	11.5	29.1	1.1	4.0	106.9	110.3	205.8	74.5
:NIM-G	198	2.1	10.8	25.3	2.8	4.8	108.5	108.9	212.6	75.0
:NIM-G	197	2.5	10.7	29.3	0.0	4.7	107.2	110.3	211.7	71.3
:NIM-G	196	1.7	7.2	26.9	2.7	4.8	107.5	108.3	207.1	74.7
:NIM-G	190	1.7	14.5	22.8	1.2	4.6	107.2	113.8	208.6	70.8
:NIM-G	72	0.0	2.3	25.4	5.3		108.4	101.9	194.4	72.6
Mean		1.6	9 .0	26.5	1.7	4.6	107.6	109.2	205.3	72.8
Standard	deviation	1.2	4.0	2.1	1.7	0.4	1.0	3.3	5.7	1.6
Recomme	ended	1.0	2.0	12.0	4.0	12.0	120.0	109.0	195.0	48.0
:W-2	189	28.0	258.8	93.3	43.7	102.2	184.7	10.4	24.4	12.0
:W-2	201	36.7	249.9	90.2	44.7	102.0	185.6	14.1	20.0	11.0
:W-2	200	28.6	247.6	93.0	47.3	103.1	184.5	11.8	21.3	11.5
:W-2	202	37.1	252.6	95.9	45.7	100.0	182.3	14.2	22.7	11.6
:W-2	198	34.2	253.3	91.6	45.0	100.4	183.7	16.5	21.6	13.4
:W-2	197	31.7	252.6	92.8	46.5	101.1	185.7	15.6	23.1	12.8
:W-2	196	33.7	257.9	94.1	47.1	102.1	185.3	14.3	24.1	11.7
·W-2	190	29.9	255.2	86.8	42.3	101.8	185.9	13.3	23.5	10.5
Mean		32.5	253.5	92.2	45.3	101.6	184.7	13.8	22.6	11.8
Standard	deviation	35	3.8	2.8	1.7	1.0	1.2	2.0	1.5	0.9
Recomme	anded	35 0	262.0	93.0	44.0	103.0	182.0	11.4	24.0	14.0
100000000		20.0				•		-		'

Appendix Ci: Data accuracy and precision - XRF trace element (Tranmet) standard tables.

Standard	Run (T)	Sc	V	Cr	Со	Cu	Ва	La	Се	Nd
:JP-1	202	5.4	33.1	2716.3	88.5	10.6	9.0	4.9	0.0	0.0
:JP-1	201	10.0	34.8	2766.1	87.5	10.1	10.8	5.1	0.0	0.0
:JP-1	200	3.6	30.3	2740.7	89.7	11.1	11.1	2.8	0.0	0.0
:JP-1	198	4.2	28.2	2777.3	90.8	10.3	12.3	4.6	0.2	0.0
:JP-1	197	5.6	35.4	2765.9	88.2	9.2	11.8	3.1	0.0	0.0
:JP-1	196	7.2	28.5	2783.3	90.7	11.1	10.0	3.2	0.2	0.0
:JP-1	72	3.7	23.1	2609.0	78.0		11.2	2.3	0.0	0.0
Mean		5.7	30.5	2736.9	87.6	10.4	10.9	3.7	0.1	0.0
Standard	deviation	2.3	4.3	60.9	4.4	0.7	1.1	1.1	0.1	0.0
:JG-3	189	8.7	70.4	24.5	13.0	4.8	452.1	20.0	32.0	10.7
:JG-3	197	8.3	74.3	21.2	13.5	4.8	452.6	20.2	32.4	12.9
Mean		8.5	72.4	22.9	13.3	4.8	452.4	20.1	32.2	11.8
Standard	deviation	0.3	2.8	2.3	0.4	0.0	0.4	0.1	0.3	1.6
Recomme	ended	8.9	73.0	23.6	11.4	6.0	453.0	20.7	40.1	16.8
:W-1	72	33.1	257.8	133.3	52.1		187.5	12.5	23.2	13.3
Recomme	ended	35.0	257.0	119.0	47.0		162.0	11.0	23.5	14.6
:BE-N	72	22.2	261.4	377.8	64.9		1181.2	83.9	163.4	69.7
Recomme	ended	22.0	235.0	360.0	60.0		1025.0	82.0	152.0	67.0
:STM-1	72	0.0	2.9	9.9	14.0		605.5	132.4	261.7	84.0
Recomme	ended	0.6	8.7	4.3	0.9		560.0	150.0	259.0	79.0
:GH	72	0.0	3.0	18.0	2.3		20.7	19.2	45.7	23.7
Recomme	ended	0.8	5.0	3.0	0.3		20.0	25.0	60.0	29.0
:JR-1	72	3.3	6.3	7.9	1.5		59.7	21.0	48.0	25.2
Recomme	ended	5.2		2.3	0.7		40.0	19.7	47.1	23.5
:JA-1	72	26.5	104.0	12.8	24.8		305.8	4.6	13.1	13.2
Recomme	ended	28.4	105.0	7.3	11.8		307.0	5.1	13.5	11.0

Element	Line	Crystal	Peak angle	Count time (seconds)	Background 1 angle
ARL 8420+					
Goniometer 1					
Р	Ka1,2	PET	89.54	50	91.97
Si	Ka1,2	PET	109.21	20	112.44
Al	Ka1,2	PET	145.12	30	148.85
Mg	Ka1,2	AX06	20.2	50	22.25
Na	Ka1,2	AX06	24.37	50	26.99
Goniometer 2					
Fe	Ka1,2	LiF200	57.52	25	60.42
Ті	Ka1,2	LiF200	86.14	40	88.8
Са	Ka1,2	LiF200	113.09	25	115.1
К	Ka1,2	LiF200	136.69	50	139.36
Mn	Ka1,2	LiF220	95.2	50	98

Appendix Cii: Data accuracy and precision - XRF operating conditions (major elements).

llimator Detector
Fine FPC
I. Coarse FPC
I. Coarse FPC
I. Coarse FPC
d. Coarse FPC
Fine FPC

Append	lix Cii: Data accur	acy and precision -	XRF operating co	nditions (trace ele	ements).
Element	Line	Crystal	Peak angle	Count time (seconds)	Background 1 angle
Philips PW1400				(0000//00)	ungio
Nb	Ka1,2	LiF220	30.43	100	33.025
Zr	Ka1,2	LiF220	32.115	80	33.025
Y	Ka1,2	LiF220	33.925	80	33.025
Sr	Ka1,2	LiF220	35.88	80	36.58
Rb	Ka1,2	LiF220	38.045	80	38.7
Th	Lai	LiF220	39.3	100	41.015
Ga	Ka1,2	LiF200	38,965	40	39.565
Zn	Ka1.2	LiF200	41.805	40	42.605
Ni	Ka1,2	LiF200	48.625	40	49.885
cont.					
Element	Count time	Background 2	Count time	Collimator	Detector
Philips PW1400	(Seconds)	angle	(Seconds)		
Nb	50	29.83	50	Fine	Scint
7r	50	29.83	50	Fine	Scint
· Y	50	20.00	50	Fine	Scint
' Sr	50	33 025	50	Fine	Scint
Bh	50	26 52	50	Fine	Scint
Th	50	29.7	50	Fine	Scint.
Ga	20	29.265	20	Fine	EDC + Soint
Ga Zn	20	20 565	20	Fine	FPC + Soint.
Ni	20	47.125	20	Fine	FPC + Scint.
ARL 8420+					
Goniometer I	164	1:000	111 04	100	114.07
Ce			110.04	100	114.07
Na	Lai		112.07	100	114.07
La	Lai		138.78	160	141./8
Ba	Lai	LIF220	154.17	100	
Goniometer 2	14.4.0	1:5000	57 50	05	00.40
Fe	Ka1,2	LIF200	57.52	25	60.42
Ti	Ka1,2	LIF200	86.14	40	88.8
Co	Ka1,2	LIF200	/7.83	100	81.03
Cr	Ka1,2	LIF200	107.02	100	109.81
V	Ka1,2	LiF200	123.17	100	126.37
Sc	Ka1,2	LiF200	97.7	100	
Ca	Ka1,2	LiF200	113.09	25	115.1
Cu	Ka1,2	LiF200	45.03	60	45.63
cont.					
ARL 8420+					
Goniometer 1					
Ce	160	110.14	160	Std. Coarse	FPC
Nd	160	110.14	160	Std. Coarse	FPC
La	80	136.78	80	Std. Coarse	FPC
Ва		153.17	100	Std. Coarse	FPC
Goniometer 2					
Fe	10	55.25	10	Fine	FPC
Ti	15	84.25	15	Fine	FPC
 Co	40	74.03	40	Fine	FPC
Cr	40	104.61	40	Fine	FPC
V	40	121 17	40	Fine	FPC
v Sc	-10	95 9	100	Fine	FPC
Ca	15	110 1	15	Fine	FPC
	30	44 43	30	Fine	FPC
<u>u</u>	50		00		

		JE	3-1a		WS 1a				
	Run 35.1.13	Run 36.1.13	Average	Recommended	Run 35.1.1	Run 36.1.1	Average		
La	37.1	36.3	37.5	38.0	27.6	26.0	25.9		
Се	67.4	62.3	<i>65.9</i>	67.0	59.4	53.0	<i>53.9</i>		
Nd	25.5	25.4	26.0	27.0	35.0	31.6	32.3		
Sm	5.1	5.0	5.1	5.0	7.5	7.1	7.1		
Eu	1.48	1.49	1.55	1.52	2.23	2.23	2.28		
Gd	4.2	4.4	4.5	4.7	7. 9	7.2	6.3		
Tb	0.60	0.60	0.63	0.70	1.04	0.95	1.08		
Yb	2.08	2.17	2.14	2.10	2.69	2.50	<i>2.58</i>		
Lu	0.27	0.35	0.28	0.31	0.45	0.37	0.36		
Та	1.65	1.63	1.62	2.00	1.14	1.07	1.06		
Th	9.45	9.40	9.61	8.80	3.22	3.10	3.10		
Hf	3.62	3.61	3.70	3.40	5.76	5.38	5.36		
Sc	29.0	30.1	28.2	29.0	30.6	30.1	29.3		
Со	39.5	39.5	<i>39.5</i>		50.4	47.2	47.1		
U	1.82	1.76	1.82	1.60	0.69		0.74		

Appendix Ciii: Data accuracy and precision - INAA standard tables.



Appendix Civ: Data accuracy and precision

Appendix Civ: Data accuracy and precision



Leicester XRF values compared with duplicate samples run by ICPMS at the Australian National University (ANU), Canberra. Concentrations are in ppm. (NB the XRF and ICPMS methods were run using different sample splits, however, the agreement is still good).

Appendix C.15



Leicester University INAA and ANU ICPMS results compared with Leicester University XRF data. This shows how XRF data of all the above elements, at these levels of concentration, may be effected by high levels of inaccuracy. Concentrations are in ppm.





Appendix C.17



Appendix C.18







Appendix C.20



Appendix C.21

Appendix D XRF Data

Appendix Di	-	Malaita Volcanics Group.
Appendix Dii	-	Maramasike Volcanics (NMAS).
Appendix Diii	-	Maramasike Volcanics (SYV).
Appendix Div	-	Makira Volcanics (MORB and OJP-type).
Appendix Dv	-	Ulawa Volcanics.

All the major analyses were done on a fused disc, and totals between 98.5 and 101.5 % were deemed acceptable (Appendix B). Trace element values are given in parts per million (ppm) and were carried out on powder pellets (Appendix B). If a value of zero is shown then it falls well below the detection limit; see Appendix C for individual element detection limits.

Nb values with a suffix † were carried out on XRF long count times (Appendix C). MB8A* is a feldspar megacryst.

Mg number (**Mg** #) = [(MgO/40.304)/((MgO/40.304) + (Fe₂O_{3(t)}/90.622))]x100 when Fe₂O₃/FeO=0.15.

Na 8.0, Fe 8.0 and Si 8.0 calculated as stated in Klein and Langmuir (1987; 1989) Na 8.0 = Na₂O + (0.373 x MgO) - 2.98. Fe 8.0 = FeO + (1.664 x MgO) - 13.313 (when $Fe_2O_3/FeO=0.15$). Si 8.0 = SiO₂ + (0.31 x MgO) - 2.48.

CIPW Norms were calculated using 'CIPW norms 3.2' computer programme (D.R. Mason, 1990), assuming a ratio of $Fe_2O_3/FeO = 0.15$.

	Kwaleunga River.													
	KW1	KW2	кwз	KW4	KW5	KW6	KW7	KW8	KW9	KW10	KW12	KW13	KW14	KW15
SiO2	50.85	50.20	51.30	50.04	50.93	50.14	51.33	50.42	49.90	50.53	49.91	49.45	49.03	49.05
TiO2	1.570	1.580	1.770	1.750	1.640	1.730	1.780	1.470	2.450	1.610	1.550	1.330	1.380	1.390
AI2O3	13.63	13.53	13.52	13.25	14.40	13.35	13.87	13.77	11.65	14.01	13.48	14.12	13.82	14.45
Fe2O3(t)	13.65	13.53	13.95	14.54	12.58	14.49	13.96	13.13	18.43	13.13	13.72	12.63	12.71	12.82
MnO	0.17	0.17	0.19	0.18	0.16	0.17	0.19	0.17	0.26	0.16	0.17	0.18	0.23	0.19
MgO	7.56	7.18	6.79	6.79	7.19	7.15	7.06	7.61	5.70	7.63	7.57	7.52	7.90	7.40
CaO	10.70	10.69	10.31	10.48	10.89	10.18	10.60	10.93	8.39	10.67	10.45	12.08	11.47	12.43
Na2O	1.93	2.06	2.11	2.18	2.37	2.05	2.18	1.91	2.21	1.99	1.94	1.84	2.48	1.84
K2O	0.145	0.136	0.142	0.159	0.187	0.165	0.164	0.136	0.296	0.090	0.156	0.144	0.121	0.060
P2O5	0.126	0.132	0.160	0.145	0.142	0.145	0.149	0.124	0.218	0.131	0.132	0.121	0.125	0.125
Total	100.31	99.21	100.24	99.25	100.48	99.57	101.29	99.68	99.49	99.96	99.08	99.40	99.27	99.76
L.O.I.	0.48	0.28	0.47	0.63	1.37	0.64	0.62	0.64	0.57	0.60	0.71	0.97	1.64	0.73
Mg #	55.46	54.40	52.25	51.22	56.24	52.60	53.21	56.58	41.02	56.65	55.37	57.24	58.29	56.48
Nb	4.7	5.5	6.1	5.9	4.9	5.6†	6.4	5.2	8.6†	5.4†	5.6	4.4†	4.4	4.7
Zr	92.8	96.9	111.7	104.2	95.8	104.4	106.4	87.2	160.6	94.1	93.7	81.9	81.4	85.1
Y	30	32	34	34	31	32	32	29	48	28	30	27	27	26
Sr	140	143	143	143	143	144	148	145	145	143	138	124	117	130
Rb	2.7	2.0	3.2	2.6	3.0	2.6	2.5	1.7	5.1	1.4	3.0	3.7	2.1	0. 9
Th	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Ga	20	20	21	21	20	21	20	19	24	19	19	20	18	20
Zn	85	91	90	90	97	94	105	90	121	91	102	81	81	85
Ni	68	65	54	52	74	54	58	73	24	67	66	101	95	102
Sc	32	40	38	38	44	42	40	37	35	40	39	33	32	37
V	333	342	389	350	360	354	379	326	459	361	335	295	294	311
Cr	133	135	74	84	157	92	94	156	7	134	135	190	191	196
Co	54	52	51	51	50	54	54	50	58	53	53	53	51	54
Cu	102	102	106	107	104	105	107	94	106	105	100	142	138	141
Ba	40	38	43	45	43	43	41	44	60	28	42	31	28	25
La	~	10	9	11	11	11	6	9	11		9	1	8	10
Ce	23	17	20	21	14	21	17	19	31	11	16	14	11	19
Na	12	13	10	10	11	14	15	14	22	11	12	11	9	12
A1203/TiO2	8.68	8.56	7.64	7.57	8.78	7.72	7.79	9.37	4.76	8.70	8.70	10.62	10.01	10.40
CaO/Al2O3	0.79	0.79	0.76	0.79	0.76	0.76	0.76	0.79	0.72	0.76	0.78	0.86	0.83	0.86
Zr/Nb	19.74	17.62	18.31	17.66	19.55	18.64	16.63	16.77	18.67	17.43	16.73	18.61	18.50	18.11
	3.06	3.03	3.28	3.10	3.11	3.27	3.29	2.98	3.32	3.33	3.13	3.00	3.03	3.29
Na 8.0	1.77	1.76	10 10	10.64	2.07	1.74	10.59	10.77	10.01	10.00	11.78	10.10	2.45	1.62
FE 0.0	50.71	10.41	50.02	10.04	50.69	10.99	10.50 51.04	50.30	12.21	50.42	11.22	10.19	10.09	10.15
51 0.0	50.71	49.95	50.92	49.00	50.00	49.00	51.04	50.50	49.19	50.42	49.70	43.30	49.00	40.00
Quartz	3.02	2.49	4.23	2.16	1.47	2.62	2.92	2.55	4.39	2.65	2.29	0.79	0	0.19
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Orthoclase	0.86	0.8	0.84	0.94	1.11	0.98	0.97	0.8	1.75	0.53	0.92	0.85	0.72	0.35
Albite	16.33	17.43	17.86	18.45	20.06	17.35	18.45	16.16	18.7	16.84	16.42	15.57	20.99	15.57
Anorthite	28.1	27.27	27	25.9	28.1	26.74	27.58	28.6	20.99	29.03	27.61	29.84	26.22	30.99
Nepheline	U A	0	0	0	0	0	0	0	U A	0	0	0	0	U
Leucite	10.06	0	10.19	20.90	20 62	18.02	10.02	20.47	16.02	10.01	10.21	0	U 24 E0	0
	19.90	20.0 22 70	13.10	20.05	20.02	25 59	13.32 24 11	20.41 21 F	10.20 27 F	25 12	25 70	24.14	27.00 13 97	44.03 21 77
Olivine	20.20 N	دی./۲ ۵	23.79 N	23.74 ۵	<u>د</u> د.بع ۱	20.00 N	<u>د</u> م	2 4 .0	27.5 0	0	23.72 N	<u>دد.</u> ۱۴ ۱۹	6.69	۲۱.// ۵
Magnetite	235	2 33	2 <u>⊿</u> 1	2.51	217	25	2 4 1	2.26	3 18	2.26	237	2 18	2 10	2 21
limonito	2.00	2.00	3.36	3 32	3 11	3 29	3 38	2.79	4 65	3.06	2.94	2.53	2.13	2.21
Anatite	03	0.31	0.38	0.34	0.34	0.34	0.35	0.29	0.52	0.31	0.31	0.29	0.3	0.3
CHM	99.15	98.04	99.04	98.26	99.41	98.32	100 1	98.54	97.92	98.82	97,89	98.32	98 17	98.65
30W	55.15	30.04	00.04	00.20	00.41	50.0 <u>C</u>		55.04	01.02	50.0E	57.00	33.0E	55.17	50.00

Appendix Di: XRF Data - Malaita Volcanic Group (OJP lavas).

	KW16	KW17	KW18	KW19	KW20	KW21	KW22	KW23	KW24	KW25	KW26	KW27	KW28	ML696
8102	50.00	40.07	50.92	40.04	40.94	40.69	40.05	40.00	50.00	40.07	40.50	40.01	50.00	50.55
5102	1 4 1 0	49.97	1 200	49.34	49.84	49.68	49.85	49.83	1 650	49.97	49.50	49.31	50.28	50.55
A12O3	14 17	14 50	12 10	14.23	14 44	14.07	1/ 02	12.86	14 22	14 11	14 20	1.210	13 71	13 57
Ee2()3(t)	12.60	12.28	12.13	12.62	12 01	14.07	14.92	13.00	19.20	19.11	19.20	14.14	13.71	14.36
MnO	0 19	0 19	0.20	0.18	0.17	0.20	0.24	0 17	0 17	0.18	0.20	0.20	0.21	0.22
MaQ	7 92	8 13	8 70	7 39	8.63	8.39	6 92	7.61	7 42	8.56	7.66	8 29	6.21	6.09
CaO	11.82	12 16	11 27	12.98	10.06	11.86	11.96	11.05	10.69	12 15	12 74	12 15	11 15	11.03
Na2O	1.78	1.67	2.79	1.73	1.49	1.96	2 15	1.96	2.06	1.62	1 70	1.81	2.13	2.16
K20	0.108	0.076	0.120	0.047	0.306	0.061	0.178	0.138	0.110	0.097	0.051	0.056	0.183	0.192
P2O5	0.126	0.097	0.089	0.103	0.104	0.106	0.132	0.127	0.153	0.099	0.103	0.103	0.147	0.151
Total	100.55	100.24	99.80	99.82	99.22	99.38	99.76	99.64	100.34	99.97	99.78	99.12	99.60	100.01
L.O.I.	1.03	0.70	1.61	0.64	3.22	1.47	0.44	0.61	0.66	0.87	0.46	0.45	0.23	0.24
Mg #	58.39	59.82	61.15	56.83	60.05	61.42	56.77	56.19	56.13	61.46	58.06	61.15	49.99	48.81
Nb	4.8	3.2†	3.3	3.8†	3.5	3.8	5.8	5.3	5.8	3.5	3.2	3.9	5.2	5.8
Zr	83.1	61.6	59.8	70.0	66.6	67.7	94.1	91.6	103.6	64.4	69.9	69.6	101.3	108.5
Y	27	23	23	25	24	24	33	29	32	23	24	25	32	34
Sr	120	115	100	122	175	193	155	146	150	112	120	129	146	149
Rb	2.1	2.4	2.6	0.6	4.9	1.4	2.1	2.7	1.6	2.6	1.5	1.6	3.3	2.7
Th	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Ga	18	19	16	18	14	16	20	20	20	17	18	19	22	22
Zn	74	64	64	79	79	75	87	87	97	76	79	7 9	100	95
Ni	101	104	75	94	95	94	82	86	72	97	94	8 9	66	62
Sc	32	34	42	36	48	39	39	37	41	31	35	43	36	29
v	292	293	315	287	301	295	384	349	358	265	280	295	334	344
Cr	179	161	92	196	216	193	117	157	141	189	191	182	120	105
Co	51	53	50	55	52	52	57	54	54	51	53	53	51	54
Cu	145	135	136	139	134	137	123	103	101	135	141	144	101	98
Ва	31	27	35	25	82	30	58	42	41	27	22	26	47	50
La	7	10	4	8	7	8	10	9	12	7	9	9	9	9
Ce	15	16	9	12	6	9	10	15	22	11	17	8	20	21
Nd	9	10	7	8	5	7	13	12	14	10	8	5	14	15
AI203/TIO2	10.05	12.39	10.16	11.86	11.37	11.73	9.50	8.88	8.62	12.71	11.93	11.69	8.46	8.03
CaO/AI2O3	0.83	0.84	0.92	0.91	0.70	0.84	0.80	0.80	0.75	0.86	0.90	0.86	0.81	0.81
Zr/Nb	17.31	19.25	18.12	18.42	19.03	17.82	16.22	17.28	17.86	18.40	21.84	17.85	19.48	18.71
Zr/Y	3.06	2.67	2.61	2.85	2.76	2.80	2.87	3.13	3.26	2.76	2.88	2.81	3.17	3.24
Na 8.0	1.75	1.72	3.06	1.51	1.73	2.11	1.75	1.82	1.85	1.83	1.58	1.92	1.47	1.45
Fe 8.0	10.91	10.90	11.98	9.96	12.28	10.96	8.51	10.96	10.38	11.43	10.26	10.78	9.17	9.31
SI 8.0	50.31	50.01	51.04	49.15	50.04	49.80	49.52	49.71	50.62	50.14	49.39	49.40	49.73	49.96
Quartz	1.74	1.18	0	0.57	2.44	0.01	0.75	1.48	2.79	1.05	0.82	0.2	2.67	2.97
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Orthoclase	0.64	0.45	0.71	0.28	1.81	0.36	1.05	0.82	0.65	0.57	0.3	0.33	1.08	1. 1 3
Albite	15.06	14.13	23.61	14.64	12.61	16.59	18.19	16.59	17.43	13.71	14.39	15.32	18.02	18.28
Anorthite	30.36	31.85	20.38	30.93	31.81	29.41	30.54	28.61	29.26	30.94	30.97	30.29	27.31	26.77
Nepheline	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Leucite	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Diopside	22.6	22.87	28.63	27.04	14.32	23.55	23.02	20.94	18.81	23.51	25.98	24.02	22.51	22.49
Hypersthene	23.89	24.13	14.92	20.58	30.22	23.86	19.86	24.5	24.52	24.71	21.6	23.34	20.97	21.1
Olivine	0	0	5.85	0	0	0	0	0	0	0	0	0	0	0
Magnetite	2.19	2.12	2.14	2.18	2.23	2.04	2.04	2.3	2.25	2.08	2.15	2.04	2.41	2.48
limenite	2.68	2.22	2.28	2.28	2.41	2.28	2.98	2.96	3.13	2.11	2.26	2.3	3.08	3.21
Apatite	0.3	0.23	0.21	0.24	0.25	0.25	0.31	0.0	0.30	0.23	0.24	0.24	0.35	0.36
SUM	99.45	99.18	90./3	90.73	90.1	90.33	90./0	90.49	99.Z	90.92	90./1	90.09	90.41	90./0

Appendix E	Di: XRF Data	a - Malaita	Volcanic	Group	(OJP	lavas)
			the second se			

-					Singga	lo River							
	ML697	ML698	ML699	ML700	TLB1	TLB2	TLB3	TLB4	TLB5	TLB7	TLB8	TLB9	TLB10
SiO2	50.08	50.50	49.29	50.11	50.98	49.71	50.55	50.56	49.35	49.39	49.44	50.07	49.81
TiO2	1.570	1.590	1.660	1.590	1.460	1.690	1.480	1.490	1.440	1.500	1.520	1.480	1.470
AI2O3	13.51	13.78	13.78	13.56	14.06	15.48	13.79	13.87	13.89	14.42	14.46	13.90	13.77
Fe2O3(t)	13.73	13.70	13.81	13.92	13.23	12.39	13.51	13.15	13.48	12.65	12.57	13.48	13.68
MnO	0.19	0.19	0.21	0.18	0.20	0.25	0.20	0.19	0.21	0.22	0.27	0.18	0.21
MgO	7.34	7.61	7.76	7.35	7.22	7.98	6.96	7.28	7.30	7.32	7.88	7.37	7.05
CaO	10.77	10.49	10.9 9	10.95	11.53	9.77	11.38	11.17	11.51	11.65	11.17	11.32	11.63
Na2O	1.99	1.96	2.00	2.00	1.99	2.04	1.95	1.99	1.93	2.05	1.99	1.95	1.93
K2O	0.166	0.163	0.076	0.138	0.161	0.081	0.141	0.149	0.142	0.158	0.164	0.157	0.142
P2O5	0.135	0.131	0.144	0.142	0.127	0.140	0.131	0.128	0.123	0.128	0.129	0.124	0.129
Total	99.48	100.12	99.72	99.96	100.95	99.52	100.10	99.98	99.37	99.50	99.59	100.03	99.82
L.O.I.	0.43	0.63	0.12	0.47	0.42	1.18	0.61	0.42	0.61	0.58	0.98	0.58	0.56
Mg #	54.59	55.54	55.82	54.28	55.10	59.15	53.67	55.45	54.91	56.54	58.50	55.14	53.68
Nb	5.2	5.2	5.4	5.1	4.6	4.9	5.1	4.6	4.6	4.7	4.6	4.5†	5.1
Zr	94.0	91.3	102.8	100.6	88.5	98.0	90.5	89.2	86.7	86.9	88.1	84.7	87.6
Ŷ	31	29	33	32	29	27	29	30	29	30	31	29	29
Sr	142	141	146	141	140	153	140	140	139	142	146	139	143
RD	3.5	3.4	1.3	2.0	2.9	1.4	2.1	3.2	2.9	2.5	3.1	2.0	3.3
in Ga	20	20	21	20	10	21	20	21	10	20	20	20	10
Ga Zn	20 92	82	96	90	98	21	99	103	87	92	88	84	84
Ni	73	73	70	70	75	88	77	73	80	84	77	70	79
Sc	34	42	36	32	35	44	32	36	36	41	35	35	35
v	326	339	354	325	318	372	314	340	323	338	352	343	305
Cr	133	158	149	130	135	152	130	128	120	131	127	124	116
Co	53	54	55	51	54	59	56	55	53	55	52	54	53
Cu	102	104	105	103	118	122	119	119	113	121	118	120	113
Ва	41	44	25	38	42	32	38	45	38	51	54	51	45
La	8	9	11	11	18	8	-3	10	11	4	9	11	9
Се	17	11	15	14	15	20	15	16	19	13	21	19	19
Nd	11	13	13	12	13	12	9	12	11	9	13	15	12
AI203/TiO2	8.61	8.67	8.30	8.53	9.63	9.16	9.32	9.31	9.65	9.61	9.51	9.39	9.37
CaO/AI2O3	0.80	0.76	0.80	0.81	0.82	0.63	0.83	0.81	0.83	0.81	0.77	0.81	0.84
Zr/Nb	18.08	17.56	19.04	19.73	19.24	20.00	17.75	19.39	18.85	18.49	19.15	18.82	17.18
Zr/Y	3.06	3.11	3.16	3.14	3.07	3.58	3.08	3.02	3.01	2. 9 4	2.88	2.90	3.05
Na 8.0	1.75	1.82	1.91	1.76	1.70	2.04	1.57	1.73	1.67	1.80	1.95	1.72	1.58
Fe 8.0	10.85	11.27	11.61	11.03	10.21	10.75	10.02	10.24	10.56	9.87	10.74	10.68	10.32
Si 8.0	49.88	50.38	49.22	49.91	50.74	49.70	50.23	50.34	49.13	49.18	49.40	49.87	49.52
Quartz	2.16	2.5	0.6	1.91	2.32	1.65	2.71	2.43	0.86	0.34	0.34	1.59	1.55
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	0
Orthoclase	0.98	0.96	0.45	0.82	0.95	0.48	0.83	0.88	0.84	0.93	0.97	0.93	0.84
Albite	16.84	16.59	16.92	16.92	16.84	17.26	16.5	16.84	16.33	17.35	16.84	16.5	16.33
Anorthite	27.44	28.32	28.4	27.62	28.96	32.84	28.46	28.47	28.82	29.68	30.04	28.71	28.49
Nepheline	0	0	0	0	0	0	0	0	0	0	0	0	0
Leucite	0	0	0	0	0	0	0	0	0	0	0	0	0
Diopside	20.77	18.89	20.8	21.34	22.66	12.12	22.48	21.5/	22.12	22.5	20.19	22.01	23.49
Olivine	24.43 0	25.97 0	25.48 0	24.38 0	22.73	28.43 0	22.5 0	23.25 0	23.29 0	22.27 0	24.78 0	23.69 0	22.49 0
Magnetite	2.37	2.36	2.38	2.4	2.28	2.14	2.33	2.27	2.32	2.18	2.17	2.32	2.36
Ilmenite	2.98	3.02	3.15	3.02	2.77	3.21	2.81	2.83	2.73	2.85	2.89	2.81	2.79
Apatite	0.32	0.31	0.34	0.34	0.3	0.33	0.31	0.3	0.29	0.3	0.31	0.29	0.31
SUM	98.3	98.93	98.53	98.74	99.82	98.47	98.93	98.84	98.21	98.4	98.51	98.87	98.64

	TLB11	TLB12	TLB13	TLB14	TLB15	TLB16	TLB17	TLB18	TLB19	TLB20	TLB21	TLB22	TLB24	TLB25
SiO2	40.04	50.00	E0 22	40.75	40.00	50.40	50.00		40.70	40.50	40.77	40.77	50.00	40.00
5102	49.84	1 200	1 520	49.75	49.26	50.18	50.36	50.63	49.78	49.52	49.77	49.77	50.23	49.90
A12O3	1.490	1/ 19	14 32	12 60	1.500	1.030	12.02	1.490	12.00	1.240	1.140	1.310	1.420	1.440
A1203	12 20	14.10	19.02	13.09	14.12	11.10	13.93	14.00	13.00	11.08	14.22	14.02	14.32	13.77
MnO	0 19	0.19	0.22	0.21	0.26	0.20	0 17	0.20	0.21	0 32	0.17	12.03	0.25	0 10
MaQ	7.03	8 20	7 56	7.02	7 94	7 70	7 75	6 71	7 1 1	7.00	9.17	9.07	9.06	7.85
CaO	11 60	11 58	11.60	11 43	11 52	10.46	11 13	11 60	11 30	13.00	12 56	11.87	9.00	11 40
Na2O	1.98	2 17	2 04	1.96	2.02	2 13	1.88	2 01	1 97	1 83	1 65	1 79	297	1.86
K20	0.153	0.543	0.166	0.130	0.138	0.146	0 124	0 137	0.150	0.045	0.083	0.107	0.054	0.096
P205	0.130	0.106	0.127	0.122	0.131	0.139	0.118	0.124	0.128	0.112	0.096	0.108	0.122	0.155
Total	99.81	99.98	100.18	99.50	99.51	99.84	99.90	100.19	99.29	99.89	100.00	100.10	99.76	100.20
L.O.I.	0.60	1.70	0.62	0.57	0.37	1.08	0.64	0.42	0.48	0.49	0.73	0.88	2.07	0.55
Mg #	54.32	61.12	58.06	53.52	55.56	59.55	57.22	53.30	54.75	57.82	60.56	58.58	59.35	56.55
Nb	4.9	3.6	4.7†	4.5†	5.3	5.3	4.7	4.6	4.6	3.9	3.3†	3.7	3.9	4.3†
Zr	89.7	63.2	87.9	89.5	90.5	96.4	83.5	85.0	87.8	72.1	63.6	71.9	78.5	78.0
Y	32	24	30	29	30	28	29	30	29	24	24	26	26	27
Sr	141	304	138	137	139	149	135	147	138	123	115	120	142	133
Rb	2.9	4.4	3.0	3.3	2.6	1.3	2.9	2.7	3.6	1.5	2.1	2.6	1.8	2.6
Th	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Ga	20	17	20	20	20	19	19	20	20	19	18	18	16	20
Zn	90	77	81	91	88	85	93	85	85	85	71	81	85	82
Ni	74	97	79	77	80	83	79	82	83	103	93	86	84	86
Sc	35	37	41	33	40	46	35	33	36	48	40	38	42	32
v	326	280	355	310	339	371	328	325	321	311	283	319	316	310
Cr	111	205	129	117	142	136	128	120	119	239	210	133	156	132
Co	54	51	54	54	53	57	52	53	50	55	52	49	51	52
Cu	119	140	116	119	119	125	115	114	118	141	137	124	123	112
Ва	39	123	43	36	53	70	33	52	40	20	29	33	30	36
La	9	8	8	9	7	8	8	9	9	7	9	6	6	7
Ce	18	13	15	12	17	13	18	14	18	15	12	14	13	12
Nd	11	9	13	9	9	10	10	11	11	0	10	8	/	12
AI203/TiO2	9.47	11.82	9.36	9.25	9.41	9.31	9.88	9.44	9.51	12.16	12.47	10.70	10.08	9.56
CaO/AI2O3	0.82	0.82	0.81	0.83	0.82	0.69	0.80	0.83	0.82	0.88	0.88	0.85	0.69	0.83
Zr/Nb	18.31	17.56	18.70	19.89	17.08	18.19	17.77	18.48	19.09	18.49	19.27	19.43	20.13	18.14
Zr/Y	2.83	2.66	2.91	3.13	2.99	3.51	2.88	2.87	3.07	2.95	2.69	2.80	3.05	2.86
Na 8.0	1.62	2.25	1.88	1.60	1.78	2.06	1.79	1.53	1.64	1.46	1.75	1.82	3.00	1.81
Fe 8.0	9.95	10.54	9.95	10.30	10.38	10.00	10.92	9.35	10.01	8.32	10.92	11.28	10.90	11.55
Si 8.0	49.54	50.15	50.19	49.45	49.06	50.11	50.28	50.23	49.50	49.21	49.85	49.79	50.25	49.85
Quartz	1.43	0	1.3	1.66	0.33	1.42	2.17	2.66	1.67	0.68	0.73	0.73	0	1.28
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Orthoclase	0.9	3.21	0.98	0.77	0.82	0.86	0.73	0.81	0.89	0.27	0.49	0.63	0.32	0.57
Albite	16.75	18.36	17.26	16.59	17.09	18.02	15.91	17.01	16.67	15. 49	13.96	15.15	25.13	15.74
Anorthite	29.16	27.35	29.43	28.17	29.05	31.43	29.21	28.94	28.59	32.8	31.15	29.91	25.58	28.94
Nepheline	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Leucite	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Diopside	22.78	24.11	22.46	22.97	22.51	16.08	20.8	23.03	22.4	26.56	25.05	23.26	18.86	21.96
Hypersthene	22.2	17.26	22.36	22.69	23.12	25.53	24.75	21.19	22.59	18.51	23.11	24.36	17.95	25.12
Olivine	0	4.14	0	0	0	0	0	0	0	0	0	0	5.74	0
Magnetite	2.29	2.02	2.12	2.36	2.28	2.05	2.25	2.28	2.28	1.98	2.08	2.21	2.14	2.34
limenite	2.83	2.28	2.91	2.81	2.85	3.1	2.00	2.83	2.//	2.30	2.17	2.49	2./	2./3
Apatite	0.31	0.25	0.3	0.29	0.31	0.33	0.28	0.29	0.0	0.27	0.23	0.26	0.29	0.37
SUM	98.67	98.98	99.12	98.32	90.35	90.02	90./Ø	99.04	90.10	90.9	90.97	33	90./1	33.00

	TLB26	TLB27	TLB28	TLB29	TLB30	TLB31	ML680	ML681	ML682	ML683	ML684	ML686	ML687	ML689
SiO2	50 38	50 34	50 23	50 75	49 54	49 59	50.61	50 55	49 33	50 11	50.38	50 36	49 43	49 97
TIO2	1.620	1.630	1.530	1.550	1.530	1.490	1.670	1.580	1.580	1.470	1.500	1.470	1.190	1.220
A12O3	13.43	13.95	14.49	14.28	14.49	13.72	13.76	13.91	14.81	13.89	14.09	13.86	14.23	14.51
Fe2O3(t)	14.13	13.53	12.39	12.94	12.92	13.52	13.74	13.75	12.44	13.43	13.24	13.27	12.63	11.35
MnO	0.20	0.19	0.27	0.21	0.29	0.1 9	0.20	0.21	0.34	0.19	0.23	0.19	0.18	0.23
MgO	7.49	7.23	7.55	7.48	7.23	8.17	7.05	7.19	6.98	7.61	7.60	7.44	7.50	8.60
CaO	9.65	10.94	11.22	11.49	11.91	10.68	10.91	11.17	11.65	11.44	11.31	11.46	12.92	12.12
Na2O	1.74	2.04	2.03	2.00	1.95	1.78	2.06	2.01	2.11	1.90	1.97	1.92	1.70	1.87
K2O	1.126	0.172	0.163	0.065	0.045	0.056	0.185	0.159	0.160	0.069	0.122	0.107	0.047	0.095
P2O5	0.132	0.144	0.128	0.132	0.134	0.125	0.142	0.132	0.135	0.124	0.128	0.124	0.106	0.106
Total	99.89	100.17	100.00	100.92	100.03	99.32	100.33	100.67	99.55	100.25	100.59	100.20	99.94	100.09
	0.79	0.63	0.71	0.41	0.49	0.66	0.50	0.50	0.57	0.56	0.65	0.52	0 66	0 78
L.O.I. Ma #	54 38	0.03 54 59	57.81	56 52	0.40 55 72	57.60	0.50 53.57	0.59	0.57	0.50	56 34	0.52	57 18	63.01
Mg #	54.50	54.50	57.01	30.52	55.72	57.00	55.57	34.04	55.70	50.05	50.54	55.70	57.10	00.01
Nb	4.8	5.3	5.3	4.8	5.5	4.6	5.5	4.8	5.2	4.7	4.4	5.2	3.9	3.2
Zr	89.5	100.0	89.0	94.0	92.1	86.2	99.6	90.6	95.4	87.0	89.2	86.7	69.9	70.1
Y	29	32	29	30	31	26	32	30	32	29	30	29	26	24
Sr	134	146	142	144	146	139	143	141	145	141	140	146	120	119
Rb	3.5	3.5	2.5	1.3	0.9	1.5	3.9	2.8	1.9	0.6	2.5	2.1	1.3	1.1
Th	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Ga	21	21	20	21	21	20	21	20	21	19	21	19	18	17
Zn	83	100	93	96	93	89	102	87	91	84	84	86	79	78
Ni	82	75	88	77	82	73	69	77	85	75	80	77	96	98
Sc	33	39	46	36	40	32	43	40	37	35	39	35	32	39
v	333	357	353	343	349	339	349	347	354	311	327	326	275	297
Cr	152	129	135	107	121	130	123	133	145	111	118	126	200	211
Co	53	52	52	53	55	53	50	52	54	52	54	52	55	53
Cu	96	107	118	119	121	116	104	106	121	119	119	115	145	142
Ba	42	48	46	34	44	34	42	44	53	30	43	57	22	25
La	10	10	6	8	11	6	9	9	11	14	10	8	10	11
Ce	10	10	12	19	20	14	12	20	10	14	10	10	12	0
Nu	14	10	10	10	10	U	14	14			10	••	10	Ū
AI203/TiO2	8.29	8.5 6	9.47	9.21	9.47	9.21	8.24	8.80	9.37	9.45	9.39	9.43	11.96	11.89
CaO/AI2O3	0.72	0.78	0.77	0.80	0.82	0.78	0.79	0.80	0.79	0.82	0.80	0.83	0.91	0.84
Zr/Nb	18.65	18.87	16.79	19.58	16.75	18.74	18.11	18.88	18.35	18.51	20.27	16.67	17.92	21.91
Zr/Y	3.12	3.10	3.11	3.19	3.01	3.37	3.10	3.06	2.94	3.03	2.96	2.99	2.74	2.88
Na 8.0	1.55	1.76	1.87	1.81	1.67	1.85	1./1	1./1	1.73	1.76	1.82	1./2	1.52	2.10
Fe 8.0	11.44	10.49	10.03	10.39	9.96	12.04	10.37	10.61	9.12	11.03	10.85	10.61	10.16	10.87
SI 8.0	50.22	50.10	50.09	50.59	49.30	49.04	50.32	50.50	49.01	49.99	50.20	50.19	49.20	50.16
Quartz	1.41	2.06	1.47	2.12	0.95	1.83	2.49	2.09	0.35	1.69	1.6	2.07	0.69	0.08
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Orthoclase	6.65	1.02	0.96	0.38	0.27	0.33	1.09	0.94	0.95	0.41	0.72	0.63	0.28	0.56
Albite	14.72	17.26	17.18	16.92	16.5	15.06	17.43	17.01	17.86	16.08	16.67	16.25	14.39	15.82
Anorthite	25.51	28.4	29.95	29.8	30.65	29.28	27.75	28.46	30.47	29.17	29.24	28.89	31.06	30.92
Nepheline	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Leucite	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Diopside	17.8	20.63	20.49	21.73	22.76	18.86	21.07	21.6	21.84	22.1	21.48	22.42	26.65	23.32
Hypersthene	26.75	23.86	23.54	23.34	22.34	27.33	23.42	23.69	21.54	24.23	24.28	23.42	21.08	23.86
Olivine	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Magnetite	2.44	2.33	2.14	2.23	2.23	2.33	2.37	2.37	2.15	2.32	2.28	2.29	2.18	1.96
limenite	3.08	3.1	2.91	2.94	2.91	2.83	3.17	3	3	2.79	2.85	2.79	2.26	2.32
Apatite	0.31	0.34	0.3	0.31	0.32	0.3	0.34	0.31	0.32	0.29	0.3	0.29	0.25	0.25
SUM	98.68	99	98.93	99.78	98.93	98.15	99.14	99.47	98.47	99.07	99.43	99.05	98.84	99.09

			Ala'ola	River									
···	ML690	ML692	AL3	AL4	AL5	AL6	AL7	AL8	AL10	AL11	AL14	AL15	AL16
SiO2	50.08	50.00	49.98	49.98	50.27	50.79	51.32	50.43	49.52	50.15	50.08	49.63	49.54
TiO2	1.190	1.190	1.600	1.620	1.590	1.580	1.760	1.610	1.590	1.590	1.060	1.500	1.190
AI2O3	14.31	14.15	13.76	13.89	13.62	13.61	13.47	14.33	13.59	13.61	15.49	13.65	14.65
Fe2O3(t)	12.28	12.47	13.61	12.73	13.50	13.69	13.80	13.03	13.92	13.77	10.82	13.48	11.75
MnO	0.18	0.18	0.17	0.19	0.17	0.19	0.20	0.16	0.20	0.19	0.17	0.19	0.18
MgO	7.53	7.76	6.98	7.53	7.28	7.35	6.90	7.60	6.99	7.02	7.39	7.26	8.91
CaO	12.85	12.77	10.84	11.23	10.89	11.10	9.86	10.48	11.42	11.20	12.50	10.90	12.03
Na2O	1.72	1.71	2.01	2.02	2.04	1.97	2.19	1.94	1.97	2.02	2.20	1.95	1.60
K2O	0.064	0.087	0.148	0.078	0.094	0.183	0.197	0.068	0.056	0.148	0.075	0.140	0.080
P2O5	0.105	0.102	0.129	0.142	0.132	0.129	0.171	0.133	0.134	0.139	0.084	0.117	0.099
Total	100.30	100.41	99.24	99.42	99.58	100.59	99.88	99.79	99.40	99.84	99.85	98.82	100.04
L.O.I.	0.35	0.67	0.57	0.44	0.52	0.75	0.55	0.64	0.36	0.41	1.57	0.57	1.08
Mg #	57.96	58.32	53.56	57.08	54.80	54.69	52.92	56.74	53.03	53.41	60.56	54.77	63.03
Nb	3.7	3.6	5.6	5.7	5.1	5.5	6.8†	5.2	5.8	5.3†	3.3	4.6†	3.4
Zr	69.0	67.6	93.5	100.4	92.3	92.0	124.5	94.6	99.8	99.0	52.4	83.0	62.1
Ŷ	24	24	30	31	29	30	34	28	32	31	21	28	22
Sr	128	122	141	144	142	136	139	146	142	142	153	140	137
RD Th	1.3	2.1	2.0	1.0	1.3	2.6	3.6	0.6	0.0	2.9	1.7	2.8	2.2
in Co	10	10	0	0	20	10	0	10	0	3	10	10	17
Ga	19	10	20	20	20	19	21	19	21	20	10	19	75
Zn	80	70 07	92	99 67	67	93	90	92	95	90 64	00 91	00 77	102
NI Sa	92	97	04	41	27	33	40	47	36	24	35	22	32
30 V	20	260	333	41 257	376	334	3/1	363	333	325	269	335	261
v Cr	207	209	97	128	133	121	341 88	145	128	124	100	147	201
	200	50	5/	53	52	51	49	145	56	55	40	51	48
Cu	141	137	104	102	102	101	98	103	101	102	212	110	119
Ba	25	31	39	27	35	38	42	25	26	44	22	38	31
la	10	7	9	12	7	8	8	7	10	8	7	8	9
Ce	9	10	20	28	17	21	29	18	14	14	9	17	8
Nd	7	9	16	15	17	14	19	12	14	14	6	15	7
A1203/TiO2	12.03	11.89	8.60	8.57	8.57	8.61	7.65	8.90	8.55	8.56	14.61	9.10	12.31
CaO/AI2O3	0.90	0.90	0.79	0.81	0.80	0.82	0.73	0.73	0.84	0.82	0.81	0.80	0.82
Zr/Nb	18.65	18.78	16.70	17.61	18.10	16.73	18.31	18.19	17.21	18.68	15.88	18.04	18.26
Zr/Y	2.84	2.85	3.14	3.23	3.22	3.05	3.65	3.37	3.14	3.18	2.52	2.96	2.89
Na 8.0	1.55	1.62	1.63	1.85	1.78	1.73	1.78	1.79	1.60	1.66	1.98	1.68	1.94
Fe 8.0	9.90	10.45	10.14	10.29	10.55	10.83	10.17	10.67	10.43	10.35	8.40	10.50	11.74
Si 8.0	49.93	49.93	49.66	49.83	50.05	50.59	50.98	50.31	49.21	49.85	49.89	49.40	49.82
Quartz	1.39	1	2.44	1.86	2.4	2.51	4.12	2.99	1.71	2.09	0	1.96	0.32
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	0
Orthoclase	0.38	0.51	0.87	0.46	0.56	1.08	1.16	0.4	0.33	0.87	0.44	0.83	0.47
Albite	14.55	14.47	17.01	17.09	17.26	16.67	18.53	16.42	16.67	17.09	18.62	16.5	13.54
Anorthite	31.14	30.68	28.09	28.6	27.73	27.75	26.34	30.19	28.07	27.63	32.17	28.08	32.56
Nepheline	0	0	0	0	0	0	0	0	0	0	0	0	0
Leucite	0	0	0	0	0	0	0	0	0	0	0	0	0
Diopside	26.28	26.33	20.58	21.58	21.03	21.89	17.81	17.27	22.96	22.39	24.02	20.86	21.65
Hypersthene	20.88	21.69	23.37	23.11	23.78	23.84	24.59	25.77	22.7	22.84	18.8	23.97	25.95
Olivine	0	0	0	0	0	0	0	0	0	0	0.8	0	0
Magnetite	2.12	2.15	2.35	2.2	2.33	2.36	2.38	2.25	2.4	2.37	1.87	2.32	2.03
Ilmenite	2.26	2.26	3.04	3.08	3.02	3	3.34	3.06	3.02	3.02	2.01	2.85	2.26
Apatite	0.25	0.24	0.31	0.34	0.31	0.31	0.41	0.32	0.32	0.33	0.2	0.28	0.23
SUM	99.25	99.34	98.05	98.31	98.42	99.41	98.68	98.66	98.19	98.65	98.93	97.65	99.01

Appendix Di: XRF Data - Malaita Volcanic Group (OJP lavas).

	AL17	ML667	ML668	ML669	ML709	ML710	ML711	ML712	ML713	ML714	ML715	ML716	ML717	ML718
6102	50.11	50.20	50 50	40.00	50.10	50.10	50 70	50.00	50.40	50.00	50.45	50.00	50.00	50.40
5102	1 200	1 650	50.52	49.29	1 500	50.18	50.72	50.00	50.46	50.22	50.15	50.00	50.20	50.42
A12O3	13.25	1/ 10	13 72	13 79	12 00	1364	12.07	12.000	12 61	12 56	12 20	1.050	12 46	12 70
Fe2()3(t)	11.20	13.23	13.72	13.70	13.99	13.04	15.61	14.25	13.01	13.00	14.07	12.10	13.40	13.70
MnO	0.20	0.24	0.18	0.21	0 17	0.20	0.22	0 10	0.10	0.20	0.19	0.20	0.20	0 10
MaQ	8 21	7 27	7 44	7.76	7 77	6.68	6 35	6.01	6.24	6.82	7 13	7 92	6.82	6.89
CaO	11 31	10.88	10 58	10 99	11 14	11 14	0.00	10.68	11 33	11 17	10.00	10.02	11 04	11 32
Na2O	2 13	2 08	2 04	2 00	1 97	2 01	2 19	2.05	2 08	2.05	2 14	1 96	2.05	2 01
K20	0.629	0 125	0 149	0.076	0.083	0.074	0.212	0 188	0 127	0 162	0 125	0.068	0 160	0.069
P205	0.106	0.141	0.142	0.144	0.122	0 135	0.166	0 147	0.140	0 139	0.120	0.141	0.140	0.142
Total	99.04	100.01	100.03	99.72	99.92	99.91	100.26	99.38	99.49	99.87	99.03	100.28	99.60	100.25
L.O.I.	1.25	1.94	0.56	0.38	0.58	0.30	0.36	0.59	0.38	0.35	0.64	0.57	0.33	0.42
Mg #	60.82	55.27	55.12	55.82	57.32	51.81	47.77	52.16	50.60	52.36	53.26	57.30	52.44	52.91
Nb	46	50	53	50	12	56	65	5.6	63	5.6	63	53	50	60
7	4.0	95.6	100 4	100.9	4.3 94.9	09.2	121.2	104.0	103 4	08.0	106.5	00.0	100 7	101 2
21	25	90.0 20	21	21	28	30.3	38	25	22	30.3	32	33.5	32	32
r Sr	169	29	145	1/0	130	1/2	145	1/2	1/8	142	142	146	145	146
Si Ph	83	22	50	17	24	0.8	4.5	42	17	28	23	17	15	20
Th	0.0	0	0.0	0	0	0.0			0	0	0	2	0	0
Ga	15	19	20	20	20	20	22	20	20	21	20	21	20	21
Zn	77	85	96	96	81	95	104	103	97	96	87	93	96	94
Ni	117	67	67	69	78	67	39	58	71	72	54	69	74	73
Sc	39	50	31	37	37	32	39	38	33	37	34	42	41	37
v	277	354	347	356	348	327	403	350	334	330	367	357	360	339
Cr	241	139	109	127	135	120	25	108	143	147	79	145	144	144
Co	52	50	53	54	51	55	54	53	51	54	53	54	53	53
Cu	129	97	102	103	92	102	115	107	106	105	104	104	105	105
Ba	69	24	33	26	29	37	55	45	43	41	39	26	24	31
La	8	9	8	10	10	8	10	9	11	6	12	11	10	9
Ce	9	17	19	16	20	18	22	25	18	15	25	15	17	23
Nd	6	15	15	15	12	10	20	14	14	11	14	12	12	11
			0.07		0.00		c co	7 00	0.45	0.40		0 50	0.00	0.51
AI203/1102	11.04	8.60	8.37	8.30	9.33	8.63	0.09	7.90	8.40 0.92	0.42	0.76	8.58	0.30	8.51
CaU/AI203	0.85	10.77	19.04	17.09	10.00	0.04	18.66	19.57	16 / 1	17.66	16.00	19.95	17.07	0.02
ZI/ND 7r/V	2.84	3 26	3 22	3 25	3.06	3 13	3 21	3.01	3 17	3 12	3 34	3 21	3 15	3 19
21/1 Na 8.0	2.04	1.81	1 84	1 91	1.89	1.52	1.58	1.65	1 43	1 61	1.82	1 93	1 61	1 60
Fe 8 0	10.69	10.29	10.92	11.61	10.93	9.96	10.83	10.58	8.99	10.17	10.79	11.41	10.14	10.15
Si 8.0	50.18	49.97	50.35	49.22	50.09	49.77	50.21	49.66	49.91	49.85	49.88	49.98	49.83	50.08
01010	00110													
Quartz	0	1.89	2.48	0.6	1.67	2.5	3.56	2.37	3.26	2.23	2.72	1.47	2.44	2.64
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Orthoclase	3.72	0.74	0.88	0.45	0.49	0.44	1.25	1.11	0.75	0.96	0.74	0.4	0.95	0.41
Albite	18.02	17.6	17.26	16.92	16.67	17.01	18.53	17.35	17.6	17.35	18.11	16.59	17.35	17.01
Anorthite	24.74	29.01	27.84	28.4	29.09	27.98	24.93	26.48	27.43	27.32	26.29	29.64	27.05	28.38
Nepheline	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Leucite	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Diopside	25.16	19.87	19.61	20.8	20.91	23.16	19.21	21.2	23.17	22.55	18.91	19.47	22.23	22.26
Hypersthene	19.42	24	24.98	25.48	24.58	21.9	24.65	23.64	20.34	22.48	25.03	25.83	22.58	22.55
Olivine	2.37	0	0	0	0	0	0	0	0	0	0	0	0	0
Magnetite	2.05	2.28	2.35	2.38	2.24	2.41	2.69	2.46	2.30	2.41	2.43	2.29	2.4	2.38
limenite	2.28	3.13	3.11	3.15	2.85	3	3.00 0.00	3.19	3.00	3.00	0.20 0.05	0.13	3.06	3.08
Apatite	0.25	0.33	0.34	0.34	0.29	0.32	0.39	0.35	0.33	0.33	0.35	0.33	0.33	0.34
SUM	98.01	98.87	98.86	98,53	98.79	98.7	90.91	90.15	30.31	90.00	97.82	99.15	90.39	99.04

Appendix Di: XRF Data - Malaita Volcanic Group	(OJP	lavas).
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											North M	Igwana	a River
	ML719	ML720	ML722	ML725	ML726	ML728	ML729	ML748	ML749	ML750	MB1	MB2	MB3
SiO2	49.99	50.53	50.85	50.23	50.07	50.99	50.70	50.45	50.70	50.93	51.92	52.87	50.99
TiO2	1.560	1.750	1.670	1.240	1.490	1.500	1.140	1.220	1.320	1.250	3.070	0.830	1.320
AI2O3	14.66	14.86	13.82	14.95	14.29	13.84	14.07	14.76	13.81	14.27	11.17	14.41	14.49
Fe2O3(t)	12.85	12.28	13.72	11.85	12.73	13.32	12.01	11.55	12.17	11.95	19.52	8.49	10.22
MnO	0.33	0.25	0.21	0.27	0.23	0.19	0.17	0.15	0.19	0.18	0.30	0.15	0.19
MgO	7.00	7.90	6.59	9.07	7.30	7.75	9.03	8.91	8.26	7.93	2.24	7.71	7.25
CaO	11.86	10.37	11.10	10.80	11.62	10.84	10.32	12.09	11.59	9.37	7.71	12.21	13.11
Na2O	2.00	2.08	2.12	1.72	1.97	1.95	1.95	2.10	2.49	4.07	2.86	3.31	2.28
K2O	0.059	0.234	0.162	0.174	0.150	0.112	1.274	0.255	0.118	0.209	0.214	0.293	0.108
P2O5	0.134	0.138	0.145	0.110	0.127	0.128	0.094	0.107	0.114	0.106	0.372	0.090	0.109
Total	100.44	100.39	100.40	100.42	99.98	100.62	100.77	101.61	100.76	100.26	99.37	100.37	100.07
L.O.I.	0.33	0.93	0.31	1.04	1.92	0.56	2.12	3.11	1.52	3.46	1.91	2.18	2.03
Mg #	55.05	59.13	51.92	63.25	56.32	56.68	62.83	63.43	60.41	59.87	20.51	67.13	61.47
Nb	5.0	5.2	6.2	3.9	5.3	5.0	3.5	4.3	4.0	4.1	11.4	2.9†	3.7
Zr	94.3	96.6	101.8	68.8	87.5	89.6	57.8	76.3	65.9	68.1	242.3	56.5	63.3
Y	32	27	33	24	32	28	23	25	22	24	72	22	24
Sr	143	146	145	119	194	137	305	118	220	174	170	116	231
Rb	0.7	3.7	2.0	1.7	2.7	2.3	8.2	2.0	2.8	2.5	2.8	5.3	1.3
Th	0	0	0	0	4	0	0	0	0	0	3	10	17
Ga T-	22	21	19	17	19	20	10	19	18	14	30	10	51
Zn	97	85	93	83	80	85	65	102	110	73	132	47	01 45
NI Se	82	01 41	25	100	13	00 /1	30	25	27	94 20	24	37	40
5C V	43	41 290	30	40	40 245	41	25	306	27	39	07 02	212	310
Cr.	140	148	122	247	177	100	207	250	235	235	0	49	84
	56	52	53	50	52	52	52	52	46	200 49	47	33	38
Cu	121	103	104	137	107	118	136	153	126	129	18	68	13
Ba	28	47	43	31	35	37	261	24	29	21	45	19	36
La	6	8	8	8	7	11	3	8	7	8	14	7	10
Ce	14	22	24	12	17	12	7	9	11	10	45	14	13
Nd	10	12	14	6	12	13	7	11	11	7	30	9	8
AI203/TiO2	9.40	8.49	8.28	12.06	9.59	9.23	12.34	12.10	10.46	11.42	3.64	17.36	10.98
CaO/AI2O3	0.81	0.70	0.80	0.72	0.81	0.78	0.73	0.82	0.84	0.66	0.69	0.85	0.90
Zr/Nb	18.86	18.58	16.42	17.64	16.51	17.92	16.51	17.74	16.48	16.61	21.25	19.48	17.11
Zr/Y	2.95	3.56	3.12	2.89	2.75	3.22	2.55	3.02	2.94	2.83	3.38	2.57	2.60
Na 8.0	1.63	2.05	1.60	2.12	1.71	1.86	2.34	2.44	2.59	4.05	0.72	3.21	2.00
Fe 8.0	9.51	10.52	9.5 9	12.09	9.91	11.17	12.16	11.56	11.02	10.28	7.40	6.90	7.64
Si 8.0	49.68	50.50	50.41	50.56	49.85	50.91	51.02	50.73	50.78	50.91	50.13	52.78	50.76
Quartz	1.44	1.74	2.97	0.95	1.5	2.75	0	0	0	0	9.65	0	0.85
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	0
Orthoclase	0.35	1.38	0.96	1.03	0.89	0.66	7.53	1.51	0.7	1.24	1.26	1.73	0.64
Albite	16.92	17.6	17.94	14.55	16.67	16.5	16.5	17.77	21.07	34.44	24.2	28.01	19.29
Anorthite	30.85	30.52	27.72	32.56	29.71	28.68	25.88	30.1	26.16	20.05	17.01	23.6	28.99
Nepheline	0	0	0	0	0	U	U	U	0	0	0	U A	U C
Leucite	0	U 16.47	01.00	16.60	U 70.07	20.02	0	0	0 25 1	U 21.14	U 16 /F	0 20 50	0 29.02
	22.42	10.4/	21.92	10.02	22.31	20.02	20.21	20.04 18.07	18 64	≤1,14 0.81	10.45	23.00	20.92
nyperstnene	21.00	∠0.00 ∧	۲1.02 ۵	29.03 ۸	<u>دد.4</u> د ۲	20.42 N	5 98	4 55	3 17	16.87	0	4 82	0.30
Magnetite	2 2 2 2	2 1 2	2.37	2 04	22	23	2.07	1.99	2.1	2.06	3.37	1.46	1.76
Ilmenite	2 96	3 32	3.17	2.36	2.83	2.85	2.17	2.32	2.51	2.37	5.83	1.58	2.51
Anatita	0.32	0.33	0.34	0.26	0.3	0.3	0.22	0.25	0.27	0.25	0.88	0.21	0.26
SUM	99.34	99.34	99.21	99.39	98.88	99.47	99.72	100.6	99.71	99.23	97.73	99.63	99.19

-	MB4	MB6	MB7	MB8A	MB8B*	MR9	MB10	MR11	MB12	MB13	MB14	MR15	MB16
						11100	MIDTO					MD10	
SiO2	49.73	49.61	50.02	50.32	48.10	50.77	49.76	43.63	49.30	50.29	48.65	50.25	46.96
TIO2	1.250	1.250	1.240	1.320	0.080	1.210	2.230	4.230	1.610	1.480	1.300	1.140	2.200
AI2O3	14.36	14.50	14.52	13.39	31.89	14.04	12.02	9.97	13.34	13.57	13.52	15.82	12.22
Fe203(t)	11.94	12.21	11.57	12.24	1.36	12.28	17.72	22.74	13.55	13.37	13.62	10.37	17.13
Mao	0.20	0.19	0.10	0.19	0.01	0.19	0.25	0.30	0.23	0.27	0.16	0.16	0.25
MgO CaO	12 23	12 /1	0.23	10.19	15 42	11 60	4.97	0.17	11 44	11 20	9.37	12 11	0.11
Na2O	1 66	1 60	1 68	2 20	2 10	2 17	9.24 2.54	9.07 1 72	2.03	1 28	1 72	1 78	1 75
K20	0.067	0.078	0.058	0.302	0.031	0 135	0 274	0.260	0 114	0.086	0.283	0 126	0 135
P205	0.111	0.110	0.109	0.110	0.015	0.079	0.217	0.117	0.123	0.131	0.100	0.078	0.109
Total	100.11	100.09	100.01	101.32	99.71	99.94	99.22	99.77	99.25	99.75	99.41	100.22	99.44
L.O.I.	1.40	1.13	1.14	1.17	0.96	1.19	2.87	0.43	1.80	3.75	1.98	0.84	3.24
Mg #	61.71	59.95	61.53	60.07	50.62	56.87	38.67	37.89	55.48	56.96	60.74	61.57	51.56
Nb	3.8	4.0†	3.7	4.2	0.0	2.7	8†	5.7	4.8	4.9	3.6	3.0	4.3
Zr	74.4	73.6	72.9	73.7	0.0	48.1	146.5	79.4	84.1	84.4	65.9	46.8	57.1
Ŷ	24	24	24	25	1	21	43	31	28	28	24	19	22
Sr	127	118	114	145	182	149	147	119	123	204	115	168	143
Rb	1.5	2.6	1.4	4.9	2.3	2.4	2.0	1.7	2.0	2.5	6.5	1.8	4.2
Th	0	0	0	0	0	0	0	0	0	0	0	0	0
Ga	17	18	18	18	17	19	23	23	20	18	17	18	25
Zn	75	76	77	67	9	58	61	117	72	80	43	59	98
Ni	110	105	111	76	4	38	28	35	68	91	147	85	102
Sc	36	30	36	30	9	44	38	66	35	28	31	38	38
v	285	282	276	292	27	346	440	1145	348	298	256	360	438
Cr	235	221	247	217	13	29	3	0	125	123	110	242	136
Co	51	53	53	48	4	49	56	78	50	50	61	41	71
Cu	136	136	137	137	5	128	72	41	89	87	77	117	123
Ba	22	20	22	50	12	27	35	22	31	26	36	29	33
La	10	6	6	ь 7	<i>'</i>	5	8	3	17	10	0 17	9	4
Ce	13	o g	9 11	12	0	3	25 17	0 8	11	12	8	3	7
Nu	• •	0		12	Ū	Ū	.,	0			Ū	0	•
AI203/TIO2	11.49	11.60	11.71	10.14		11.60	5.39	2.36	8.29	9.17	10.40	13.88	5.55
CaO/AI2O3	0.85	0.86	0.86	0.82		0.83	0.77	0.91	0.86	0.83	0.79	0.83	0.87
Zr/Nb	19.58	18.40	19.70	17.55		17.81	18.31	13.93	17.52	17.22	18.31	15.60	13.28
Zr/Y	3.17	3.11	3.00	2.97		2.32	3.39	2.57	3.03	3.01	2.80	2.50	2.57
Na 8.0	1.87	1.00	10.45	2.30		1.00	10.37	16.74	10.05	11.04	2.24	8.01	1.00
FE 0.0	11.32	10.04	50.00	50.38		50 52	48.82	43.06	49 15	50.25	49.07	50.06	46.99
310.0	40.00	43.00	00.00	00.00		00.02	40.02	40.00	10.10	00.20	10.07	00.00	40.00
Quartz	0.67	1.01	1.32	0	0.52	1.54	2.96	0	0.57	4.27	0	1.24	0
Corundum	0	0	0	0	0.25	0	0	0	0	0	0	0	0
Orthoclase	0.4	0.46	0.34	1.78	0.18	0.8	1.62	1.54	0.67	0.51	1.67	0.74	0.8
Albite	14.05	13.54	14.22	19.38	18.53	18.36	21.49	14.55	17.18	11.68	14.55	15.06	14.81
Anorthite	31.53	32.15	31.91	25.37	76.41	28.17	20.59	18.72	26.95	30.58	28.34	34.81	25.09
Nepheline	0	0	0	0	0	0	0	U	U	U	U	U	0
Leucite	0 00	0	U 22 75	0000	0	0/1⊑	20.10	21 51	0 22 04	0 20 22	10 60	0 24 22	U 22 12
Diopsiae	23.20	∠3.30 22 F7	20.10	23.3 22	3 28 U	24.10	20.10	23.94	23.07	20.02	25.05	29.00 19.01	23.28
	24.40 0	23.31 N	22.00 ۸	23 0 57	0.20	۲.0 <i>1</i>	20.0 4 0	3.84	0	0	3.86	0	4 46
Magnetite	2.06	2 1 1	2	2.11	0.23	2.12	3.06	3.92	2.34	2.31	2.35	1.79	2.95
limenite	2.37	2.37	2.36	2.51	0.15	2.3	4.24	8.03	3.06	2.81	2.47	2.17	4.18
Apatite	0.26	0.26	0.26	0.26	0.04	0.19	0.51	0.28	0.29	0.31	0.24	0.18	0.26
SUM	99.08	99.03	99.03	98.27	99.6	98.7	97.7	96.23	98.08	98.59	98.22	99.33	97.96

Appendix Di: XF	F Data - Malaita	Volcanic	Group	(OJP lavas	s).
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	Kware'	e River						South N	uth Ngwanaa River				
	ML175	ML177	ML212	ML213	ML216	ML217	ML219	ML299	ML313	ML314	ML316	ML317	ML319
SiO2	49.47	50.52	51.07	51.43	49.97	51.22	50.01	50.88	50.04	50.83	50.74	50.71	50.59
TiO2	1.480	1.450	1.510	1.350	1.250	1.600	1.010	1.480	1.430	1.910	1.020	1.610	2.070
AI2O3	13.71	13.82	13.86	14.01	13.17	13.75	14.31	13.80	13.60	13.20	18.99	14.00	12.86
Fe2O3(t)	13.37	13.21	13.05	12.54	13.11	13.40	11.83	13.22	13.58	14.43	8.21	11.18	15.64
MnO	0.20	0.22	0.22	0.19	0.19	0.1 9	0.17	0.18	0.20	0.21	0.14	0.21	0.23
MgO	7.05	7.66	7.49	8.12	10.38	7.47	10.02	7.89	8.59	7.16	6.47	7.32	6.90
CaO	11.82	8.40	11.02	11.03	10.72	10.81	11.84	10.68	10.91	9.46	12.09	11.63	8.83
Na2O	2.16	3.53	1.92	1.86	1.57	1.95	1.43	1.86	1.78	2.30	1.98	2.29	2.36
K20 B205	0.115	1.681	0.142	0.136	0.199	0.128	0.065	0.106	0.122	0.366	0.425	1.042	0.277
F205	0.120	100 60	100 41	100.90	100 66	100 64	100 79	100.22	100.26	100.04	100 15	100 13	0.109
Totar	33.00	100.00	100.41	100.00	100.00	100.04	100.70	100.22	100.50	100.04	100.15	100.10	00.00
L.O.I.	0.43	2.38	0.69	1.00	2.65	0.68	1.06	0.56	0.76	0.85	1.20	0. 9 9	0.82
Mg #	54.25	56.59	56.34	59.28	64.03	55.62	65.57	57.30	58.72	52.73	63.92	59.55	49.80
Nb	4.8	4.3	4.9	4.1	3.8	5.4	2.9†	4.9	4.2	5.7	2.9	5.1	6.0
Zr	92.3	77.1	87.7	87.0	57.6	95. 9	55.0	88.0	80.9	107.4	53.4	94.9	117.8
Y	29	26	29	28	21	31	20	29	27	33	19	26	38
Sr	134	290	131	133	185	136	109	130	126	139	163	148	142
Rb	2.5	8.3	2.7	2.5	3.9	2.7	2.9	1.8	2.9	2.6	3.6	5.6	2.6
in Co	0	10	10	10	16	0	17	0	10	0	10	0	0
Ga	20	79	19	80	65	20	67	19	80	21 57	30	16	22 61
Ni	78	75	73	85	160	72	162	74	100	45	82	84	37
Sc	41	39	43	38	36	40	30	35	26	41	29	39	41
v	319	310	332	299	268	354	230	328	302	354	236	379	378
Cr	160	125	155	160	167	137	175	158	140	37	112	157	15
Co	53	51	54	51	60	50	54	50	55	53	32	51	53
Cu	109	109	112	109	98	102	124	104	104	70	10	84	98
Ва	38	43	35	37	28	33	20	33	41	40	34	79	38
La	9	6	9	8	5	9	6		10	12	5	6	8
Ce	10	14	16	20	10	16	8	17	10	19	4	14	21
Nd	11	10	14	13	7	13	8	10	11	12	10	8	17
AI203/TIO2	9.26	9.53	9.18	10.38	10.54	8.5 9	14.17	9.32	9.51	6.91	18.62	8.70	6.21
CaO/A12O3	0.86	0.61	0.80	0.79	0.81	0.79	0.83	0.77	0.80	0.72	0.64	0.83	0.69
Zr/Nb	19.23	17.93	17.90	21.22	15.16	17.76	18.97	17.96	19.26	18.84	18.41	18.61	19.63
	3.10	2.98	3.06	3.07	2.72	3.13	2.79	3.01	2.95	3.27	2.80	3.64	3.12
Fe 8.0	10.05	10.93	10.50	11 11	15.37	10.78	13 65	11.32	12.00	11 16	4 60	2.04 8.59	11 78
Si 8.0	49.18	50.41	50.91	51.47	50.71	51.06	50.64	50.85	50.22	50.57	50.27	50.50	50.25
Quartz	0.21	0	3.2	3.06	0	3.46	0.22	3.13	1.17	2.48	1.68	0	2.84
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	0
Orthoclase	0.68	9.93	0.84	0.8	1.18	0.76	0.38	0.63	0.72	2.16	2.51	6.16	1.64
Albite	18.28	27.93	16.25	15.74	13.29	16.5	12.1	15.74	15.06	19.46	16.75	19.38	19.97
Anorthite	27.38	16.9	28.78	29.48	28.3	28.39	32.44	28.99	28.76	24.61	41.68	24.84	23.68
Nepheline	0	1.05	0	0	0	0	0	0	0	0	0	0	0
Leucite	0	0	0	0	10.04	0	0	0	0	0	0	0	0
Diopside	25.1/	19.8	20.7	20.08	19.84	20.12	20.98	19.1	20.22	17.59	14.4/	20.22	13.86
Olivine	ι <u>21.21</u> Λ	U 18 54	24.1 0	20.00 N	0.57	2 0	20.40 N	20.1 N	0	0	0.79	2.84	0
Magnetite	231	2 28	2 25	2.16	2.26	2.31	2.04	2.28	2.34	2.49	1.42	1.93	2.7
limenite	2.81	2.75	2.87	2.56	2.37	3.04	1.92	2.81	2.72	3.63	1.94	3.06	3.93
Apatite	0.3	0.28	0.29	0.3	0.22	0.32	0.21	0.3	0.29	0.41	0.22	0.32	0.4
SUM	98.35	99.47	99.28	99.71	99.51	99.5	99.75	99.08	99.2	98.8	99.45	99.17	98.58

Appendix Di: XRF Data - Malaita Volcanic Group	(OJP lavas)).
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-						Longan	Longana River			Akemar	Akemare River		
	ML322	ML323	ML325	ML329	ML331	ML341	ML344	ML348	ML350	ML362	ML367	ML368	
SiO2	53.12	48.78	48.85	52.32	51.62	50.37	49.91	49.27	50.54	49.86	50.74	50.55	
TiO2	2.080	3.720	3.890	1.130	1.020	1.500	1.750	3.810	1.380	1.500	1.500	1.630	
AI2O3	10.50	10.99	10.39	14.74	15.27	14.45	12.59	11.26	13.30	13.63	13.82	14.12	
Fe2O3(t)	18.33	20.00	20.44	10.09	9.52	12.27	16.14	20.31	13.23	13.68	13.15	13.57	
MnO	0.19	0.33	0.27	0.19	0.15	0.25	0.18	0.26	0.20	0.17	0.22	0.31	
MgO	0.56	3.34	3.67	7.08	7.77	7.66	7.98	4.93	8.69	7.98	7.47	7.98	
CaO	11.52	9.62	9.83	12.16	13.14	11.33	9.79	7.33	10.51	10.65	11.08	9.89	
Na2O	3.43	2.86	2.89	2.73	2.21	1.96	1.82	2.69	1.88	1.88	1.91	1.85	
K20	0.136	0.141	0.105	0.452	0.138	0.142	0.130	0.142	0.305	0.167	0.145	0.205	
P205	100.20	0.185	0.188	0.106	0.084	0.126	100 46	0.259	0.120	0.132	100 17	0.132	
Iotai	100.29	99.90	100.52	101.01	100.92	100.07	100.46	100.25	100.15	99.05	100.17	100.24	
L.O.I.	7.17	1.35	4.68	0.53	1.72	1.21	0.82	0.53	2.89	0.54	0.74	1.21	
Mg #	6.43	27.30	28.76	61.21	64.73	58.40	52.64	35.31	59.63	56.74	56.09	56.94	
Nb	15.7	8.0	11.3	2.9	3.1	5.3	6.1	10.7	4.2	5.7	5.4	4.5	
Zr	298.0	144.9	144.6	52.3	51.9	87.0	114.8	160.0	79.9	94.4	89.7	90.3	
Y	86	46	56	21	20	30	35	51	27	31	29	29	
Sr	158	156	160	134	182	155	123	144	147	119	134	130	
Rb	1.0	0.5	0.6	5.2	2.0	2.8	2.7	0.0	4.6	6.7	2.2	1.8	
Th	0	0	0	0	0	0	0	0	0	0	0	2	
Ga	25	24	21	18	17	19	20	25	17	20	19	20	
Zn	37	75	54	64	51	101	81	84	65	44	94	75	
Ni	0	0	0	42	81	77	93	0	111	75	71	76	
Sc	34	40	42	49	45	41	39	63	39	42	45	49	
V	27	215	251	300	270	341	308	376	287	348	334	367	
Cr	0	2	0	74	1/3	183	193	4/ 70	203	138	148	188	
C0	32	57	49	38	42	51	00 101	13	23 94	57	52 110	52 04	
Ba	54	52	28	40	31	83	30	41	47	88	52	34 24	
19	13	11	9	7	8	9	13	9		9	7	8	
Ce	37	23	32	12	7	17	25	26	15	15	16	12	
Nd	28	20	25	6	6	10	15	18	12	10	13	12	
			o 07	40.04	44.07	0.00	7.10	0.00	0.04		0.01		
A1203/1102	5.05	2.95	2.67	13.04	14.97	9.63	7.19	2.90	9.64	9.09	9.21	0.00	
CaU/AI2U3	10.00	0.00	12.95	18.02	16 7/	16.42	18.82	14 95	19.02	16 56	16.61	20.07	
Z//ND	3 45	3 12	2 56	2 47	2 60	2 87	3 26	3 14	2.96	3.08	3.09	3.07	
Na 8.0	0.66	1.13	1.28	2.39	2.13	1.84	1.82	1.55	2.14	1.88	1.72	1.85	
Fe 8.0	3.57	9.64	10.58	7.25	7.90	10.11	14.01	12.56	12.66	11.87	10.56	11.77	
Si 8.0	50.81	47.34	47.51	52.03	51.55	50.26	49. 9 0	48.32	50.75	49.85	50.58	50.54	
Quartz	7.33	3.09	2.67	0	0.66	1.83	2.25	4.35	1.24	1.56	2.83	2.85	
Corundum	0	0	0	0	0	0	0	0	0	0	0	0	
Orthoclase	0.8	0.83	0.62	2.67	0.82	0.84	0.77	0.84	1.8	0.99	0.86	1.21	
Albite	29.02	24.2	24.46	23.1	18.7	16.59	15.4	22.76	15.91	15.91	16.16	15.65	
Anorthite	12.85	16.73	15.07	26.63	31.34	30.21	25.8	18.23	26.95	28.26	28.71	29.62	
Nepheline	0	0	0	0	0	0	0	0	0	0	0	0	
Leucite	0	0	0	0	0	0	0	0	0	0	0	0	
Diopside	36.81	25.53	27.75	27.03	27.14	20.71	18.04	14.01	20.06	19.57	21	15.35	
Hypersthene	3.8	16.9	16.83	16.43	17.66	23.56	30.31	26.97	21.81	26.66	∠4.05 ∩	28.64	
Olivine	0	0	0	0.12	0	0	0 279	0 3 E	2.20	U 0	0 7 7 7	0	
Magnetite	3.10	3.40 7.07	3.52 7 20	1.74 2.15	1.04	2.12	2.70	7 24	2.20	2.30	2.21	2.34	
Anatite	1.05	0.44	0 45	0.25	0.2	0.3	0.38	0.61	0.28	0.31	0.3	0.31	
SUM	98.78	98.24	98.76	100.1	100.1	99	99.06	98.52	99.01	98.47	99.03	99.07	

			Naa'au	River		Lo' <u>omae R</u> iver	Taeloa River	
	ML369	ML371	ML543	ML544	ML545	ML208	ML12	
SiO2	49.52	51.28	50.24	50.30	50.34	53.27	50.49	
TiO2	1.220	0.900	1.310	1.450	1.740	0.800	1.490	
AI2O3	14.63	17.48	13.54	13.67	13.50	14.61	13.73	
Fe2O3(t)	12.79	8.71	13.14	13.22	13.91	9.13	13.15	
MnO	0.19	0.14	0.25	0.27	0.23	0.15	0.20	
MgO	9.33	6.60	8.89	7.60	6.86	7.04	7.02	
CaO	10.64	13.25	11.10	11.35	10.39	11.30	11.42	
Na2O	1.72	1.90	1.54	1.86	2.17	3.61	1.96	
K2O	0.161	0.115	0.239	0.130	0.776	0.154	0.139	
P2O5	0.116	0.080	0.104	0.111	0.138	0.150	0.126	
Total	100.32	100.46	100.35	99.97	100.04	100.22	99.72	
L.O.I.	3.32	1.52	1.63	0.82	0.87	1.73	0.30	
Mg #	62.12	63.01	60.34	56.38	52.58	63.42	54.55	
Nb	3.7	2.7†	3.8	3.9	4.8	4.5	5.1	
Zr	73.1	49.0	68.5	77.5	92.2	93.1	92.8	
Y	25	18	25	26	32	31	30	
Sr	156	221	152	135	209	137	137	
Rb	2.4	2.5	5.9	2.8	4.9	3.1	2.3	
Th	0	0	0	0	0	3	0	
Ga	18	18	17	19	21	16	19	
Zn	68	51	76	80	5 9	41	84	
Ni	154	74	114	81	60	64	73	
Sc	26	39	44	40	45	40	42	
v	218	224	308	337	377	191	331	
Cr	191	237	184	156	99	118	166	
Co	54	35	54	50	54	35	50	
Cu	73	88	106	114	86	17	108	
Ва	31	35	492	142	167	31	52	
La	10	10	10	10	15	10	10	
Ce	10	14	10	10	15	10	13	
Nu	0		5	12	0	10		
AI203/TIO2	11.99	19.42	10.34	9.43	7.76	18.26	9.21	
CaO/Al2O3	0.73	0.76	0.82	0.83	0.77	0.77	0.83	
Zr/Nb	19.76	18.15	18.03	19.87	19.21	20.69	18.20	
Zr/Y	2.95	2.69	2.80	2.96	2.90	2.97	3.10	
Na 8.0	2.22	1.38	1.88	1.71	1.75	3.26	1.60	
Fe 8.0	13.34	5.25	12.91	10.83 50.19	10.20	0.34	9.01	
51 0.0	49.93	50.65	50.52	50.16	45.55	52.97	50.19	
Quartz	0	2.64	1.66	2.15	0.72	0	2.73	
Corundum	0	0	0	0	0	0	0	
Orthoclase	0.95	0.68	1.41	0.77	4.59	0.91	0.82	
Albite	14.55	16.08	13.03	15.74	18.36	30.55	16.59	
Anorthite	31.72	38.83	29.33	28.57	24.8	23.21	28.26	
Nepheline	0	0	0	U A	0	0	U	
Leucite	10.05	0	0	0	01.40	U 06.09	0	
Diopside	16.65	21.56	20.58	22.29	21.43	∠0.U0 10.79	22.0	
riypersthene	29.81	10.51	28.2	24.01	22.93 0	12./0 0.46	22	
	0.73	1 =	U 7 0 0	0 0 00	21	2.40	0 2 27	
Magnetite	2.21	1.5	2.27	2.20	2.4 3.2	1.57	2.21	
	2.32 0.27	0.10	2.49 0.25	2.10 0.26	0.33	0.36	0.3	
Apaule CIM	0.27	0.19	0.20	0.20	98.85	90 44	98.59	
SOM	33.ZI	33.1	33.22	90.0Z	00.00	00.44	30.33	

Appendix Di: XRF Data - Malaita Volcanic Group (OJP lavas).

	Ma <u>mafua Ri</u> ver	Kw <u>awau Riv</u> er	Kw <u>diafaa Ri</u> ver	
	ML43	ML83	ML178	
SiO2	46.57	45.24	46.46	
TiO2	2.450	2.340	2.320	
Al2O3	11.88	10.89	11.24	
Fe2O3(t)	11.40	10.88	11.26	
MnO	0.13	0.19	0.15	
MgO	6.66	8.28	7.23	
CaO	15.98	18.68	17.64	
Na2O	2.71	2.19	2.21	
K20	0.940	0.620	0.839	
P2O5	1.244	0.686	0.570	
Total	99.97	100.00	99.92	
L.O.I.	3.90	5.48	5.80	
Mg #	56.78	63.12	59.08	
Nb	47.3	52.2	41.7	
Zr	195.3	190.7	171.7	
Y	29	27	25	
Sr	582	662	650	
Rb	17.6	16.2	18.9	
Th	5	7	6	
Ga	22	18	18	
Zn	112	97	96	
Ni	117	202	205	
Sc	19	22	21	
v	271	213	208	
Cr	238	321	320	
Co	44	50	47	
Cu	70	154	80	
Ва	466	641	449	
La	37	51	40	
Ce	79	80	70	
Nd	35	36	39	
AI203/TIO2	4.8	4.7	4.8	
CaO/Al2O3	1.3	1.7	1.6	
Zr/Nb	4.1	3.7	4.1	
Zr/Y	6.9	7.2	6.8	
Na 8.0	2.2	2.3	1.9	
Fe 8.0	7.7	9.5	8.4	
Si 8.0	46.2	45.3	46.2	
Quartz	0	0	0	
Corundum	0	0	0	
Orthoclase	5.56	0.04	4.96	
Albite	10.68	0	3.63	
Anorthite	17.48	18.05	18.27	
Nepheline	6.64	10.04	8.17	
Leucite	0	2.84	0	
Diopside	44.57	57.29	53. 76	
Hypersthene	0	0	0	
Olivine	4.71	2.95	2.55	
Magnetite	1.97	1.88	1.94	
limenite	4.65	4.44	4.41	
Apatite	2.95	1.62	1.35	
SUM	99.19	99.16	99.03	

Appendix Dii: XRF Data - Maramasike Volcanics (NMAS).

_		So	outh and Sm	nall Malaita		
	9536	9538	9960	9971	9980	10317
SiO2	47.22	52.34	52.41	54.68	52.34	51.99
TiO2	3.170	2.710	2.670	3.540	2.480	3.390
A12O3	11.58	12.23	12.41	14.60	11.72	12.67
Fe2O3(t)	11.97	11.10	10.94	9.54	11.05	10.94
MnO	0.14	0.13	0.12	0.09	0.13	0.12
MgO	9.33	8.77	8.29	3.79	9.86	7.18
CaO	11.56	7.99	8.09	6.44	8.11	8.04
Na2O	1.77	2.31	2.30	2.8 9	2.08	2.65
K20	1.64	1.57	1.40	3.35	1.51	2.61
P2O5	0.826	0.626	0.551	1.022	0.565	0.933
Total	99.20	99.78	99.18	99.93	99.85	100.52
L.O.I.	3.98	1.460	0.70	0.45	0.82	1.03
Mg #	63.67	63.983	63.02	47.18	66.74	59.61
Nb	36.6	24.7	21.9	27.7	22.0	36.2
Zr	304.4	267.0	238.2	507.4	268.7	397.9
Y	29	27	26	31	24	33
Sr	662	637	644	942	728	989
Rb	26.7	24.3	24.0	38.8	20.9	37.5
Th	3	4	3	2	3	4
Ga	19	20	21	27	20	24
Zn	113	108	103	93	94	113
Ni	278	229	212	27	222	126
Sc	17	21	18	18	22	19
V	181	162	146	1/3	152	188
Cr	356	356	316	16	393	257
Co	47	45	44	30	4/	43
Cu	40	38	33	24	30	34
Ва	435	363	349	021	405	724
La	42	26	24	4/	29	50
Ce	92	56 24	22	64	07 25	61
NG	44	34	33	04	35	01
AI203/TiO2	3.65	4.513	4.65	4.12	4.73	3.74
CaO/AI2O3	1.00	0.653	0.65	0.44	0.69	0.63
Zr/Nb	8.32	10.81	10.88	18.32	12.21	10.99
Zr/Y	10.42	10.04	9.06	16.21	11.29	12.24
Na 8.0	2.27	2.60	2.41	1.32	2.78	2.35
Fe 8.0	12.16	10.48	9.62	1.71	12.00	8.02
SI 8.0	47.63	52.58	52.50	53.37	52.92	51.74
Quartz	0	3.15	4.29	6.83	2.8	0.9
Corundum	0	0	0	0	0	0
Orthoclase	9.7	9.28	8.27	19.77	8.89	15.43
Albite	14.98	19.55	19.46	24.46	17.6	22.42
Anorthite	18.81	18.36	19.4	16.99	18.2	14.97
Nepheline	0	0	0	0	0	0
Leucite	0	0	0	0	0	0
Diopside	27.25	14.24	14.17	7.34	15.11	15.76
Hypersthene	7.55	25.79	24.45	13.11	28.41	19.72
Olivine	9.98	0	0	0	0	0
Magnetite	2.06	1.91	1.89	1.65	1.91	1.89
Ilmenite	6.02	5.15	5.07	6.72	4./1	6.44
Apatite	1.96	1.48	1.31	2.42	1.34	2.21
SUM	98.3	98.91	98.32	99.28	98.97	99.73

Appendix Diii: XRF Data - Maramasike Volcanics (SYV).

	9267	9279	9291	9483	10245	10252	10267	10268	MK1	МКЗ	MK5	MK6	MK13	MK20
5:02	50.06	40.20	F2 90	10.06	40.06	40.90	50.00	50.57	50.15	50.74	49.40	40.70	51.00	47.01
5102 TiO2	1 670	1 810	1 320	40.20 2 570	49.00	49.00	1 080	1 640	1 150	1 710	40.40	49.72	1 540	47.21
A1203	14 89	18 56	14 44	11 19	19.89	14 44	14 52	13 49	15.41	14 89	19 57	22 16	14 96	17 42
Fe2O3(t)	11 21	8 90	8 90	11.15	7 11	10.11	14.52	13 58	10.28	11 56	7 21	5 58	11.35	5.87
MnO	0.24	0.00	0.00	0.15	0.11	0.16	0.17	0.21	0.17	0.20	0.11	0.00	0.19	0.09
MaQ	6.48	9.03	9 44	11.39	6.57	7 69	9.20	7.09	8.33	6.55	14 26	6.96	7.00	14 13
CaO	12.27	8.14	8.59	9.90	11.17	10.65	11.38	10.90	12.14	10.70	9.50	13.88	10.50	14.72
Na2O	2.72	3.40	3.10	2.87	2.55	3.87	2.32	2.32	2.32	3.06	1.15	1.66	2.82	0.72
K20	0.62	0.25	0.73	0.29	1.02	0.59	0.13	0.18	0.17	0.17	0.19	0.03	0.41	0.02
P2O5	0.211	0.358	0.133	0.781	0.103	0.166	0.089	0.139	0.099	0.193	0.018	0.042	0.143	0.013
Total	100.36	99.99	99.79	99.07	98.43	98.94	100.60	100.12	100.22	99.77	100.65	100.60	100.00	100.36
		~ ^		07		4.0	07		4.0		4.0	10	10	
L.O.I.	4.4	0.9	3.0	2.7	2.0	4.0	2.7	1.0	1.8	0.9	4.8	1.8	1.2	3.3
MG #	56.52	69.52	70.46	68.70	67.51	63.10	64.43	54.00	64.56	56.02	81.64	/3./2	58.10	84.41
Nb	5.9	3.8	2.7	30.1	3.2	3.7	2.5	5.7	1.7†	4†	0.4†	0.8†	2†	0.5†
Zr	103.9	188.6	86.5	188.7	54.3	98.2	52.6	90.5	56. 9	122.3	2	11.4	101.1	0
Y	35	30	29	33	18	32	19	29	26	41	5	10	35	5
Sr	700	465	94	322	169	212	217	145	176	180	233	173	152	182
Rb	21.5	3.5	5. 9	4.0	17.7	9.3	2.5	2.7	3.5	2	2.8	1.4	7.2	0.8
Th	0	0	0	4	0	0	0	0	0	2	0	0	0	0
Ga	19	19	14	18	17	16	15	19	16	20	11	16	18	10
Zn	61	44	177	90	48	78	65	80	60	75	64	29	72	26
NI	57	135	86	216	73	71	136	59	79	49	275	113	56	434
Sc	30	15	28	18	22	41	29	36	40	40	15	25	39	29
v	282	146	244	203	169	262	238	333	312	339	62	166	348	83
Cr	1/1	142	327	347	270	234	243	51	294	1/9	1/3	579	1/3	1251
Co	40	35	33	51	38	38	50	53	51	51	49	29	52	42
Cu	25	31	94	52	09	57	127	130	94	95	05	10	45	10
Ба	10	10	35	04 22	57	43 6	22	37	5	55	20	2	40	12
La	18	22	10	82	8	13	- 6	13	3	10	1	2	11	2
Nd	15	15	9	02 45	6	12	4	13	6	13	1	2	11	1
114	10	10	Ŭ	40	Ŭ		-	••	Ū	10	•	Ū	••	•
AI203/TIO2	8.9	10.3	10.9	4.4	23.4	10.0	13.4	8.2	13.4	8.7	103.0	47.1	9.7	96.8
CaO/AI2O3	0.8	0.4	0.6	0.9	0.6	0.7	0.8	0.8	0.8	0.7	0.5	0.6	0.7	0.8
Zr/Nb	17.6	49.6	32.0	6.3	17.0	26.5	21.0	15.9	33.5	30.6	5.0	14.3	50.6	0.0
Zr/Y	3.0	6.2	3.0	5.7	3.1	3.1	2.7	3.1	2.2	3.0	0.4	1.1	2.9	0.0
Na 8.0	2,2	3.8	3.6	4.1	2.0	3.8	2.8	2.0	2.4	2.5	3.5	1.3	2.5	3.0
Fe 8.0	7.3	8.7	9.2	14.8	3.3	7.9 40.7	11.4 50.7	10.5	9.0	7.7	14.5	2.3	8.2	13.0
51 0.0	49.0	49.7	55.5	49.3	40.0	49.7	50.7	50.5	50.5	50.5	50.4	49.4	50.6	49.1
Quartz	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.37	0.00	0.00	0.00	0.14	0.00	0.00
Corundum	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.24	0.00	0.00	0.00
Orthoclase	3.65	1.47	4.31	1.70	6.01	3.50	0.77	1.09	1.03	1.02	1.11	0.20	2.42	0.14
Albite	23.02	28.77	26.23	24.29	21.58	27.61	19.63	19.63	19.63	25.89	9.73	14.05	23.86	6.09
Anorthite	26.60	34.65	23.33	16.80	39.82	20.28	28.82	25.85	31.12	26.39	47.02	52.92	26.96	44.23
Nepheline	0.00	0.00	0.00	0.00	0.00	2.78	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Diopside	27.13	2.92	14.93	22.34	12.14	25./1	22.08	22.61	23.21	21.00	0.00	12.63	19.92	22.82
Hyperstnene	/.16	13.12	23.76	10.89	5.29	0.00	18.41	22.62	10.69	17.60	27.83	18.22	20.04	7.70
Ulivine	0.27	12.53	2.11	13.43	9.09	1 74	0.07	0.00	4.40	1.19	1 24	0.00	0.01	17.49
Magnetite	1.93	1.03	1.00	∠.∪I ∕I ¤¤	1.23	1.74 275	1.9/ 2.0F	2.34 2.11	1.// 2.19	1.33	1.24	0.90	1.90	1.01
	0.1/ 0.E0	3.44 0.95	2.01 0.00	4.00	1.01	2.10 0.20	2.00	0.11	2.10	0.20	0.00	0.09	2.92 0.24	0.02
Apatite	0.50	0.00	0.32	00.1	0.24 07 90	0.00	0.21	0.00	0.20	0.40	100.04	100 11	0.04	0.03
30M	33.42	33.20	35.04	30.10	31.02	30.07	33.01	30.30	33.33	30.19	100.02	100.11	33.02	33.00

	MK25	MK30	MK40	MK54_	MK56	MK58	MK62	MK66	MK69	MK71	MK92	MK93	MK97	MK99
8102	50 47	40.60	40.02	E1 14	50 E1	E0 21	40.90	50.00	50 70	50.60	E1 00	E1 04	47 71	51 10
5102	1 820	2 000	40.93	1 520	2 030	1 570	49.00	1 180	1 650	1 480	1 580	1 560	47.71	1 120
A1203	14 97	14 35	13 76	13.56	14 29	13 43	13.88	14 42	14 10	13.37	13 79	13.24	14.38	16.54
Fe2O3(t)	12.18	12.63	13.44	12.63	13.11	13.72	13.69	11.57	12.96	13.16	12 59	12.22	12.83	9.68
MnO	0.21	0.21	0.24	0.19	0.21	0.21	0.23	0.20	0.20	0.25	0.21	0.25	0.21	0.16
MgO	5.90	6.83	5.88	7.76	5.26	7.11	7.32	8.45	6.55	8.02	7.75	7.23	6.85	6.58
CaO	11.36	10.19	13.10	10.19	10.21	10.40	10.53	11.87	10.51	9.19	10.95	10.46	13.49	11.75
Na2O	3.06	3.68	2.48	2.86	3.69	3.05	2.25	1.90	2.30	2.90	1.99	3.06	2.28	2.69
K2O	0.18	0.10	0.35	0.17	0.22	0.23	0.21	0.22	0.93	0.81	0.12	0.56	0.16	0.18
P2O5	0.196	0.199	0.146	0.130	0.192	0.135	0.137	0.107	0.180	0.124	0.130	0.135	0.199	0.096
Total	100.35	99.88	99.87	100.15	99.72	100.16	99.68	100.18	100.16	99.93	100.19	100.05	99.89	99.89
	33	0.2	36	14	10	10	28	16	20	1 9	34	31	37	31
L.O.I. Ma #	52 13	54.87	49 59	58.01	47 43	53.81	54 59	62 15	53 19	57.81	58.06	57.09	54.56	60.45
ing "	02.10	04.07	40.00	00.01	11.40	00.01	04.00	02.10	00.10	07.01	00.00	07.00	0	
Nb	4.6†	2.6†	5.7†	3.6†	3†	4.9†	4.9†	3.8†	5.7†	4.4†	4.9†	4.9†	3.3†	3†
Zr	74.8	161	128.9	92.8	150.4	93.7	87.9	66.1	97.4	84.3	83.9	85.2	123.1	52.4
Y	28	48	44	31	47	32	30	24	32	29	28	29	42	19
Sr	424	134	201	136	138	143	234	144	198	165	196	151	236	248
Rb	2.3	0	0	8.2	2.1	1.3	1.5	2.4	16.9	5.1	4.1	6	0	1.5
In	0	0	0	2	0	0	0	10	10	17	10	10	10	10
Ga	16	21	21	17	22	21	20	18	19	71	18	18	10	18
	67	// 66	78	90	03 07	/ I 61	71	104	09 70	/ I 60	03 75	65 77	10	49 52
NI So	28	40	30 /1	09 46	20	01 //1	28	104	41	46	75	28	42	34
v	306	40	380	346	418	348	312	281	331	324	350	326	373	258
Cr	141	209	62	134	60	95	73	196	133	131	114	120	96	25
Co	54	54	52	65	54	59	60	58	57	60	56	57	54	43
Cu	•	•										•••	•	
Ba	46	25	35	33	29	43	78	41	53	56	35	66	50	25
La	6	4	6	6	4	7	6	3	10	5	8	5	6	4
Ce	11	13	16	10	14	21	9	8	21	16	17	15	20	10
Nd	10	14	16	12	17	14	9	8	14	11	11	9	17	9
AI203/TiO2	8.2	6.9	8.8	8.9	7.0	8.6	8.5	12.2	8.5	9.0	8.7	8.5	8.1	14.8
CaO/AI2O3	0.8	0.7	1.0	0.8	0.7	0.8	0.8	0.8	0.7	0.7	0.8	0.8	0.9	0.7
Zr/Nb	16.3	61.9	22.6	25.8	50.1	19.1	17.9	17.4	17.1	19.2	17.1	17.4	37.3	17.5
Zr/Y	2.7	3.3	2.9	3.0	3.2	2.9	2.9	2.7	3.0	2.9	3.0	3.0	3.0	2.7
Na 8.0	2.3	3.2	1.7	2.8	2.7	2.7	2.0	2.1	1.8	2.9	1.9	2.8	1.9	2.2
Fe 8.0	7.4	9.2	8.6	10.5	7.4	10.7	11.0	10.5	9.1	11.5	10.5	9.4	9.4	5.9
Si 8.0	49.8	49.2	48.3	51.1	49.7	50.0	49.6	50.4	50.3	50.6	51.0	51.1	47.4	50.7
Quartz	0.00	0.00	0.00	0.00	0.00	0.00	0.59	0.27	0.60	0.00	3.03	0.00	0.00	0.18
Corundum	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Orthoclase	1.06	0.56	2.07	0.98	1.31	1.34	1.22	1.28	5.51	4.80	0.70	3.32	0.94	1.06
Albite	25.8 9	31.14	20.99	24.20	31.23	25.81	19.04	16.08	19.46	24.54	16.84	25.89	19.29	22.76
Anorthite	26.58	22.36	25.38	23.68	21.78	22.29	27.16	30.18	25.39	21.06	28.34	20.74	28.53	32.53
Nepheline	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Diopside	23.64	22.22	32.16	21.42	23.05	23.53	20.01	22.93	21.21	19.54	20.68	24.94	30.65	20.62
Hypersthene	13.84	4.19	6.09	22.34	9.58	14.34	24.68	23.97	21.08	15.86	24.03	14.64	3.62	17.89
Olivine	2.27	11.71	6.41	1.07	5.09	6.00	0.00	0.00	0.00	7.61	0.00	4.09	9.69	0.00
Magnetite	2.10	2.18	2.32	2.18	2.20	2.3/	2.30	2.00	2.23	2.27	2.17	2.11	2.21	1.6/
	3.40 0.40	3.97 0.47	2.90 0.35	∠.89 ∩ 21	0.00 0.45	2.30	3.10 0.22	2.24 0.25	0.10	2.01 0.20	3.00 0 21	2.90 0.30	3.30 0 47	2.13
Apaule	0.40	0.47 02 20	0.00	00.01	0.40 03 80	0.02	0.32	0.20 QQ 10	0.40 QQ AF	0.29	0.31	0.32	0.47	0.23
SUM	33.31	30.00	30.73	33.00	30.00	30.37	30.43	33.13	33.05	30.00	33.11	33.00	30.19	33.00

MK100 MK105 M	MK106 MK109	MK116 MK118	MK122 MK126	MK143 MK153	MK159 MK161

SiO2	51.80	50.34	49.79	49.97	50.46	48.19	49.97	49.65	57.34	44.77	51.32	49.59
TiO2	1.310	1.770	1.600	1.080	1.560	3.920	1.520	1.720	0.740	2.100	1.270	2.260
AI2O3	13.97	13.36	14.00	14.15	13.44	12.81	14.27	14.31	15.95	14.90	13.33	12.77
Fe2O3(t)	10.67	14.08	13.5 9	11.09	12.99	11.28	12.87	13.32	8.0 9	14.21	11.79	15.27
MnO	0.21	0.22	0.20	0.18	0.23	0.16	0.20	0.21	0.15	0.23	0.23	0.26
MgO	8.39	7.72	7.24	9.05	7.85	5.49	7.56	7.20	3.60	9.70	7.86	7.15
CaO	10.04	9.75	10.93	12.19	10.68	11.86	11.80	10.95	11.30	11.79	10.63	10.23
Na2O	3.09	2.21	2.42	1.49	2.51	2.54	1.88	2.34	2.18	1.48	3.11	1.39
K2O	0.61	0.24	0.09	0.15	0.10	3.05	0.12	0.15	0.74	0.25	0.42	0.24
P2O5	0.114	0.151	0.136	0.089	0.139	0.794	0.134	0.142	0.106	0.237	0.111	0.206
Total	100.19	99.85	100.00	99.45	99.97	100.09	100.34	99.98	100.18	99.66	100.07	99.36
L.O.I.	1.8	2.1	2.3	2.1	1.4	4.5	1.0	0.5	5.8	5.3	3.0	7.1
Mg #	63.87	55.21	54.50	64.72	57.61	52.25	56.91	54.86	50.01	60.55	59.98	51.29
Nb	4.1†	5.8†	5.5†	3.1†	5.1†	83.3†	5.2†	5.6†	0.8†	3.8†	3.8†	7.2†
Zr	72.5	102.9	94.2	52.9	92.8	584.4	88.9	100.8	44.1	117.3	70.4	131.5
Y	25	32	30	20	29	37	29	32	18	48	25	39
Sr	149	130	151	133	144	1020	141	153	653	836	142	88
Bb	54	1.8	0	2	0	33.3	0	0	12.5	32	8.6	1.6
Th	0	0	0	0	0	10	õ	õ	2	0	0	0
Ga	15	21	20	17	18	25	20	22	- 18	15	13	17
Zn	74	71	78	61	81	107	71	92	60	88	70	94
Ni	117	52	72	115	80	72	91	60	6	34	100	44
Sc	42	42	41	39	41	20	33	40	26	36	37	38
V	207	251	247	261	3/8	312	348	380	202	408	285	300
v C-	100	901	120	201	146	226	144	100	20	70	106	112
Cr	199	02	130	200	50	230	50	122	20	72	50	50
0	60	01	01	55	29	40	29	60	32	50	50	59
Cu	07	66	67	100	00	010		47	107	FF	00	50
Ва	87	55 -	5/	133	38	010	24	4/	107	55	23	56
La	4		8	2	10	90 177	17	8	0	10	5	<i>'</i>
Ce		21	14	12	12	177	17	10	9	19	12	24
Na	0	15	13	0	12	00	14	14	0	15	10	10
AI203/TiO2	10.7	7.5	8.8	13.1	8.6	3.3	9.4	8.3	21.6	7.1	10.5	5.7
CaO/AI2O3	0.7	0.7	0.8	0.9	0.8	0.9	0.8	0.8	0.7	0.8	0.8	0.8
Zr/Nb	17.7	17.7	17.1	17.1	18.2	7.0	17.1	18.0	55.1	30.9	18.5	18.3
Zr/Y	2.9	3.3	3.1	2.6	3.2	15.8	3.0	3.2	2.4	2.5	2.8	3.4
Na 8.0	3.2	2.1	2.1	1.9	2.5	1.6	1.7	2.0	0.5	2.1	3.1	1.1
Fe 8.0	9.5	11.9	10.8	10.9	11.0	5.9	10.5	10.4	0.0	14.9	9.9	12.3
Si 8.0	51.9	50.3	49.6	50.3	50.4	47.4	49.8	49.4	56.0	45.3	51.3	49.3
Quartz	0.00	1.67	0.00	1.28	0.02	0.00	1.35	0.00	13.11	0.00	0.00	5.24
Corundum	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Orthoclase	3.59	1.39	0.55	0.90	0.60	18.03	0.69	0.86	4.36	1.45	2.49	1.40
Albite	26.15	18.70	20.48	12.61	21.24	13.73	15.91	19.80	18.45	12.52	26.32	11.76
Anorthite	22.46	25.84	27.06	31.47	25.11	14.54	30.16	28.12	31.56	33.29	21.17	27.91
Nepheline	0.00	0.00	0.00	0.00	0.00	4.20	0.00	0.00	0.00	0.00	0.00	0.00
Diopside	21.73	17.82	21.73	23.19	22.22	32.36	22.68	20.90	19.83	19.45	25.32	17.89
Hypersthene	14.73	27.05	22.77	24.86	24.13	0.00	23.01	23.27	9.14	8.11	12.39	26.44
Olivine	6.04	0.00	0.53	0.00	0.00	5.10	0.00	0.00	0.00	16.63	6.66	0.00
Magnetite	1.84	2.43	2.34	1.91	2.24	1.95	2.22	2.30	1.40	2.45	2.03	2.63
Ilmenite	2.49	3.36	3.04	2.05	2.96	7.45	2.89	3.27	1.41	3.99	2.41	4.29
Apatite	0.27	0.36	0.32	0.21	0.33	1.88	0.32	0.34	0.25	0.56	0.26	0.49
SUM	99.28	98.63	98.83	98.48	98.85	99.25	99.21	98.84	99.50	98.46	99.05	98.06

Appendix Dv: XRF Data - Ulawa Volcanics.

	UL2	UL3	UL4	UL5	UL6	UL7	UL8	UL9	UL11	UL12
	40.07	44.07	44.07	45.00	50.04	50.40	<u> </u>			
5102	42.37	44.97	44.87	45.28	50.61	50.46	50.09	50.58	50.57	50.25
1102	11.000	11.00	11.900	11 50	1.270	1.190	14.00	1.350	1.240	1.200
Ee202(+)	10.66	10.99	0.00	11.02	10.00	10.57	14.00	10.20	14.97	10.11
MnO	0.16	0.15	5.55 0 11	0.17	0.14	0.14	0.14	0.14	0.13	0.14
MaQ	5 30	7 55	5.03	7.04	10.79	0.14	11 10	10.14	11 21	0.14
CaO	23.18	10.23	20.66	18.00	0.63	9.79 11 17	8 70	9.64	8 95	11 20
Na2O	20.10	2.68	20.00	2 79	1 /0	1 55	1 /2	1.54	1 44	1 55
K2O	1.67	1 47	1 14	1 52	0.16	0.12	0.22	0.18	0.24	0.11
P205	0.563	0.512	0.591	0.589	0.10	0.12	0.23	0.10	0.24	0.106
Total	99 90	100.36	100.00	100.04	99.96	100 43	99.43	100 16	100.05	100.38
, etai	00.00	100.00	100.00	100.01	00.00	100.40	00.40	100.10		100100
L.O.I.	11.4	9.3	9.0	7.9	2.1	1.7	2.4	1.9	2.4	1.6
Mg #	52.78	60.94	57.17	58.76	69.41	67.60	68.78	68.15	69.25	66.56
NIS	A1 0+	44 8+	A1 5+	11 2+	2 0+	2 4+	26+	A 1+	2 9+	2.8+
7r	140.0	150.8	141.6	1/0 0	3.0 j 65 1	59.2	5.01 65.5	4.1	63.0	68 7
21 V	20	22	20	20	24	22	200.0	24	23	24
sr	629	677	573	623	126	151	123	131	124	121
Bb	30.0	22.9	40.4	45 5	1.8	11	20	24	17	09
Th	6	6	6	7	2	0	0	0	0	0.0
Ga	16	17	16	, 18	18	17	17	17	17	18
Zn Zn	85	100	98	114	61	61	60	64	61	64
Ni	231	220	332	295	136	144	122	125	125	135
Sc	18	19	17	17	39	35	36	39	37	34
v	164	180	156	165	290	286	280	303	276	282
Cr	303	330	328	320	287	361	251	257	257	284
Co	52	51	59	60	55	56	54	55	54	56
Ва	653	717	612	684	23	26	24	26	24	28
La	32	37	33	36	5	5	5	3	2	6
Се	61	72	63	69	13	11	14	12	10	13
Nd	30	34	29	31	9	9	9	9	10	10
AI203/TIO2	6.23	5.91	6.05	5.91	11.87	12.92	11.75	11.32	12.07	12.09
CaO/AI2O3	2.07	1.74	1.72	1.57	0.64	0.73	0.59	0.63	0.60	0.75
Zr/Nb	3.37	3.37	3.41	3.38	17.13	17.12	18.19	16.88	16.82	18.08
Zr/Y	7.19	6.98	6.98	7.38	2.76	2.59	3.05	2.90	2.81	2.85
Na 8.0	1.98	2.52	1.98	2.44	2.53	2.22	2.61	2.44	2.64	2.17
Fe 8.0	5.02	8.49	5.26	7.98	12.99	11.40	14.27	12.69	14.10	11.57
Si 8.0	41.53	44.83	44.23	44.98	51.47	51.01	51.08	51.33	51.57	50.76
Quartz	0.00	0.00	0.00	0.00	1.94	1.32	1.57	1.97	1./5	1.20
Corundum	0.00	0.00	0.00	0.00	0.97	0.70	1.36	1.05	1.39	0.63
Orthoclase	0.00	0.00	0.00	0.00	12.61	13.12	12.02	13.03	12.19	13.12
Albite	12.28	13.81	16.98	14.42	33.98	34.63	33.33	34.26	33.69	33.96
Anorthite	13.66	12.29	12.61	12.79	0.00	0.00	0.00	0.00	0.00	0.00
Nepheline	7.74	6.80	5.27	7.05	0.00	0.00	0.00	0.00	0.00	0.00
Diopside	36.85	49.91	51.61	52.91	10.63	16.38	7.82	10.46	8.16	17.39
Hypersthene	0.00	0.00	0.00	0.00	34.41	29.06	37.74	33.74	37.36	28.64
Olivine	5.00	4.68	0.00	2.7 9	0.00	0.00	0.00	0.00	0.00	0.00
Magnetite	1.84	1.88	1.72	1.92	1.84	1.82	1.97	1.89	1.93	1.88
Ilmenite	3.42	3.55	3.76	3.70	2.41	2.26	2.39	2.56	2.36	2.37
Apatite	1.33	1.21	1.40	1.40	0.26	0.23	0.25	0.26	0.25	0.25
CS	16.93	5.38	5.86	2.21	0.00	0.00	0.00	0.00	0.00	0.00
SUM	99.05	99.51	99.23	99.19	99.05	99.52	98.44	99.22	99.07	99.43
Appendix Dv: XRF Data - Ulawa Volcanics.

	UL13	UL14	UL15	UL16	UL18
SiO2	50.15	50.52	50.29	50.92	49.41
TiO2	1.290	1.230	1.090	1.260	2.300
AI2O3	15.59	14.88	15.29	15.82	13.12
Fe2O3(t)	10.89	10.71	10.81	10.41	10.15
MnO	0.15	0.13	0.14	0.12	0.12
MgO	9.02	9.86	10.12	9.09	5.95
CaO	11.46	11.43	10.95	10.83	13.48
Na2O	1.63	1.55	1.47	1.66	3.04
K20	0.11	0.11	0.11	0.20	1.90
P2O5	0.111	0.102	0.091	0.103	0.613
Total	100.39	100.51	100.36	100.41	100.08
L.O.I.	1.5	1.5	1.8	1.6	3.5
Ma #	65.06	67.43	67.79	66.25	56.86
			.		
Nb	4†	3.8†	3.1†	3.4†	53.5†
Zr	71.5	65.1	54.4	62.8	181.4
Y	24	22	21	20	25
Sr	129	121	126	133	783
Rb	1	2.2	2	0.5	44.6
Th	0	0	0	0	6
Ga	18	17	16	18	21
Zn	66	63	58	67	93
Ni	144	137	140	141	127
Sc	32	35	35	40	17
v	262	284	264	303	181
v C-	203	204	204	240	101
Ur O-	208	328	239	249	233
	55	54 0 7	50	59	48
Ba	26	27	21	25	834
La	3	3	4	6	43
Ce	11	14	8	12	81
Nd	10	9	8	12	37
AI203/TIO2	12.09	12.10	14.03	12.56	5.70
CaO/AI2O3	0.74	0.77	0.72	0.68	1.03
Zr/Nb	17.88	17.13	17.55	18.47	3,39
7r/N	2 98	2 96	2 60	3 11	7 23
No 8 0	2 01	2.00	2.00	2 07	2 28
50 9 0	10 62	11 66	10 11	10.25	5 15
FE O.U	50.47	51 10	50.05	51 26	18 77
51 0.0	ວປ.4/	51.10	50.95	51.20	40.//
Quartz	1.24	1.27	1.12	2.28	0.00
Corundum	0.63	0.62	0.64	1.21	11.23
Orthoclase	13.79	13.12	12.44	14.05	15.71
Albite	34.91	33.34	34.80	35.11	16.54
Anorthite	0.00	0.00	0.00	0.00	5.43
Nonholino	0.00	0.00	0.00	0.00	0.70
Disposido	17.04	10 40	15 20	14 61	39 10
Diopsiae	17.31	10.40	10.39	14.01	30.19
nyperstnene	20.99	20.37	30.88	21.02	0.00
Olivine	0.00	0.00	0.00	0.00	4.63
Magnetite	1.88	1.85	1.86	1.80	1.75
Ilmenite	2.45	2.34	2.07	2.39	4.37
Apatite	0.26	0.24	0.22	0.24	1.45
CS	0.00	0.00	0.00	0.00	0.00
SUM	99.46	99.60	99.42	99.52	99.30

Appendix E INAA and ICPMS data (including PGEs)

Appendix Ei	-	Leicester INAA data.
Appendix Eii	-	ANU ICPMS data (including PGEs).

					<u>~</u>	heurin		Cester		iala.					
Sam. no.	KW6	KW9	KW10	KW13	KW15	KW17	KW19	TLB1	TLB9	TLB12	TLB13	TLB14	TLB20	TLB21	TLB25
La	5.8	9.2	5.3	4.4	4.6	3.2	3.6	5.0	5.0	3.5	5.2	5.1	3.3	3.3	4.6
Ce	15.4	24.1	13.1	11.5	11.8	7.8	9.7	13.4	13.6	9.4	13.6	12.9	9.9	9.5	11.9
Nd	11.9	17.4	10.4	9.5	9.6	6.4	7.6	10.2	10.6	7.8	10.1	9.6	8.2	5.6	9.4
Sm	3.5	5.2	3.2	2.8	2.9	2.3	2.5	3.1	3.0	2.4	3.2	3.2	2.5	2.3	2.9
Eu	1.40	1.85	1.28	1.13	1.11	0.86	1.06	1.25	1.27	0.97	1.23	1.21	1.02	1.01	1.16
Gd	4.6	6.0	4.1	4.1		2.9		4.8	4.4	3.7	4.5	3.6	2.8	2.7	4.2
Tb	0.90	1.08	0.78	0.74	0.71	0.60	0.55	0.70	0.68	0.60		0.71	0.63	0.60	0.75
Yb	2.95	4.15	2.89	2.53	2.49	2.10	2.13	2.72	2.81	2.23	2.82	2.73	2.38	2.27	2.52
Lu	0.48	0.66	0.41	0.39	0.38	0.32	0.35	0.40	0.45	0.37	0.38	0.41	0.35	0.37	0.40
Та	0.36	0.53	0.34	0.28	0.28	0.21	0.24	0.28	0.30	0.22	0.29	0.31	0.24	0.21	0.28
Th	0.62	0.93	0.58	0.44	0.43	0.31	0.37	0.59	0.53	0.36	0.56	0.61	0.33	0.33	0.47
Hf	2.80	4.05	2.50	2.23	2.26	1.67	1.88	2.33	2.41	1.87	2.36	2.30	1.92	1.85	2.17
Sc	48.4	43.5	50.8	46.4	48.3	47.7	49.5	48.2	48.6	48.7	50.6	48.2	51.2	49.9	48.3
Co	51.3	68.8	53.3	53.0	61.3	53.6	53.8	52.0	53.0	54.3	53.9	53.1	59.2	53.8	55.9
U	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	0.13	b.d.l.	b.d.l.	b.d.l.	0.13	b.d.l.	0.15	b.d.l.	b.d.l.	0.02
Sam. no.	TLB27	TLB31	MB6	AL2	AL11	AL12	AL15	ML219	ML543	ML729	MB2	MB8b			
La	5.7	4.3	3.7	14.9	5.6	4.3	5.2	2.9	4.1	3.0	3.8	0.3			
Ce	14.6	12.8	10.5	13.1	14.6	10.1	13.5	7.9	11.1	8.8	10.7	0.3			
Nd	12.2	10.8	8.4	13.7	11.2	8.1	10.7	6.1	7.5	6.5	8.3				
Sm	3.4	3.1	2.6	3.0	3.4	2.8	3.0	2.0	2.5	2.2	2.3	0.0			
Eu	1.41	1.20	1.08	0.74	1.29	0.85	1.23	0.85	0.97	0.89	1.07	0.19			
Gd	4.7	4.2	3.5	3.6	4.7	3.7	4.3	2.8	3.3		3.5				
Tb	0.80	0.80	0.74	0.42	0.76	0.66	0.75	0.50	0.61	0.55	0.74	0.12			
Yb	3.04	2.81	2.36	1.40	2.85	3.12	2.63	1.88	2.21	1.97	2.34	0.03			
Lu	0.46	0.43	0.41	0.22	0.43	0.50	0.43	0.30	0.41	0.32	0.40	0.01			
Та	0.34	0.30	0.24	0.06	0.34	0.08	0.31	0.18	0.24	0.19	0.24	0.32			
Th	0.60	0.54	0.36	0.37	0.59	0.50	0.56	0.26	0.42	0.33	0.35				
Hf	2.85	2.41	2.11	0.59	2.58	2.47	2.33	1.56	2.05	1.73	2.08	0.01			
Sc	48.9	48.3	48.2	7.0	47.4	26.6	48.8	43.4	45.5	46.8	54.7	0.6			
Co	52.8	53.8	53.7	8.5	52.8	30.8	56.1	60.9	57.0	55.1	33.4	1.3			
U	0.13	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	0.02	b.d.l.	b.d.l.	b.d.l.	0.01	b.d.l.			

Appendix Ei: Leicester INAA data.

	KW6	KW10	KW13	KW17	KW19	TLB9	TLB13	TLB14
Li (ppm)	5.8642	8.8377	4.7602	7.2818	4.6331	4.6189	5.219	6.511
Ве	0.58429	0.57002	0.46756	0.38979	0.38218	0.47728	0.537	0.537
Ρ	551.54	494.69	461.78	356.88	370.83	484.03		
Sc	47.997	48.517	47.345	48.332	47.997	46.883	47.773	48.129
Ti	9785.8	9404.3	8007	6929.6	6746.1	8816.8	9655.1	10629
V	390.69	390.94	358.98	324.07	324.4	356.05	324.83	373.64
Cr	137.24	191.24	284.62	249.92	275.88	144	90.185	108.04
Co	45.282	49.209	48.704	49.72	48.53	50.346	52.859	50.997
Ni	56.832	71.976	96.643	99.574	89.891	77.092	79.08	79.704
Cu	100.89	96.885	137.8	129.47	132.49	118.21	115.03	117.31
Zn	103.04	104.46	88.66	75.313	81.022	94.197	91.432	93.473
Ga	17.795	17.505	16.564	15.352	15.403	17.926	18.152	18.306
Rb	1.8777	0.35287	2.204	1.1595	0.40622	2.2902	2.36	1.43
Sr	139.94	142.57	122.58	114.42	116.39	136.8	132.74	134.3
Y	32.206	28.394	26.407	22.648	23.12	28.059	29.271	29.195
Zr	100.6	91.767	79.875	62.578	63.634	82.759	80.326	81.066
Nb	5.8405	5.3686	4.5338	3.3761	3.4151	4.8359	5.043	5.091
Мо	0.78976	0.64756	0.33984	0.33902	0.36205	0.7458	0.398	0.551
Ag	0.022597	0.024351	0.039375	0.018023	0.026999	0.042395		
Cd	0.082994	0.14126	0.15814	0.063427	0.1071	0.079219		
Sn	0.99897	0.96511	0.95016	0.56646	0.62342	1.0152	0.715	0.797
Sb	0.010545	0.0098708	0.0078595	0.0070321	0.0040485	0.013148	0.005	0.007
Cs	0.013467	0.010036	0.0074363	0.0064744	0.0046814	0.012346	0.027	0.006
Ва	34.445	21.595	22.062	14.979	11.557	38.442	31.535	25.878
La	5.8416	5.5403	4.358	3.2448	3.4217	4.9566	5.227	5.29
Ce	15.351	14.348	11.637	8.7878	9.229	13.115	13.971	13.926
Pr	2.343	2.1588	1.798	1.3884	1.4479	1.9743	2.143	2.142
Nd	11.745	10.784	9.305	7.2201	7.6296	10.052	10.607	10.567
Sm	3.6723	3.3655	2.974	2.3921	2.4904	3.1711	3.238	3.262
Eu	1.3318	1.2409	1.092	0.91427	0.93366	1.1726	1.21	1.216
Tb	0.80488	0.7381	0.668	0.55703	0.57048	0.71038	0.782	0.768
Gd	4.5234	4.1038	3.724	3.0387	3.1985	3.9925	4.262	4.213
Dy	5.1194	4.6113	4.218	3.5674	3.6919	4.5139	4.851	4.721
Но	1.1036	0.98957	0.925	0.77274	0.80972	0.98877	1.059	1.026
Er	3.0854	2.8013	2.544	2.1857	2.2316	2.7287	3.001	2.936
Yb	3.0314	2.8035	2.483	2.1354	2.19	2.6358	2.692	2.687
Lu	0.45208	0.42014	0.372	0.32169	0.3298	0.39556	0.402	0.395
Hf	2.5589	2.3723	2.06	1.6203	1.656	2.1513	2.109	2.138
Та	0.36789	0.34162	0.292	0.21653	0.21807	0.30673	0.335	0.34
W	0.070806	0.044395	0.049	0.03694	0.037672	0.062099		
TI	0.014717	0.014142	0.011	0.0051494	0.010881	0.0094124		
Pb	0.4536	1.2258	0.783	0.22614	0.55032	0.34145	0.303	0.489
Th	0.56915	0.51513	0.373	0.26948	0.28111	0.45265	0.487	0.476
U	0.14084	0.13072	0.101	0.074421	0.075694	0.11358	0.118	0.117
Cu (ppb)		96885	137800	129470	132490		115030	117310
Re		1.45206	0.88108	0.72944	0.748		0.42477	1.172
Au		0.17761	3.1183	2.3911	1.2219		0.04068	0.28641
Pd		0.303	5.56871	1.92457	1.48		0.108	0.123
Pt		0.52	5.86226	3.10062	2.677		0.197	0.179
Rh		0.016	0.27624	0.2418533	0.196		0.007	0.003
Ru		0.0529	0.10398	0.09137	0.08/6		0.0318	0.0072
Ir		0.0294	0.16068	0.084	0.115		0.0056	0.0093
Ni		71976	96643	99574	89891		79080	79704

Appendix Eii: ANU ICPMA data.

Li to U is in ppm, Cu to Ni (including PGEs) is in ppb

	TLB21	TLB25	MB6	AL11	AL15	ML543	ML219
Li (ppm)	6.505	5.922	5.1395	8.2457	6.5085	5.9016	5.2262
Be	0.394	0.493	0.4474	0.45952	0.46107	0.32663	0.30991
Ρ			430.89	532.95	450.74	381.65	317.26
Sc	49.356	45.727	47.783	48.006	46.143	44.616	42.642
Ti	8107.6	9497.8	7303.5	9579.1	8961	7519.9	5890.3
v	340.94	337.54	318.36	367.92	369.38	323.99	267.35
Cr	137.09	97.616	271.84	129.37	184.35	190.12	207.06
Co	52 468	53 688	51 914	49 189	50 161	52 847	55.07
Ni	94 296	90.955	107.84	64.555	82.076	112.78	158.16
Cu	139.6	115.67	144 74	96 676	105.65	108 74	129.94
7n	75 706	94 226	88 388	96,699	105.00	92 138	75 837
211	16.09	94.220	17 001	190.099	17 614	15 907	14 626
Cia Dia	10.28	1 4 9 2	1 011	1 6045	1 0 0 1 0	5 6020	1 020
HD	0.819	1.483	1.011	1.6945	1.8318	5.6929	1.237
Sr	112.13	127.79	115.69	140.98	130.04	123.29	107.77
Y 	22.118	26.697	23.914	29.689	26.908	22.632	19.343
Zr	55.558	72.191	70.255	91.867	79.152	64.735	52.788
Nb	3.223	4.545	3.8967	5.5382	4.7353	3.8752	2.898
Мо	0.323	0.374	0.39633	0.6975	0.36845	0.31833	0.25878
Ag			0.022057	0.024354	0.040104	0.21939	0.022093
Cd			0.091291	0.075225	0.16973	0.72672	0.064842
Sn	0.616	0.748	0.68725	0.83971	1.1249	0.60693	0.54311
Sb	0.002	0.003	0.011831	0.010265	0.013281	0.0083792	0.0074179
Cs	0.002	0.009	0.0056872	0.0079681	0.0094739	0.020926	0.017823
Ва	16.027	22.985	17.425	33.92	27.24	408.05	11.777
La	3.217	4.798	3.561	5.7841	4.9881	3.9599	2.839
Се	8.788	12.643	9.735	15.228	12.904	10.495	7.771
Pr	1.415	1.936	1.518	2.2484	1.9159	1.5714	1.206
Nd	7.162	9.548	7.956	11.288	9.715	8.0031	6.322
Sm	2.351	2.954	2.626	3.5269	3.0309	2.5676	2.099
Eu	0.916	1.108	0.989	1.2798	1.1429	0.97851	0.811
ть	0.577	0.7	0.599	0.77551	0.67777	0.57895	0.489
Gd	3.152	3.871	3.332	4.3743	3.8053	3.2864	2.694
Dv	3.569	4.298	3.829	4.9371	4.3052	3.7489	3.129
, Ho	0 779	0.942	0.836	1.0724	0.94137	0.82178	0.689
Fr	2 233	2 708	2 301	2 9411	2 603	2 2466	1.9
Li Vh	2.200	2,308	2.001	2 8566	2.505	2 1846	1 845
	0.207	0 257	0.339	0 43341	0.37522	0.3203/	0 277
	1 402	1 014	1 909	0.40041	2 047	1 7947	1 20
nî Te	1.493	1.911	1.020	2.38//	2.04/	0.24025	0.197
18	0.215	0.299	0.240	0.00000	0.23302	0.24323	0.167
<u>۷۷</u>			0.046	0.004020	0.000011	0.01010	0.040
	0.000	0 51 4	0.005	0.024/31	0.013546	0.01012	0.005
Pb	0.239	0.514	0.1/5	0.35424	1.2093	0.2288	0.221
Th	0.255	0.427	0.282	0.5307	0.44133	0.3527	0.229
U	0.068	0.104	0.084	0.1305	0.10911	0.089825	0.063
Cu (ppb)	139600	115670	144740		105650	108740	129940
Ке	1.66681	2.95569	1.14012		1.42284	0.912/5	1.323/2
Au	1.2541	0.0594	1.4355		0.11436	0.60262	1.5522
Pd	4.76458	0.101	2.45384		0.414	0.30061	3.82417
Pt	4.48459	0.18	3.23091		0.682	0.48162	3.69317
Rh	0.1672667	0.003	0.1754933		0.027	0.0383467	0.2666667
Ru	0.13464	0.0162	0.06815		0.0838	0.05268	0.16508
Ir	0.12568	0.0086	0.1516		0.0346	0.01807	0.13098

Appendix F Radiogenic isotopes

Sample no.	description		Age (m.y.)	Bb	Sr	87Sr/86Sr (t=0)	87Sr/86Sr (t=age)	Sm	PN	143Nd/144Nd (t=0)	143Nd/144Nd (t=age)	eps Nd
AL15	tholeiite	r c	122 122	2.8ppm	139ppm 25.2ug/g	0.704203 0.704187	0.70410 0.704187	3.01ppm 0.34ug/g	9.78ppm 0.91ug/g	0.512832 0.512860	0.512683 0.512680	4.0 3.9
KW13	tholeiite	r u	122 122	3.7ppm	125ppm 21.3ug/g	0.703857 0.703865	0.70371 0.703865	2.99ppm 0.36ug/g	9.51ppm 0.92ug/g	0.512903 0.512960	0.512751 0.512768	5.3 5.6
KW6	tholeiite	r UL	122 122	2.6ppm	142ppm 26.1ug/g	0.704237 0.704197	0.70415 0.704197	3.69ppm 0.30ug/g	12.05ppm 0.76ug/g	0.512819 0.512875	0.512671 0.512684	3.7 4.0
TLB25	tholeiite	г Г	122 122	2.6ppm	132ppm 22.7ug/g	0.70419 0.704173	0.70409 0.704173	2.93ppm 0.31ug/g	9.43ppm 0.76ug/g	0.512828 0.512857	0.512678 0.512660	3.8 3.5
MB2	gabbro	<u>ب</u>	122	2.56ppm	97.9ppm	0.704411	0.70428	1.88ppm	5.32ppm	0.512825	0.512654	3.4
MB6	tholeiite		122	0.67ppm	106ppm	0.703732	0.70370	1.33ppm	3.36ppm	0.512975	0.512784	5.9
ML219	tholeiite	ب	122	0.6ppm	95.05ppm	0.703804	0.70377	1ppm	2.55ppm	0.512981	0.512792	6.1
MB8B	feldspar megacryst	ب_	122	0.795ppm	179.8ppm	0.703718	0.70370	0.04ppm	0.18ppm	0.512968	0.512861	7.4
ML43	alk basalt	_	44	10.67ppm	582.95ppm	0.70373	0.70370	4.89ppm	18.09ppm	0.512879	0.512832	4.9
ML178	alk basalt		44	8.855ppm	582.65ppm	0.703814	0.70379	3.8ppm	14.01ppm	0.512867	0.512820	4.7
AL12	ash layer	ب	44	8.16ppm	128ppm	0.705456	0.70534	1.99ppm	5.92pppm	0.512990	0.512931	6.8
AL2	mudstone	٦	110	4.35ppm	148.7ppm	0.707207	0.70707	1.89ppm	9.37ppm	0.512374	0.512286	-4.1

Appendix F: Radiogenic isotopes

Sample no.	description		176/177Hf	206/204Pb (t=0)	207/204Pb (t=0)	208/204Pb (t=0)	Pb conc (ppm)	U conc (ppb)	206Pb/204Pb (t=age)	207Pb/204Pb (t=age)
AL15	tholeiite	UL		18.121	15.481	38.122	0.785	110	17.953	15.473
		L		18.400	15.526	38.406	0.176	75	17.884	15.501
KW13	tholeiite	UL		18.455	15.536	38.334	0.838	96	18.316	15.529
		L	0.283240	18.683	15.527	38.472	0.157	67.5	18.160	15.502
KW6	tholeiite	UL		18.298	15.489	38.361	0.485	138	17.954	15.472
		L	0.283042	18.600	15.506	38.523	0.131	81.5	17.844	15.469
TLB25	tholeiite	UL		18.276	15.519	38.320	0.584	103	18.063	15.509
		L	0.283169	18.646	15.488	38.436	0.111	78.5	17.787	15.446
MB2	gabbro	L	0.283048	18.573	15.522	38.606	0.16	78	17.979	15.493
MB6	tholeiite	L	0.283071	18.692	15.575	38.508	0.158	61	18.222	15.552
ML219	tholeiite	L	0.283089	18.643	15.561	38.416	0.143	50	18.218	15.540
MB8B	feldspar megacryst	L		18.366	15.560	38.242	0.032	neg.*		
ML43	alk basalt	L		18.821	15.559	38.854	1.281	461	18.663	15.552
ML178	alk basalt	L		18.823	15.555	38.853	1.285	327	18.711	15.549
AL12	ash layer	L		18.691	15.589	38.493	1.709	152	18.652	15.587
AL2	mudstone	UL		18.573	15.631	38.628	2.42	37	18.556	15.630

*U concentration of MB8A indistinguishable from the blank.

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Appendix G Electron microprobe data

Appendix Gi -	(Olivines
Appendix Gii -	I	Pyroxenes
Appendix Giii -	I	Feldspars
Appendix Giv -	(Oxide minerals
Appendix Gv		Alteration minerals

The cations portions have been recalulated on the basis of;

4 oxygens for olivines
6 oxygens for pyroxenes
32 oxygens for feldspars
4 oxygens for oxide minerals - magnetites
3 oxygens for oxide minerals - ilmenites

22 oxygens for alteration minerals.

Appendix	Gi: El	lectron	micro	probe	data	- olivines.
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slide	Information	SiO2	TiO2	AI2O3	Cr2O3	FeO	MnO	MgO	CaO	Na2O	K20	NiO	Total
ML219	olivine core fresh	38.25	0.01	0.02	0.02	27.14	0.46	36.12	0.30	0.01	0.01	0.03	102.40
ML219	olinive core fresh	37.90	0.00	0.01	0.01	28.15	0.40	35.05	0.32	0.01	0.01	0.13	102.00
ML219	olivine fresh core	38.31	0.00	0.00	0.01	28.25	0.40	35.58	0.30	0.02	0.00	0.07	102.90
ML219	olivine fresh at rim	39.31	0.01	0.01	0.02	28.16	0.37	36.82	0.30	0.02	0.01	0.16	105.20
ML219	fresh centre of olivine	38.06	0.01	0.01	0.02	28.29	0.45	35.20	0.30	0.01	0.01	0.12	102.50
ML219	olivine (bleb)	38.18	0.03	0.00	0.00	28.89	0.43	36.33	0.31	0.01	0.01	0.13	104.30
ML219	olivine large (close to rim)	38.59	0.01	0.01	0.03	24.59	0.41	38.30	0.33	0.02	0.00	0.11	102.40
ML219	olivine larger at rim	38.29	0.01	0.01	0.02	25.31	0.33	37.57	0.32	0.00	0.00	0.13	102.00
ML219	middle of olivine	38.87	0.01	0.03	0.03	24.09	0.35	38.41	0.31	0.00	0.02	0.10	102.20
ML219	middle of olivine	38.86	0.01	0.00	0.00	24.27	0.38	38.59	0.29	0.01	0.00	0.11	102.50
ML219	olivine	38.14	0.01	0.03	0.03	27.89	0.44	35.54	0.28	0.02	0.01	0.08	102.50
ML219	fresh olivine core	38.98	0.00	0.02	0.01	23.85	0.33	38.81	0.33	0.00	0.01	0.11	102.50
ML219	fresh olivine core	38.99	0.01	0.03	0.02	22.31	0.35	39.74	0.31	0.02	0.00	0.15	101.90
ML219	bleb in olivine (darker BSE)	39.78	0.01	0.04	0.02	23.10	0.33	40.34	0.30	0.02	0.01	0.09	104.00
ML219	bleb in olivine (darker BSE)	37.65	0.00	0.05	0.03	24.13	0.40	37.42	0.28	0.04	0.01	0.08	100.10
ML219	brown bleb in olivine	38.95	0.00	0.03	0.03	23.39	0.37	39.26	0.34	0.01	0.00	0.10	102.50
ML219	olivine core	38.70	0.03	0.02	0.03	23.88	0.37	38.71	0.31	0.01	0.02	0.09	102.20
ML219	olivine rim, next to cpx	38.96	0.02	0.02	0.03	23.89	0.34	38.78	0.31	0.01	0.01	0.10	102.50
ML219	olivine	39.02	0.01	0.02	0.03	23.57	0.37	39.14	0.34	0.01	0.02	0.09	102.62
ML219	olivine	38.89	0.01	0.03	0.02	23.47	0.36	39.07	0.35	0.01	0.00	0.13	102.34
ML219	olivine altered rim	38.87	0.03	0.01	0.00	23.65	0.37	39.12	0.31	0.01	0.00	0.14	102.51
ML219	olivine fresh core	38.16	0.02	0.03	0.00	27.81	0.45	36.43	0.30	0.01	0.00	0.10	103.31
ML219	olivine fresh core	37.65	0.00	0.03	0.03	28.17	0.47	35.54	0.28	0.03	0.01	0.09	102.30
ML219	olivine fresh core	37.60	0.01	0.00	0.02	28.29	0.44	35.45	0.30	0.01	0.01	0.04	102.17

Appendix (Gi:	Electron	microp	orobe	data	- olivines.
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slide	information	ο	Si	Ti	AI	Cr	Fe2	Mn	Mg	Ca	Na	к	Ni	Total
ML219	olivine core fresh	4	0.99	0.00	0.00	0.00	0.59	0.01	1.40	0.01	0.00	0.00	0.00	3.01
ML219	olinive core fresh	4	0.99	0.00	0.00	0.00	0.62	0.01	1.37	0.01	0.00	0.00	0.00	3.01
ML219	olivine fresh core	4	0.99	0.00	0.00	0.00	0.61	0.01	1.38	0.01	0.00	0.00	0.00	3.01
ML219	olivine fresh at rim	4	1.00	0.00	0.00	0.00	0.60	0.01	1.39	0.01	0.00	0.00	0.00	3.00
ML219	fresh centre of olivine	4	0.9 9	0.00	0.00	0.00	0.62	0.01	1.37	0.01	0.00	0.00	0.00	3.01
ML219	olivine (bleb)	4	0.98	0.00	0.00	0.00	0.62	0.01	1.39	0.01	0.00	0.00	0.00	3.02
ML219	olivine large (close to rim)	4	0.99	0.00	0.00	0.00	0.53	0.01	1.47	0.01	0.00	0.00	0.00	3.01
ML219	olivine larger at rim	4	0.9 9	0.00	0.00	0.00	0.55	0.01	1.45	0.01	0.00	0.00	0.00	3.01
ML219	middle of olivine	4	1.00	0.00	0.00	0.00	0.52	0.01	1.47	0.01	0.00	0.00	0.00	3.00
ML219	middle of olivine	4	0.99	0.00	0.00	0.00	0.52	0.01	1.47	0.01	0.00	0.00	0.00	3.01
ML219	olivine	4	0.99	0.00	0.00	0.00	0.61	0.01	1.38	0.01	0.00	0.00	0.00	3.01
ML219	fresh olivine core	4	1.00	0.00	0.00	0.00	0.51	0.01	1.48	0.01	0.00	0.00	0.00	3.00
ML219	fresh olivine core	4	1.00	0.00	0.00	0.00	0.48	0.01	1.51	0.01	0.00	0.00	0.00	3.00
ML219	bleb in olivine (darker BSE)	4	1.00	0.00	0.00	0.00	0.48	0.01	1.51	0.01	0.00	0.00	0.00	3.00
ML219	bleb in olivine (darker BSE)	4	0.99	0.00	0.00	0.00	0.53	0.01	1.47	0.01	0.00	0.00	0.00	3.01
ML219	brown bleb in olivine	4	0.99	0.00	0.00	0.00	0.50	0.01	1.49	0.01	0.00	0.00	0.00	3.01
ML219	olivine core	4	0.99	0.00	0.00	0.00	0.51	0.01	1.48	0.01	0.00	0.00	0.00	3.01
ML219	olivine rim, next to cpx	4	1.00	0.00	0.00	0.00	0.51	0.01	1.48	0.01	0.00	0.00	0.00	3.00
ML219	olivine	4	0.99	0.00	0.00	0.00	0.50	0.01	1.49	0.01	0.00	0.00	0.00	3.01
ML219	olivine	4	0.99	0.00	0.00	0.00	0.50	0.01	1.49	0.01	0.00	0.00	0.00	3.01
ML219	olivine altered rim	4	0.99	0.00	0.00	0.00	0.51	0.01	1.49	0.01	0.00	0.00	0.00	3.01
ML219	olivine fresh core	4	0.99	0.00	0.00	0.00	0.60	0.01	1.40	0.01	0.00	0.00	0.00	3.01
ML219	olivine fresh core	4	0.99	0.00	0.00	0.00	0.62	0.01	1.39	0.01	0.00	0.00	0.00	3.01
ML219	olivine fresh core	4	0.99	0.00	0.00	0.00	0.62	0.01	1.39	0.01	0.00	0.00	0.00	3.01

Appendix Gii: Electron microprobe data - pyroxenes.

slide	information	SiO2	TiO2	A12O3 (Cr2O3	FeO	MnO	MgO	CaO	Na2O	К2О	NiO	Total
TLB25	pyroxene forming ophtic texture	51.72	0.63	2.93	0.25	7.64	0.24	17.15	19.61	0.25	0.02	0.04	100.48
TLB25	pyroxene in groundmass	47.93	0.50	2.25	0.01	24.86	0.59	6.08	13.91	0.49	0.05	0.02	96.69
TLB25	fresh cpx forming subopht-text	50.49	0.98	2.19	0.01	16.40	0.43	14.20	15.07	0.24	0.01	0.00	100.02
TLB25	small pyroxene in groundmass	50.29	0.56	1.15	0.03	23.47	0.70	9.71	14.25	0.22	0.01	0.02	100.41
TLB25	pyroxene forming ophtic texture	52.89	0.45	2.52	0.47	6.58	0.19	17.21	20.61	0.22	0.01	0.02	101.17
TLB25	pyroxene forming ophtic texture	51.80	0.68	3.06	0.23	6.87	0.17	16.39	20.72	0.22	0.02	0.02	100.18
TLB25	pyroxene forming ophtic texture	51.57	0.66	2.86	0.08	8.30	0.20	16.17	20.09	0.24	0.02	0.03	100.22
TLB25	pyroxene in groundmass	49.04	0.61	1.46	0.01	22.62	0.58	6.39	18.24	0.29	0.01	0.01	99.26
MB8b	pyroxene groundmass	51.19	0.55	1.20	0.02	20.41	0.56	16.76	8.79	0.09	0.01	0.02	99.60
MB8a	pyroxene forming ophtic texture	51.52	0.59	2.17	0.05	9.32	0.22	16.35	18.74	0.26	0.14	0.02	99.38
MB8a	pyroxene forming ophtic texture	51.29	0.65	2.48	0.04	8.98	0.22	15.87	20.45	0.21	0.00	0.05	100.24
MB8a	pyroxene forming ophtic texture	51.44	0.54	1.31	0.01	12.27	0.36	15.77	17.76	0.20	0.00	0.02	99.68
KW9	pyroxene	49.42	0.70	1.26	0.03	23.14	0.56	11.67	12.81	0.16	0.01	0.00	99.76
KW9	pyroxene centre	49.91	0.60	1.17	0.01	22.31	0.63	12.48	12.61	0.16	0.01	0.02	99.91
KW9	cpx pheno (high order colours)	50.47	0.82	1.94	0.04	13.94	0.36	13.64	18.69	0.25	0.01	0.01	100.17
KW9	small pyroxene (pheno?)	48.86	0.75	1.50	0.04	18.90	0.50	11.56	16.74	0.21	0.01	0.02	99.09
KW9	pyroxene (altered middle)	51.10	0.55	1.26	0.04	17.78	0.48	14.85	13.81	0.18	0.01	0.02	100.08
KW9	pyroxene (altered middle)	49.83	0.62	1.05	0.04	21.13	0.54	11.29	15.28	0.17	0.02	0.00	99.97
KW9	pyroxene enclosed in a feidspal	51.06	0.64	1.23	0.04	19.85	0.52	15.25	11.28	0.16	0.01	0.03	100.07
KW9	pyroxene (nign order colours)	50.32	0.67	1.21	0.00	22.26	0.62	13.11	11.93	0.12	0.02	0.00	100.20
KW7	pyroxene in groundmass	52.30	0.57	2.28	0.23	7.91	0.22	17.36	19.42	0.22	0.00	0.04	100.55
KW7	pyroxene in groundmass	50.17	1.12	3.70	0.06	10.69	0.25	15.24	18.31	0.23	0.00	0.02	99.79
KW7	pyroxene in groundmass	50.46	0.85	3.47	0.05	10.30	0.25	15.42	17.27	0.28	0.03	0.01	98.39
KW7	pyroxene in groundmass	52.43	0.50	1.57	0.04	11.30	0.33	19.02	14.11	0.13	0.00	0.04	99.47
KW7	pyroxene pheno	52.01	0.50	2.42	0.31	7.26	0.17	17.06	19.95	0.24	0.01	0.03	99.96
KW7	pyroxene in groundmass	48.91	1.13	2.94	0.04	14.28	0.33	14.11	10.39	0.26	0.02	0.01	90.42
	pyroxene forming opni. text	52.43	0.40	1.70	0.19	/.24 6.91	0.14	16.99	19.00	0.20	0.01	0.07	99.71 100.47
KW7	pyroxene centre pyroxene in groundmass	50.04	0.54	2.40 3.44	0.42	10.84	0.17	16.24	16.33	0.23	0.00	0.00	98.33
KW6	pyroxene in groundmass	51.16	0.74	3.17	0.14	8.49	0.21	16.39	19.66	0.23	0.00	0.01	100.20
KW6	cpx in adms (high order colours	50.64	0.87	3.07	0.03	9.35	0.20	15.92	18.95	0.28	0.01	0.04	99.36
KW6	cpx in gdms (low order colours)	51.89	0.60	2.68	0.23	7.86	0.16	16.99	19.38	0.22	0.01	0.02	100.04
KW6	large pyroxene pheno	52.26	0.50	2.52	0.32	6.65	0.19	16.84	20.68	0.26	0.02	0.04	100.28
KW6	large pyroxene pheno	51.22	0.70	2.50	0.03	11.99	0.30	16.17	16.32	0.27	0.02	0.02	99.54
KW6	large pyroxene pheno	52.13	0.54	2.56	0.30	6.88	0.22	16.98	20.00	0.22	0.01	0.01	99.85
KW6	smaller cpx (darker body colour	52.96	0.44	2.12	0.37	6.71	0.16	16.99	20.80	0.23	0.01	0.01	100.80
KW6	cpx pheno core	52.66	0.48	2.49	0.34	7.48	0.22	17.44	20.18	0.22	0.00	0.04	101.55
KW6	cpx pheno	52.12	0.49	2.62	0.42	7.29	0.18	17.42	19.53	0.26	0.01	0.01	100.35
KW6	cpx pheno	52.23	0.55	2.65	0.34	7.45	0.25	17.36	19.43	0.22	0.02	0.01	100.51
KW6	cpx pheno	52.14	0.56	2.85	0.38	7.05	0.15	16.79	20.40	0.21	0.01	0.00	100.54
KW6	cpx rim	52.03	0.56	2.58	0.36	6.72	0.19	16.83	21.30	0.21	0.00	0.07	100.85
KW6	cpx very egde (lighter BSE)	50.47	0.93	2.72	0.02	13.18	0.32	14.05	18.46	0.26	0.00	0.05	100.46
KW6	cpx pheno core	51.66	0.49	2.42	0.30	6.73	0.17	16.81	21.08	0.22	0.01	0.01	99.90
KW6	cpx pheno	52.22	0.53	2.49	0.41	0.07	0.10	17.00	21.27	0.22	0.00	0.02	101.03
KW6	cpx pneno	52.29	0.40	2.42	0.33	6.75	0.19	17.00	21.07	0.23	0.01	0.03	100.00
KW6	cpx pheno	52.25	0.55	2.40	0.35	6.45	0.19	17.01	21.10	0.23	0.01	0.02	100.03
KW6	cpx pheno rim (next plag pheno	52.90	0.42	2.00	0.30	16.82	0.19	1/.07	14 51	0.21	0.00	0.02	100.79
r.vv6	chx blieuo um (puður BSE)	50.22	0.30	2.34	0.04	0.02	0.40	1-1.00	19.01	0.20	0.01	0.02	100.12
KW1	cpx (forming subopht texture)	53.07	0.48	2.40	0.36	6.73	0.20	17.46	19.90	0.21	0.00	0.01	100.80
KW1	cpx (forming subopht texture)	52.93	0.52	2.58	0.45	6.75	0.22	17.10	20.61	0.24	0.01	0.02	101.40
KW1	cpx middle	51.81	0.78	3.05	0.06	9.86	0.27	14.06	14.07	0.24	0.03	0.05	100.10
KW1	cpx miaaie/rim	49.90	1.03	2.3/	0.01	17.04	0.40	14.24	14.0/	0.24	0.01	0.02	100.10
KW1		50.22	0.90	1.00	0.00	12 10	0.00	JE 10	17.024	0.22	0.00	0.02	101.20
	opx (enclosing opaques)	J1.44	1.29	2.3U 2 81	0.02	15.40	0.23	12 04	18 70	0.20	0.02	0.00	100.20
KW1	cpx forming ophitic texture	52.82	0.59	2.70	0.46	6.70	0.17	16.71	20.66	0.24	0.02	0.02	101.10

Appendix Gii: Electron microprobe data - pyroxenes.

slide	information	0	SI	Ti	AI	Cr	Fe2	Mn	Mg	Ca	Na	к	Ni	Total
TLB25	pyroxene forming ophtic texture	6	1.90	0.02	0.13	0.01	0.23	0.01	0.94	0.77	0.02	0.00	0.00	4.03
TLB25	pyroxene in groundmass	6	1.96	0.02	0.11	0.00	0.85	0.02	0.37	0.61	0.04	0.00	0.00	3.99
TLB25	fresh cpx forming subopht-text	6	1.92	0.03	0.10	0.00	0.52	0.01	0.80	0.61	0.02	0.00	0.00	4.01
TLB25	small pyroxene in groundmass	6	1.96	0.02	0.05	0.00	0.77	0.02	0.57	0.60	0.02	0.00	0.00	4.00
TLB25	pyroxene forming ophtic texture	6	1.92	0.01	0.11	0.01	0.20	0.01	0.93	0.80	0.02	0.00	0.00	4.01
TLB25	pyroxene forming ophtic texture	6	1.91	0.02	0.13	0.01	0.21	0.01	0.90	0.82	0.02	0.00	0.00	4.01
TLB25	pyroxene forming ophtic texture	6	1.91	0.02	0.12	0.00	0.26	0.01	0.89	0.80	0.02	0.00	0.00	4.02
TLB25	pyroxene in groundmass	6	1.96	0.02	0.07	0.00	0.75	0.02	0.38	0.78	0.02	0.00	0.00	4.00
MB8b	pyroxene groundmass	6	1.95	0.02	0.05	0.00	0.65	0.02	0.95	0.36	0.01	0.00	0.00	4.01
MB8a	pyroxene forming ophtic texture	6	1.92	0.02	0.10	0.00	0.29	0.01	0.91	0.75	0.02	0.01	0.00	4.02
MB8a	pyroxene forming ophtic texture	6	1.91	0.02	0.11	0.00	0.28	0.01	0.88	0.81	0.02	0.00	0.00	4.03
MB8a	pyroxene forming ophtic texture	6	1.94	0.02	0.06	0.00	0.39	0.01	0.89	0.72	0.01	0.00	0.00	4.03
KW9	pyroxene	6	1.93	0.02	0.06	0.00	0.76	0.02	0.68	0.54	0.01	0.00	0.00	4.02
KW9	pyroxene centre	6	1.94	0.02	0.05	0.00	0.73	0.02	0.72	0.53	0.01	0.00	0.00	4.02
KW9	cpx pheno (high order colours)	6	1.91	0.02	0.09	0.00	0.44	0.01	0.77	0.76	0.02	0.00	0.00	4.03
KW9	small pyroxene (pheno?)	6	1.91	0.02	0.07	0.00	0.62	0.02	0.68	0.70	0.02	0.00	0.00	4.04
KW9	pyroxene (altered middle)	6	1.94	0.02	0.06	0.00	0.57	0.02	0.84	0.56	0.01	0.00	0.00	4.02
KW9	pyroxene (altered middle)	6	1.94	0.02	0.05	0.00	0.69	0.02	0.00	0.64	0.01	0.00	0.00	4.02
KW9 KW9	pyroxene enclosed in a leidspar pyroxene (high order colours)	6	1.95 1.94	0.02	0.06	0.00	0.83	0.02	0.87	0.40	0.01	0.00	0.00	4.01
KW7	pyroxene in groundmass	6	1.92	0.02	0.10	0.01	0.24	0.01	0.95	0.76	0.02	0.00	0.00	4.02
KW7	pyroxene in groundmass	6	1.88	0.03	0.16	0.00	0.33	0.01	0.85	0.73	0.02	0.00	0.00	4.02
KW7	pyroxene in groundmass	6	1.90	0.02	0.15	0.00	0.33	0.01	0.87	0.70	0.02	0.00	0.00	4.01
KW7	pyroxene in groundmass	6	1.94	0.01	0.07	0.00	0.35	0.01	1.05	0.56	0.01	0.00	0.00	4.01
KW7	pyroxene pheno	6	1.92	0.01	0.11	0.01	0.22	0.01	0.94	0.79	0.02	0.00	0.00	4.02
KW7	pyroxene in groundmass	6	1.88	0.03	0.13	0.00	0.46	0.01	0.81	0.68	0.02	0.00	0.00	4.03
KW7	pyroxene forming opht. text	6	1.93	0.01	0.08	0.01	0.22	0.00	0.97	0.77	0.01	0.00	0.00	4.02
KW7	pyroxene centre	6	1.91	0.02	0.11	0.01	0.21	0.01	0.92	0.82	0.02	0.00	0.00	4.02
KW7	pyroxene in groundmass	6	1.89	0.02	0.15	0.00	0.34	0.01	0.91	0.66	0.02	0.00	0.00	4.01
KW6	pyroxene in groundmass	6	1.89	0.02	0.14	0.00	0.26	0.01	0.90	0.78	0.02	0.00	0.00	4.02
KW6	cpx in gdms (high order colours	6	1.89	0.02	0.14	0.00	0.29	0.01	0.89	0.76	0.02	0.00	0.00	4.02
KW6	cpx in gdms (low order colours)	6	1.91	0.02	0.12	0.01	0.24	0.01	0.93	0.77	0.02	0.00	0.00	4.02
KW6	large pyroxene pheno	6	1.92	0.01	0.11	0.01	0.20	0.01	0.92	0.81	0.02	0.00	0.00	4.02
KW6	large pyroxene pheno	6	1.92	0.02	0.11	0.00	0.38	0.01	0.90	0.66	0.02	0.00	0.00	4.02
KW6	large pyroxene pheno	6	1.92	0.02	0.11	0.01	0.21	0.01	0.93	0.79	0.02	0.00	0.00	4.01
KW6	smaller cpx (darker body colour	6	1.93	0.01	0.09	0.01	0.20	0.00	0.92	0.81	0.02	0.00	0.00	4.01
KW6	cpx pheno core	6	1.91	0.01	0.11	0.01	0.23	0.01	0.94	0.79	0.02	0.00	0.00	4.02
KW6	cpx pheno	6	1.91	0.01	0.11	0.01	0.22	0.01	0.95	0.77	0.02	0.00	0.00	4.02
KW6	cpx pheno	6	1.91	0.02	0.11	0.01	0.23	0.01	0.95	0.70	0.02	0.00	0.00	4.02
KWO	cpx pheno	6	1.91	0.02	0.12	0.01	0.22	0.00	0.92	0.84	0.02	0.00	0.00	4.01
KWG	cpx very ende (lighter BSE)	6	1.91	0.02	0.12	0.00	0.41	0.01	0.79	0.74	0.02	0.00	0.00	4.02
KW6		6	1.91	0.01	0.11	0.01	0.21	0.01	0.93	0.83	0.02	0.00	0.00	4.03
KW6		6	1.91	0.01	0.11	0.01	0.21	0.00	0.92	0.83	0.02	0.00	0.00	4.02
KW6	cpx pheno	6	1.91	0.01	0.10	0.01	0.21	0.01	0.93	0.83	0.02	0.00	0.00	4.02
KW6	cpx pheno	6	1.91	0.01	0.10	0.01	0.21	0.01	0.93	0.83	0.02	0.00	0.00	4.03
KW6	cpx pheno rim (next plag pheno	6	1.93	0.01	0.09	0.01	0.20	0.01	0.93	0.83	0.02	0.00	0.00	4.01
KW6	cpx pheno rim (bright BSE)	6	1.91	0.03	0.11	0.00	0.53	0.01	0.82	0.59	0.02	0.00	0.00	4.02
KW1	cpx (forming subopht texture)	6	1.93	0.01	0.10	0.01	0.20	0.01	0.95	0.78	0.02	0.00	0.00	4.01
KW1	cpx (forming subopht texture)	6	1.92	0.01	0.11	0.01	0.20	0.01	0.92	0.80	0.02	0.00	0.00	4.01
KW1	cpx middle	6	1.91	0.02	0.13	0.00	0.30	0.01	0.94	0.67	0.02	0.00	0.00	4.01
KW1	cpx middle/rim	6	1.90	0.03	0.11	0.00	0.54	0.01	0.81	0.61	0.02	0.00	0.00	4.03
KW1	cpx rim	6	1.94	0.03	0.08	0.00	0.72	0.02	0.53	0.67	0.02	0.00	0.00	4.00
KW1	cpx (enclosing opaques)	6	1.91	0.03	0.13	0.00	0.42	0.01	0.84	0.68	0.02	0.00	0.00	4.02
KW1	cpx in groundmass	D C	1.89	0.04	0.13	0.00	0.40	0.01	0.09	0.77	0.02	0.00	0.00	4.02
KW1	cpx forming opnitic texture	0	1.92	0.02	0.12	0.01	0.20	0.01	0.01	0.01	0.02	0.00	0.00	<u>, 101</u>

Ap	pendix	Gii: E	lectron	micropro	be c	lata -	pyroxenes.
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slide	information	SiO2	TiO2	AI2O3	Cr2O3	FeO	MnO	MgO	CaO	Na2O	K2O	NIO	Total
KW14	pyroxene altered in groundmas	49.77	1.11	7 40	0.11	9 13	0.18	14 40	14 82	1.07	0.04	0.05	98.08
KW14	pyroxene altered in groundmas	49.04	1 21	5 16	0.07	12 13	0.28	12.03	18 69	0.84	0.03	0.01	99.49
KW14	pyroxene altered in groundmas:	48.61	1.43	4.60	0.11	12.35	0.30	14.98	17.06	0.26	0.01	0.02	99.73
MB6	pyroxene in groundmass	49.21	1.32	4.95	0.28	8.71	0.19	14.46	20.19	0.26	0.00	0.05	9 9.62
MB6	pyroxene in groundmass	49.59	1.23	2.53	0.00	15.24	0.40	13.45	17.02	0.23	0.01	0.05	99.75
MB6	cpx (next to opaque)	48.83	1.36	2.78	0.01	17.73	0.39	11.67	16.69	0.22	0.01	0.00	99.69
MB6	pyroxene in groundmass	49.29	1.35	4.70	0.35	8.09	0.16	14.88	20.24	0.24	0.03	0.02	99.35
MB6	pyroxene in groundmass	49.50	1.12	5.39	0.19	7.84	0.17	14.23	20.88	0.30	0.02	0.02	99.66
MB6	cpx (forming subopht texture)	51.63	0.63	2.66	0.10	10.18	0.29	15.66	18.91	0.22	0.01	0.03	100.32
MB6	cpx (forming subopht texture)	51.10	0.91	2.78	0.02	11.53	0.30	15.15	18.43	0.22	0.01	0.00	100.45
MB2	pyroxene	51.38	0.53	1.93	0.04	8.32	0.29	16.10	19.75	0.19	0.00	0.02	98.55
MB2	pyroxene	52.12	0.47	1.85	0.03	8.53	0.24	16.34	20.12	0.25	0.01	0.01	99.97
MB2	pyroxene	52.19	0.52	1.88	0.04	9.17	0.23	16.29	19.16	0.22	0.01	0.02	99.73
MB2	pyroxene	52.13	0.53	1.85	0.02	8.84	0.22	16.31	19.57	0.20	0.00	0.02	99.69
MB2	pyroxene	52.13	0.53	1.77	0.03	9.43	0.27	16.42	19.32	0.21	0.01	0.02	100.14
MB2	pyroxene	52.29	0.43	1.82	0.04	8.09	0.22	16.38	20.04	0.19	0.01	0.04	99.55
MB2	pyroxene	50.13	0.03	0.19	0.00	19.29	0.48	7.78	20.72	0.25	0.01	0.01	98.89
KW18	pyroxene centre	53.11	0.43	1.56	0.01	12.21	0.25	19.35	14.22	0.19	0.01	0.04	101.38
KW18	pyroxene edge	51.49	0.66	2.23	0.01	13.00	0.32	15.46	17.13	0.24	0.01	0.08	100.63
KW18	pyroxene middle	52.21	0.50	2.66	0.08	8.89	0.27	18.41	17.49	0.20	0.00	0.06	100.77
KW18	pyroxene centre	53.14	0.38	1.56	0.08	7.92	0.19	17.89	19.61	0.17	0.01	0.06	101.01
KW18	pyroxene edge	53.30	0.41	1.79	0.06	7.91	0.18	17.76	19.64	0.21	0.00	0.03	101.29
KW13	pyroxene middle	50.76	0.88	3.23	0.19	8.39	0.23	15.01	21.34	0.27	0.01	0.02	100.33
KW13	pyroxene	50.81	0.80	1.91	0.03	14.41	0.35	13.18	18.86	0.23	0.01	0.05	100.64
KW13	pyroxene	51.50	0.76	2.71	0.17	9.36	0.27	15.54	20.11	0.23	0.01	0.03	100.69
KW13	pyroxene	51.21	0.80	3.14	0.13	7.72	0.19	16.32	20.68	0.24	0.00	0.02	100.45
KW13	pyroxene	50.73	0.77	3.35	0.22	6.60	0.17	15.96	21.39	0.20	0.01	0.03	99.43
KW13	pyroxene edge	50.75	0.67	3.57	0.72	6.51	0.17	16.37	20.91	0.22	0.01	0.04	99.94
KW13	cpx at very edge near feldspar	49.28	1.00	2.20	0.00	14.75	0.35	13.11	18.16	0.31	0.01	0.03	99.20
KW13	pyroxene	50.97	0.83	2.80	0.04	10.14	0.20	15.24	19.66	0.25	0.01	0.03	100.17
KW13	pyroxene	50.96	0.94	2.32	0.03	11.38	0.28	14.47	19.61	0.25	0.01	0.03	100.28
KW13	pyroxene	52.58	0.41	1.59	0.05	9.59	0.24	17.55	18.12	0.18	0.02	0.04	100.37
KW13	pyroxene	51.15	0.78	3.02	0.03	8.94	0.20	15.64	20.24	0.23	0.00	0.06	100.29
KW13	pyroxene	51.30	0.80	2.10	0.06	14.41	0.34	13.40	18.96	0.25	0.01	0.03	101.66
ML219	pyroxene (next to olivine)	51.64	0.65	2.14	0.01	9.95	0.22	16.17	18.65	0.21	0.00	0.05	99.69
ML219	pyroxene (next to olivine)	53.18	0.52	2.58	0.07	7.23	0.15	18.02	19.74	0.21	0.01	0.02	101.70
ML219	pyroxene core of above	52.67	0.50	2.73	0.08	6.55	0.18	17.05	20.91	0.18	0.01	0.07	100.90
ML219	pyroxene	53.04	0.44	1.59	0.03	8.61	0.19	16.85	19.58	0.18	0.00	0.02	100.50
ML219	cpx groundmass (light BSE)	47.97	0.37	1.55	0.03	30.93	1.31	8.08	9.00	0.12	0.06	0.02	99.44
ML219	pyroxene	53.48	0.31	1.83	0.29	5.80	0.17	18.01	20.76	0.18	0.01	0.04	100.90
ML219	pyroxene	52.04	0.55	3.26	0.41	5.94	0.11	16.95	21.14	0.20	0.01	0.03	100.60
ML219	pyroxene rim	53.72	0.39	1.70	0.09	6.58	0.16	17.64	20.15	0.17	0.02	0.05	100.70
ML219	cpx (next to rim just probed)	40.20	0.03	6.08	0.03	19.50	0.23	19.24	1.37	0.07	0.37	0.01	87.13
ML219	pyroxene	52.92	0.41	1.72	0.10	6.25	0.15	17.47	20.76	0.22	0.01	0.03	100.00
ML219	pyroxene	51.99	0.49	2.97	0.27	6.24	0.14	17.39	20.07	0.20	0.01	0.04	99.81
ML219	pyroxene	52.59	0.43	2.83	0.31	5.97	0.16	16.92	20.83	0.20	0.00	0.01	100.30
ML219	pyroxene	52.36	0.50	2.32	0.02	7.54	0.18	16.51	20.76	0.21	0.01	0.04	100.50
ML219	pyroxene	52.26	0.60	2.14	0.03	9.34	0.24	15.89	19.51	0.22	0.01	0.00	100.20
ML219	cpx in groundmass	48.45	1.71	2.51	0.02	20.64	0.44	8.88	17.28	0.24	0.04	0.02	100.20
ML219	cpx in groundmass	49.19	1.29	2.02	0.01	20.77	0.50	7.92	18.78	0.24	0.02	0.01	100.80
ML219	pyroxene(?)	49.58	0.72	2.20	0.02	28.45	0.80	11.13	8.13	0.12	0.02	0.03	101.20
ML219	cpx in groundmass	49.38	1.33	3.02	0.02	15.09	0.32	13.25	17.15	0.33	0.02	0.04	99.95
ML219	cpx in groundmass	50.25	0.98	2.24	0.02	15.31	0.41	12.97	18.08	0.29	0.02	0.05	100.60
ML219	cpx in groundmass	48.71	0.67	1.46	0.04	26.17	0.71	8.17	13.21	0.23	0.04	0.03	99.44
ML219	cpx in groundmass	48.06	1.69	3.07	0.04	21.74	0.50	6.66	19.47	0.26	0.01	0.01	101.51
ML219	pyroxene	51.44	0.48	0.72	0.02	23.66	0.60	18.51	4.59	0.08	0.00	0.02	100.12

Appendix Gii: Electron microprobe data - pyroxenes.

KW14 pyroxene altered in groundmass 6 1.86 0.03 0.33 0.00 0.29 0.01 0.80 0.59 0.08 0.00 0.40 KW14 pyroxene altered in groundmass 6 1.86 0.03 0.23 0.00 0.38 0.01 0.68 0.76 0.06 0.00 0.40 KW14 pyroxene altered in groundmass 6 1.83 0.04 0.22 0.01 0.84 0.68 0.02 0.00 0.40 MB6 pyroxene in groundmass 6 1.84 0.04 0.12 0.01 0.27 0.01 0.81 0.81 0.02 0.00 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.04 0.11 0.00 0.57 0.01 0.67 0.69 0.02 0.00 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.03 0.24 0.01 0.25 0.00 0.83 0.20 0.00 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.03
KW14 pyroxene altered in groundmass 6 1.86 0.03 0.23 0.00 0.38 0.01 0.68 0.76 0.06 0.00 4.02 KW14 pyroxene altered in groundmass 6 1.83 0.04 0.20 0.00 0.39 0.01 0.84 0.69 0.02 0.00 4.03 MB6 pyroxene in groundmass 6 1.84 0.04 0.22 0.01 0.27 0.01 0.81 0.02 0.00 0.00 4.02 MB6 cpx (next to opaque) 6 1.89 0.04 0.11 0.00 0.49 0.01 0.77 0.70 0.02 0.00 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.03 0.24 0.01 0.79 0.83 0.02 0.00 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.03 0.24 0.01 0.79 0.83 0.02 0.00 0.00 4.02 MB6 pyroxene forming subopht texture) 6 1.92 0.02 0
KW14 pyroxene altered in groundmass 6 1.83 0.04 0.20 0.00 0.39 0.01 0.84 0.69 0.02 0.00 4.03 MB6 pyroxene in groundmass 6 1.84 0.04 0.22 0.01 0.27 0.01 0.81 0.81 0.02 0.00 4.02 MB6 pyroxene in groundmass 6 1.89 0.04 0.11 0.00 0.49 0.01 0.77 0.70 0.02 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.04 0.21 0.01 0.25 0.00 0.63 0.81 0.02 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.04 0.21 0.01 0.25 0.00 0.83 0.81 0.02 0.00 4.02 MB6 pyroxene in groundmass 6 1.92 0.02 0.12 0.00 0.32 0.01 0.83 0.81 0.02 0.00 4.02 MB6 cpx (forming subopht texture) 6 1.91 0.03 0.12
MB6 pyroxene in groundmass 6 1.84 0.04 0.22 0.01 0.27 0.01 0.81 0.81 0.02 0.00 0.00 4.02 MB6 pyroxene in groundmass 6 1.89 0.04 0.11 0.00 0.49 0.01 0.77 0.70 0.02 0.00 0.00 4.02 MB6 cpx (next to opaque) 6 1.89 0.04 0.13 0.00 0.57 0.01 0.67 0.69 0.02 0.00 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.04 0.21 0.01 0.24 0.01 0.79 0.83 0.02 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.03 0.24 0.01 0.79 0.83 0.02 0.00 4.02 MB6 cpx (forming subopht texture) 6 1.92 0.02 0.12 0.00 0.36 0.01 0.80 0.01 0.00 4.02 MB2 pyroxene 6 1.93 0.02 0.00 0.26
MB6 pyroxene in groundmass 6 1.89 0.04 0.11 0.00 0.49 0.01 0.77 0.70 0.02 0.00 0.00 4.02 MB6 cpx (next to opaque) 6 1.89 0.04 0.13 0.00 0.57 0.01 0.67 0.69 0.02 0.00 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.04 0.21 0.01 0.25 0.00 0.83 0.81 0.02 0.00 4.02 MB6 pyroxene in groundmass 6 1.84 0.03 0.24 0.01 0.24 0.01 0.79 0.83 0.02 0.00 4.01 MB6 cpx (forming subopht texture) 6 1.92 0.02 0.12 0.00 0.36 0.01 0.80 0.01 0.00 4.01 MB2 pyroxene 6 1.93 0.02 0.09 0.00 0.26 0.01 0.80 0.01 0.00 4.02 MB2 pyroxene 6 1.93 0.01 0.08 0.00 0.29 </td
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MB6 pyroxene in groundmass 6 1.84 0.03 0.24 0.01 0.24 0.01 0.79 0.83 0.02 0.00 0.00 4.01 MB6 cpx (forming subopht texture) 6 1.92 0.02 0.12 0.00 0.32 0.01 0.87 0.75 0.02 0.00 0.00 4.01 MB6 cpx (forming subopht texture) 6 1.91 0.03 0.12 0.00 0.32 0.01 0.87 0.75 0.02 0.00 0.00 4.01 MB2 pyroxene 6 1.93 0.02 0.09 0.00 0.26 0.01 0.90 0.80 0.01 0.00 4.02 MB2 pyroxene 6 1.93 0.01 0.08 0.00 0.26 0.01 0.90 0.80 0.02 0.00 0.00 4.02 MB2 pyroxene 6 1.94 0.01 0.08 0.00 0.29 0.01 0.90 0.76 0.02 0.00 0.00 4.01 MB2 pyroxene 6 1.94<
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MB2 pyroxene 6 1.93 0.02 0.08 0.00 0.29 0.01 0.91 0.77 0.02 0.00 0.00 4.02 MB2 pyroxene 6 1.94 0.01 0.08 0.00 0.25 0.01 0.91 0.91 0.02 0.00 0.00 4.01 MB2 pyroxene 6 1.94 0.01 0.08 0.00 0.25 0.01 0.91 0.91 0.02 0.00 0.00 4.01 MB2 pyroxene 6 1.99 0.00 0.01 0.00 0.64 0.02 0.46 0.88 0.02 0.00 0.00 4.02 KW18 pyroxene centre 6 1.94 0.01 0.07 0.00 0.37 0.01 1.05 0.56 0.01 0.00 4.02 KW18 pyroxene edge 6 1.92 0.02 0.10 0.01 0.86 0.69 0.01 0.00 4.02 KW18 pyroxene middle 6 1.94 0.01 0.07 0.00 0.24 <th< td=""></th<>
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MB2 pyroxene 6 1.99 0.00 0.01 0.00 0.64 0.02 0.46 0.88 0.02 0.00 0.00 4.02 KW18 pyroxene centre 6 1.94 0.01 0.07 0.00 0.37 0.01 1.05 0.56 0.01 0.00 4.02 KW18 pyroxene edge 6 1.92 0.02 0.10 0.01 0.01 0.66 0.69 0.02 0.00 4.02 KW18 pyroxene middle 6 1.91 0.01 0.00 0.41 0.01 0.86 0.69 0.02 0.00 4.02 KW18 pyroxene middle 6 1.91 0.01 0.01 0.01 0.01 0.05 0.01 0.00 0.00 4.02 KW18 pyroxene centre 6 1.94 0.01 0.07 0.00 0.24 0.01 0.07 0.01 0.01 0.00 4.02 KW18 pyroxene centre 6
KW18 pyroxene centre 6 1.94 0.01 0.07 0.00 0.37 0.01 1.05 0.56 0.01 0.00 4.02 KW18 pyroxene edge 6 1.92 0.02 0.10 0.00 0.41 0.01 0.86 0.69 0.02 0.00 4.02 KW18 pyroxene middle 6 1.91 0.01 0.11 0.00 0.27 0.01 1.00 0.69 0.01 0.00 4.02 KW18 pyroxene centre 6 1.94 0.01 0.07 0.00 0.24 0.01 0.97 0.77 0.01 0.00 0.00 4.02
KW18 pyroxene edge 6 1.92 0.02 0.10 0.00 0.41 0.01 0.86 0.69 0.02 0.00 4.02 KW18 pyroxene middle 6 1.91 0.01 0.11 0.00 0.27 0.01 1.00 0.69 0.01 0.00 4.02 KW18 pyroxene centre 6 1.94 0.01 0.07 0.00 0.24 0.01 0.97 0.77 0.01 0.00 0.00 4.02
KW18 pyroxene middle 6 1.91 0.01 0.11 0.00 0.27 0.01 1.00 0.69 0.01 0.00 4.03 KW18 pyroxene centre 6 1.94 0.01 0.07 0.00 0.24 0.01 0.97 0.77 0.01 0.00 4.02
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KW18 pyroxene edge 6 1.94 0.01 0.08 0.00 0.24 0.01 0.96 0.77 0.02 0.00 4.02
KW13 pyroxene middle 6 1.89 0.02 0.14 0.01 0.26 0.01 0.83 0.85 0.02 0.00 4.03
KW13 pyroxene 6 1.92 0.02 0.09 0.00 0.46 0.01 0.74 0.76 0.02 0.00 4.02
KW13 pyroxene 6 1.91 0.02 0.12 0.00 0.29 0.01 0.86 0.80 0.02 0.00 4.02
KW13 pyroxene 6 1.89 0.02 0.14 0.00 0.24 0.01 0.90 0.82 0.02 0.00 0.00 4.03
KW13 pyroxene 6 1.89 0.02 0.15 0.01 0.21 0.01 0.88 0.85 0.01 0.00 4.02
KW13 pyroxene edge 6 1.88 0.02 0.16 0.02 0.20 0.01 0.90 0.83 0.02 0.00 4.02
KW13 cpx at very edge near feldspar 6 1.90 0.03 0.10 0.00 0.47 0.01 0.75 0.75 0.02 0.00 0.00 4.04
KW13 pyroxene 6 1.90 0.02 0.12 0.00 0.32 0.01 0.85 0.79 0.02 0.00 4.02
KW13 pyroxene 6 1.91 0.03 0.10 0.00 0.36 0.01 0.81 0.79 0.02 0.00 4.02
KW13 pyroxene 6 1.94 0.01 0.07 0.00 0.30 0.01 0.96 0.72 0.01 0.00 4.02
KW13 pyroxene 6 1.90 0.02 0.13 0.00 0.28 0.01 0.86 0.80 0.02 0.00 4.02
KW13 pyroxene 6 1.92 0.02 0.09 0.00 0.45 0.01 0.75 0.76 0.02 0.00 4.02
ML219 pyroxene (next to olivine) 6 1.93 0.02 0.09 0.00 0.31 0.01 0.90 0.75 0.02 0.00 0.00 4.02
ML219 pyroxene (next to olivine) 6 1.92 0.01 0.11 0.00 0.22 0.00 0.97 0.76 0.02 0.00 0.00 4.02
ML219 pyroxene core of above 6 1.92 0.01 0.12 0.00 0.20 0.01 0.93 0.82 0.01 0.00 0.00 4.01
ML219 pyroxene 6 1.95 0.01 0.07 0.00 0.26 0.01 0.92 0.77 0.01 0.00 0.00 4.01
ML219 cpx groundmass (light BSE) 6 1.94 0.01 0.07 0.00 1.05 0.05 0.49 0.39 0.01 0.00 0.00 4.01
ML219 pyroxene 6 1.94 0.01 0.08 0.01 0.18 0.01 0.97 0.81 0.01 0.00 0.00 4.01
ML219 pyroxene 6 1.90 0.02 0.14 0.01 0.18 0.00 0.92 0.83 0.01 0.00 0.00 4.02
ML219 pyroxene rim 6 1.95 0.01 0.07 0.00 0.20 0.00 0.96 0.79 0.01 0.00 0.00 4.00
ML219 cpx (next to rim just probed) 6 1.75 0.00 0.31 0.00 0.71 0.01 1.25 0.06 0.01 0.02 0.00 4.11
ML219 pyroxene 6 1.94 0.01 0.07 0.00 0.19 0.00 0.96 0.82 0.02 0.00 0.00 4.02
ML219 pyroxene 6 1.91 0.01 0.13 0.01 0.19 0.00 0.95 0.79 0.01 0.00 0.00 4.02
ML219 pyroxene 6 1.92 0.01 0.12 0.01 0.18 0.00 0.92 0.82 0.01 0.00 0.00 4.01
ML219 pyroxene 6 1.93 0.01 0.10 0.00 0.23 0.01 0.90 0.82 0.02 0.00 0.00 4.02
ML219 pyroxene 6 1.94 0.02 0.09 0.00 0.29 0.01 0.88 0.77 0.02 0.00 0.00 4.01
ML219 cpx in groundmass 6 1.89 0.05 0.12 0.00 0.67 0.01 0.52 0.72 0.02 0.00 0.00 4.01
ML219 cpx in groundmass 6 1.92 0.04 0.09 0.00 0.08 0.02 0.46 0.78 0.02 0.00 0.00 4.01
ML219 pyroxene(?) 6 1.93 0.02 0.10 0.00 0.93 0.03 0.65 0.34 0.01 0.00 0.00 4.00
ML219 cpx in groundmass 6 1.88 0.04 0.14 0.00 0.48 0.01 0.75 0.70 0.02 0.00 0.00 4.03
ML219 cpx in groundmass 6 1.91 0.03 0.10 0.00 0.49 0.01 0.73 0.74 0.02 0.00 0.00 4.03
ML219 cpx in groundmass 6 1.95 0.02 0.07 0.00 0.87 0.02 0.49 0.57 0.02 0.00 0.00 4.01
MI 219 pvroxene 6 1.96 0.01 0.03 0.00 0.75 0.02 1.05 0.19 0.01 0.00 4.02

slide	Information	SiO2	TiO2	AI2O3	Cr2O3	FeO	MnO	MgO	CaO	Na2O	K2O	NiO	Total
KW7	Plag pheno core	50.94	0.06	29.74	0.02	0.67	0.02	0.24	14.33	3.50	0.06	0.02	99.60
KW7	Plag pheno middle	49.97	0.04	30.35	0.02	0.71	0.03	0.21	14.99	3.02	0.04	0.02	99.40
KW7	Plag pheno rim	50.64	0.05	30.01	0.04	0.88	0.03	0.27	14.60	3.29	0.03	0.00	99.84
KW7	Plag pheno core	49.43	0.06	31.20	0.00	0.64	0.03	0.21	15.76	2.70	0.01	0.03	100.07
KW7	Plag pheno	48.54	0.04	31.28	0.01	0.67	0.03	0.20	15.91	2.55	0.02	0.02	99.27
KW7	feldspar in subopht. texture	50.29	0.02	30.31	0.02	0.70	0.02	0.24	14.80	3.18	0.05	0.01	99.64
KW7	feld sticking out of subopht text	50.42	0.03	29.89	0.03	0.62	0.01	0.23	14.66	3.25	0.02	0.02	99.18
KW7	feld (alt) in subopht texture	56.25	0.09	26.38	0.04	1.07	0.01	0.11	10.23	5.63	0.15	0.02	99.98
KW7	feld on rim of pyroxene	54.42	0.14	27.20	0.01	1.08	0.03	0.17	11.51	4.88	0.10	0.02	99.56
KW7	middle of feld pheno in opht tex	49.24	0.01	31.22	0.04	0.68	0.02	0.22	15.80	2.63	0.05	0.01	99.92
KW7	middle/rim of feld pheno above	48.80	0.09	30.67	0.04	0.67	0.00	0.21	15.62	2.69	0.03	0.02	98.84
KW7	plagioclase in groundmass	52.19	0.07	28.62	0.04	0.97	0.03	0.26	13.23	4.10	0.05	0.00	99.56
KW7	plagioclase in groundmass	52.63	0.08	28.28	0.05	1.20	0.01	0.29	12.77	4.12	0.08	0.02	99.53
KW7	plagioclase in groundmass	57.59	0.12	25.54	0.02	0.96	0.03	0.14	9.03	6.25	0.18	0.04	99.90
KW7	core of plagioclase in gdms	53.65	0.09	27.76	0.04	0.97	0.02	0.15	12.03	4.77	0.09	0.02	99.59
KW7	plagioclase in groundmass	52.18	0.11	27.70	0.03	1.37	0.01	0.33	12.64	4.61	0.11	0.00	99.09
KW9	Large zoned feldspar pheno	53.60	0.07	28.02	0.02	0.74	0.02	0.12	12.17	4.77	0.10	0.02	99.65
KW9	Large zoned feldspar pheno	56.21	0.05	26.46	0.04	0.59	0.03	0.08	10.18	5.77	0.14	0.04	99.59
KW9	Large zoned feldspar pheno	58.41	0.04	24.93	0.02	0.74	0.03	0.05	8.15	6.80	0.21	0.02	99.40
KW9	feldspar phenocryst	55.09	0.08	27.05	0.03	0.58	0.04	0.05	10.90	5.42	0.10	0.00	99.34
KW9	Feldspar enclosing olivines	53.65	0.08	27.83	0.02	0.65	0.03	0.11	11.94	4.89	0.10	0.02	99.32
KW9	feldspar enclosing pyroxene	55.02	0.06	27.23	0.03	0.59	0.02	0.07	10. 9 5	5.38	0.12	0.02	99.49
KW6	feld in groundmass	55.72	0.02	27.31	0.04	0.79	0.01	0.12	10.67	5.51	0.13	0.02	100.34
KW6	feld pheno middle	49.56	0.01	31.65	0.04	0.57	0.01	0.21	15.47	2.99	0.03	0.02	100.56
KW6	feld pheno rim	51.70	0.10	28.52	0.01	1.92	0.01	0.88	12.93	3.70	0.09	0.02	99.88
KW6	middle of feld pheno	51.23	0.04	30.16	0.04	0.62	0.01	0.20	14.20	3.60	0.05	0.02	100.17
KW6	rim of feld pheno	49.45	0.05	30.46	0.02	1.27	0.03	0.51	14.68	2.94	0.05	0.02	99.48
KW6	middle of feld	50.67	0.03	30.93	0.04	0.65	0.00	0.20	14.65	3.22	0.05	0.00	100.44
KW6	middle/rim feldspar	49.63	0.02	31.78	0.04	0.61	0.00	0.17	15.62	2.75	0.04	0.02	100.68
KW6	rim feldspar	54.76	0.03	28.45	0.02	0.98	0.00	0.10	11.69	5.11	0.13	0.01	101.28
KW6	feldspar in groundmass	53.01	0.05	29.52	0.02	0.90	0.03	0.19	12.86	4.38	0.08	0.02	101.06
KW6	plag pheno (ocsilltory zoning)	50.08	0.03	30.98	0.02	0.58	0.02	0.20	15.83	2.90	0.04	0.02	100.70
KW6	plag pheno (ocsilltory zoning)	50.70	0.08	28.92	0.04	2.21	0.02	0.76	13.89	3.18	0.09	0.03	99.92
KW6	plag pheno (ocsilltory zoning)	50.55	0.04	30.61	0.02	0.54	0.02	0.24	14.95	3.21	0.05	0.02	100.25
KW6	plag pheno (ocsilltory zoning)	50.66	0.02	30.23	0.00	0.56	0.02	0.20	14.49	3.26	0.08	0.02	99.54
KW6	plag pheno (ocsilltory zoning)	50.97	0.06	30.32	0.04	0.62	0.02	0.22	14.58	3.21	0.06	0.02	100.12
KW6	plag pheno	50.00	0.01	30.69	0.01	0.52	0.03	0.21	15.31	3.01	0.05	0.02	99.86
KW6	plag pheno	48.92	0.02	31.41	0.01	0.62	0.00	0.20	16.03	2.61	0.03	0.01	99.86
KW1	feld in ophitic texture	52.13	0.05	29.94	0.04	0.92	0.02	0.29	13.63	3.63	0.08	0.01	100.70
KW1	feld in ophitic texture	52.21	0.02	29.87	0.02	0.83	0.03	0.25	13.31	3.82	0.08	0.01	100.50
KW1	feld small in middle of cpx	51.24	0.05	30.10	0.01	0.89	0.01	0.31	14.05	3.43	0.07	0.02	100.20
KW1	feld small in middle of cpx	51.12	0.03	30.36	0.00	0.72	0.02	0.23	14.06	3.38	0.06	0.02	100.00
KW1	feld in groundmass	54.01	0.03	28.54	0.04	0.85	0.02	0.14	12.10	4.61	0.09	0.02	100.50
KW1	feld in groundmass	55.36	0.11	27.33	0.04	1.29	0.02	0.22	11.17	5.16	0.14	0.00	100.80
KW1	feld in ophitic texture	50.37	0.06	30.88	0.04	0.66	0.03	0.25	14.85	3.09	0.03	0.02	100.30
KW1	feld in ophitic texture	51.26	0.09	30.25	0.04	0.88	0.03	0.23	13.87	3.55	0.06	0.03	100.30
KW1	feld edge of above	53.52	0.07	28.71	0.04	1.06	0.01	0.24	12.28	4.41	0.07	0.00	100.40
KW13	feldspar	51.24	0.06	30.37	0.02	0.79	0.03	0.22	14.14	3.56	0.07	0.00	100.50
KW13	feld at egde of cpx	48.54	0.07	31.94	0.04	0.67	0.03	0.22	16.16	2.42	0.05	0.02	100.16
KW13	feld in middle of cpx	50.48	0.05	30.81	0.04	0.65	0.06	0.22	15.02	3.38	0.05	0.02	100.78
KW13	feldspar	51.79	0.05	29.82	0.02	1.07	0.03	0.15	13.36	4.03	0.09	0.03	100.44
KW13	teld groundmass	51.04	0.04	28.84	0.00	2.01	0.02	0.56	13.36	3.57	0.15	0.01	99.60
KW13	teld groundmass	54.77	0.69	23.73	0.05	4.00	0.01	0.91	8.96	5.79	0.36	0.00	99.27
KW13	teidspar	51.97	0.10	27.45	0.03	2.58	0.01	0.74	12.17	4.20	0.14	0.02	99.41
KW13	teldspar	51.09	0.10	29.34	0.01	1.37	0.03	0.45	13.51	3.76	0.10	0.01	99.77
KW13	teidspar	49.80	0.07	30.18	0.04	1.73	0.01	0.47	14.54	3.06	0.07	0.02	99.99

slide	information	0	Si	Ті	AI	Cr	Fe2	Mn	Mg	Ca	Na	к	Ni	Total
KW7	Plag pheno core	32	9.35	0.01	6.44	0.00	0.10	0.00	0.07	2.82	1.25	0.01	0.00	20.05
KW7	Plag pheno middle	32	9.21	0.01	6.59	0.00	0.11	0.01	0.06	2.96	1.08	0.01	0.00	20.03
KW7	Plag pheno rim	32	9.29	0.01	6.49	0.01	0.14	0.01	0.07	2.87	1.17	0.01	0.00	20.05
KW7	Plag pheno core	32	9.06	0.01	6.74	0.00	0.10	0.01	0.06	3.10	0.96	0.00	0.00	20.04
KW7	Plag pheno	32	8.98	0.01	6.82	0.00	0.10	0.01	0.06	3.16	0.92	0.01	0.00	20.06
KW7	feldspar in subopht. texture	32	9.24	0.00	6.56	0.00	0.11	0.00	0.07	2.91	1.13	0.01	0.00	20.05
KW7	feld sticking out of subopht text	32	9.30	0.00	6.50	0.00	0.10	0.00	0.06	2.90	1.16	0.01	0.00	20.03
KW7	feld (alt) in subopht texture	32	10.17	0.01	5.62	0.01	0.16	0.00	0.03	1.98	1.97	0.04	0.00	20.01
KW7	feld on rim of pyroxene	32	9.92	0.02	5.85	0.00	0.17	0.01	0.05	2.25	1.73	0.02	0.00	20.01
KW7	middle of feld pheno in opht tex	32	9.05	0.00	6.76	0.01	0.10	0.00	0.06	3.11	0.94	0.01	0.00	20.04
KW7	middle/rim of feld pheno above	32	9.07	0.01	6.72	0.01	0.10	0.00	0.06	3.11	0.97	0.01	0.00	20.05
KW7	plagioclase in groundmass	32	9.57	0.01	6.19	0.01	0.15	0.01	0.07	2.60	1.46	0.01	0.00	20.06
KW7	plagioclase in groundmass	32	9.64	0.01	6.11	0.01	0.18	0.00	0.08	2.51	1.46	0.02	0.00	20.03
KW7	plagioclase in groundmass	32	10.39	0.02	5.43	0.00	0.15	0.01	0.04	1.75	2.19	0.04	0.01	20.00
KW7	core of plagioclase in gdms	32	9.80	0.01	5.98	0.01	0.15	0.00	0.04	2.35	1.69	0.02	0.00	20.05
KW7	plagioclase in groundmass	32	9.64	0.02	6.03	0.00	0.21	0.00	0.09	2.50	1.65	0.03	0.00	20.17
KW9	Large zoned feldspar pheno	32	9.78	0.01	6.02	0.00	0.11	0.00	0.03	2.38	1.69	0.02	0.00	20.06
KW9	Large zoned feldspar pheno	32	10.19	0.01	5.65	0.01	0.09	0.01	0.02	1.98	2.03	0.03	0.01	20.01
KW9	Large zoned feldspar pheno	32	10.55	0.01	5.31	0.00	0.11	0.01	0.01	1.58	2.38	0.05	0.00	20.00
KW9	feldspar phenocryst	32	10.03	0.01	5.81	0.00	0.09	0.01	0.01	2.13	1.91	0.02	0.00	20.02
KW9	Feldspar enclosing olivines	32	9.81	0.01	6.00	0.00	0.10	0.01	0.03	2.34	1.73	0.02	0.00	20.06
KW9	feldspar enclosing pyroxene	32	10.01	0.01	5.84	0.00	0.09	0.00	0.02	2.13	1.90	0.03	0.00	20.03
KW6	feld in groundmass	32	10.04	0.00	5.80	0.01	0.12	0.00	0.03	2.06	1.93	0.03	0.00	20.03
KW6	feld pheno middle	32	9.04	0.00	6.80	0.01	0.09	0.00	0.06	3.02	1.06	0.01	0.00	20.09
KW6	feld pheno rim	32	9.49	0.01	6.17	0.00	0.30	0.00	0.24	2.54	1.32	0.02	0.00	20.09
KW6	middle of feld pheno	32	9.34	0.01	6.48	0.01	0.10	0.00	0.05	2.78	1.27	0.01	0.00	20.05
KW6	rim of feld pheno	32	9.13	0.01	6.63	0.00	0.20	0.01	0.14	2.90	1.05	0.01	0.00	20.08
KW6	middle of feld	32	9.22	0.00	6.64	0.01	0.10	0.00	0.05	2.86	1.14	0.01	0.00	20.03
KW6	middle/rim feldspar	32	9.04	0.00	6.82	0.01	0.09	0.00	0.05	3.05	0.97	0.01	0.00	20.04
KW6	rim feldspar	32	9.82	0.00	6.01	0.00	0.15	0.00	0.03	2.25	1.78	0.03	0.00	20.07
KW6	feldspar in groundmass	32	9.56	0.01	6.27	0.00	0.14	0.01	0.05	2.49	1.53	0.02	0.00	20.07
KW6	plag pheno (ocsilitory zoning)	32	9.13	0.00	6.65	0.00	0.09	0.00	0.05	3.09	1.02	0.01	0.00	20.06
KW6	plag pheno (ocsilitory zoning)	32	9.34	0.01	6.28	0.01	0.34	0.00	0.21	2.74	1.14	0.02	0.00	20.09
KW6	plag pneno (ocsilitory zoning)	32	9.23	0.01	0.59	0.00	0.08	0.00	0.06	2.92	1.14	0.01	0.00	20.05
KW6	plag pheno (ocsilitory zoning)	32	9.30	0.00	0.54	0.00	0.09	0.00	0.05	2.00	1.10	0.02	0.00	20.02
KWO	plag pheno (ocsilitory zoning)	32	9.30	0.01	0.52	0.01	0.10	0.00	0.06	2.00	1.14	0.01	0.00	20.00
KW6	plag pheno	32 32	9.17 9.00	0.00	6.81	0.00	0.08	0.00	0.05	3.16	0.93	0.01	0.00	20.04
KW1	feld in ophitic texture	32	9.44	0.01	6.39	0.01	0.14	0.00	0.08	2.64	1.27	0.02	0.00	20.00
KW1	feld in ophitic texture	32	9.47	0.00	6.39	0.00	0.12	0.00	0.07	2.59	1.34	0.02	0.00	20.01
KW1	feld small in middle of cpx	32	9.35	0.01	6.47	0.00	0.13	0.00	0.08	2.75	1.21	0.02	0.00	20.02
KW1	feld small in middle of cpx	32	9.33	0.00	6.53	0.00	0.11	0.00	0.06	2.75	1.20	0.01	0.00	20.00
KW1	feld in groundmass	32	9.76	0.00	6.08	0.01	0.13	0.00	0.04	2.34	1.62	0.02	0.00	20.00
KW1	feld in groundmass	32	9.97	0.02	5.80	0.01	0.20	0.00	0.06	2.15	1.80	0.03	0.00	20.03
KW1	feld in ophitic texture	32	9.19	0.01	6.64	0.01	0.10	0.00	0.07	2.90	1.09	0.01	0.00	20.03
KW1	feld in ophitic texture	32	9.34	0.01	6.50	0.01	0.13	0.00	0.06	2.71	1.25	0.01	0.00	20.03
KW1	feld edge of above	32	9.69	0.01	6.13	0.01	0.16	0.00	0.06	2.38	1.55	0.02	0.00	20.01
KW13	feldspar	32	9.32	0.01	6.51	0.00	0.12	0.00	0.06	2.76	1.25	0.02	0.00	20.05
KW13	feld at egde of cpx	32	8.91	0.01	6.91	0.01	0.10	0.00	0.06	3.18	0.86	0.01	0.00	20.06
KW13	feld in middle of cpx	32	9.18	0.01	6.61	0.01	0.10	0.01	0.06	2.93	1.19	0.01	0.00	20.11
KW13	feldspar	32	9.42	0.01	6.40	0.00	0.16	0.00	0.04	2.60	1.42	0.02	0.00	20.09
KW13	feld groundmass	32	9.41	0.01	6.27	0.00	0.31	0.00	0.15	2.64	1.28	0.04	0.00	20.11
KW13	feld groundmass	32	10.15	0.10	5.18	0.01	0.62	0.00	0.25	1.78	2.08	0.09	0.00	20.25
KW13	feldspar	32	9.61	0.01	5.98	0.00	0.40	0.00	0.20	2.41	1.50	0.03	0.00	20.16
KW13	feldspar	32	9.38	0.01	6.35	0.00	0.21	0.00	0.12	2.66	1.34	0.02	0.00	20.10
KW13	feldspar	32	9.16	0.01	6.55	0.01	0.27	0.00	0.13	2.87	1.09	0.02	0.00	20.10

Appendix G	iiii: Electron	microprobe	data -	feldspars.
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slide	Information	SiO2	TiO2	AI2O3	Cr2O3	FeO	MnO	MgO	CaO	Na2O	K2O	NiO	Total
KW18	feld centre	51.86	0 02	29.91	0.01	0.82	0.02	0.18	13 59	3 92	0.09	0.01	100.43
KW18	feld centre	53.59	0.09	28.97	0.04	0.85	0.01	0.13	12.69	4.67	0.08	0.00	101.12
KW18	feld centre	52.68	0.05	29.81	0.01	0.72	0.01	0.16	13.26	4.20	0.06	0.01	100.97
KW18	feld gdmass	56.06	0.05	27.09	0.00	0.39	0.03	0.12	7.97	7.57	0.08	0.02	99.38
KW18	feldspar	51.44	0.03	30.46	0.04	0.76	0.01	0.20	14.51	3.71	0.04	0.00	101.20
KW18	feldspar	52.38	0.07	30.04	0.04	0.79	0.01	0.17	13.71	4.04	0.07	0.00	101.32
KW18	feldspar	56.40	0.04	27.38	0.02	0.68	0.01	0.07	10.38	5.85	0.12	0.02	100.97
KW18	groundmass feld ????	62.45	0.01	22.23	0.04	3.61	0.02	0.37	3.81	8.72	0.94	0.02	102.22
KW14	altered feldspar in groundmass	58.33	0.01	24.91	0.01	0.16	0.02	0.00	0.40	12.06	0.19	0.02	96.11
KW14	altered feldspar	49.50	1.37	26.49	0.01	5.02	0.23	0.30	10.96	4.14	1.02	0.02	99.06
TLB25	feldspar in subopht texture	50.46	0.03	29.98	0.00	0.89	0.02	0.25	14.36	3.26	0.04	0.02	99.31
TLB25	feldspar in subopht texture	51.11	0.16	25.28	0.03	2.42	0.04	2.72	14.72	3.15	0.06	0.03	99.72
TLB25	feld pheno in middle/rim	51.84	0.02	29.01	0.04	0.74	0.01	0.16	13.51	3.99	0.07	0.03	99.42
TLB25	feldspar pheno rim	53.26	0.08	27.38	0.06	1.23	0.02	0.32	11.79	4.67	0.11	0.01	98.93
TLB25	feld pheno core	50.99	0.12	26.01	0.03	2.66	0.02	1.15	11.47	3.94	0.07	0.02	96.48
TLB25	feld in groundmass	61.90	0.06	20.71	0.03	1.73	0.02	0.67	4.07	7.09	2.17	0.01	98.46
TLB25	alt dark rim of gdms feld	53.70	0.08	27.86	0.04	1.08	0.01	0.13	11.91	4.73	0.09	0.00	99.63
TLB25	feld in groundmass	55.34	0.09	27.11	0.01	0.81	0.03	0.09	10.95	5.37	0.12	0.02	99.94
TLB25	very small feld in opht text	53.40	0.06	29.12	0.00	0.95	0.01	0.18	12.56	4.44	0.06	0.02	100.80
TLB25	feld enclosed in ophtic texture	50.76	0.07	31.08	0.05	0.60	0.02	0.21	14.79	3.32	0.06	0.02	100.98
TLB25	large pheno feldspar	52.70	0.04	29.51	0.01	0.72	0.03	0.17	13.05	4.27	0.06	0.00	100.56
TLB25	large pheno feldspar-middle	54.60	0.07	28.63	0.02	0.89	0.01	0.09	11.81	4.84	0.09	0.00	101.05
TLB25	large pheno feldspar	53.75	0.04	28.71	0.02	0.85	0.00	0.13	12.12	4.88	0.10	0.03	100.63
TLB25	large pheno feldspar	54.02	0.03	28.56	0.04	0.77	0.03	0.10	11.93	5.00	0.10	0.04	100.62
TLB25	large pheno feldspar	54.42	0.09	28.20	0.00	0.70	0.00	0.07	11.27	5.26	0.09	0.00	100.10
TLB25	large pheno feldspar - v rim	57.25	0.11	25.59	0.04	0.84	0.00	0.18	8.50	6.56	0.42	0.01	99.50
TLB25	feld pheno in sub opth texture	52.02	0.06	29.27	0.01	0.96	0.00	0.22	12.84	4.09	0.05	0.00	99.52
TLB25	dark rim of feld	52.14	0.05	26.46	0.04	2.09	0.05	0.87	10.98	4.96	0.10	0.01	97.75
TLB25	feldspar in groundmass	58.08	0.05	26.15	0.01	0.58	0.01	0.05	8.90	6.51	0.12	0.00	100.46
TLB25	larger feld (middle)	50.30	0.04	30.73	0.04	0.57	0.00	0.15	14.69	3.35	0.05	0.02	99.94
TLB25	larger feld (dark rim)	57.47	0.05	25.13	0.04	0.72	0.01	0.10	8.57	6.11	0.16	0.02	98.38
ML219	feldspar	47.81	0.01	32.02	0.01	0.84	0.04	0.13	16.24	2.11	0.03	0.02	99.26
ML219	feld core	48.07	0.02	32.38	0.04	0.56	0.03	0.21	16.76	2.08	0.02	0.00	100.20
ML219	feld middle	48.80	0.05	31.96	0.02	0.68	0.03	0.20	16.16	2.45	0.04	0.00	100.40
ML219	smaller feld encased in cpx	49.59	0.03	30.84	0.04	0.68	0.03	0.16	15.03	2.85	0.04	0.04	99.33
ML219	small feld encased in cpx	49.30	0.03	31.31	0.02	0.73	0.01	0.19	15.56	2.71	0.05	0.00	99.91
ML219	small feld in cpx (core)	49.84	0.07	30.75	0.03	1.01	0.02	0.49	15.02	2.81	0.06	0.02	100.10
ML219	small feld in cpx (rim)	50.34	0.10	27.12	0.04	2.16	0.02	2.94	15.41	2.75	0.08	0.04	101.00
ML219	large feld edge cpx core	50.66	0.06	30.84	0.01	0.72	0.03	0.17	14.94	3.11	0.05	0.01	100.60
ML219	large feld core at edge of cpx	50.00	0.02	30.65	0.01	0.66	0.00	0.19	14.56	3.12	0.05	0.02	99.28
ML219	large feld rim at edge of cpx	54.25	0.07	24.30	0.02	3.44	0.07	2.54	9.15	4.88	0.17	0.03	98.92
ML219	large feld core/rim at edge of cr	55.96	0.12	27.27	0.01	0.99	0.03	0.11	10.36	5.56	0.10	0.04	100.60
ML219	feldspar	49.08	0.03	30.16	0.04	1.21	0.03	0.50	14.67	2.90	0.03	0.00	98.65
ML219	feldspar	50.00	0.05	30.76	0.01	0.75	0.02	0.23	14.74	2.88	0.05	0.02	99.51
ML219	feld core encased in cpx	48.41	0.04	32.08	0.01	0.57	0.02	0.22	16.38	2.16	0.02	0.03	99.94
ML219	feld rim encased in cpx	49.75	0.04	30.79	0.04	0.70	0.03	0.27	15.20	2.79	0.04	0.01	99.66
ML219	small feld in cpx	48.92	0.05	29.70	0.02	1.89	0.05	1.27	14.54	2.43	0.07	0.02	98.96
ML219	feldspar	49.62	0.02	31.18	0.01	0.62	0.02	0.24	15.33	2.72	0.05	0.02	99.83
ML219	feld (between two ophtic cpxs)	50.77	0.03	30.59	0.01	0.70	0.03	0.23	14.48	3.14	0.04	0.02	100.00
ML219	feld enclosed in cpx	50.00	0.04	31.02	0.03	0.71	0.03	0.23	14.94	2.87	0.04	0.03	99.94
ML219	feld enclosed in cox	50.96	0.04	30.17	0.05	0.66	0.01	0.18	14.38	3.31	0.02	0.02	99.80
ML219	feld enclosed in cox	50.39	0.04	30.52	0.04	0.65	0.01	0.21	14.79	3.05	0.05	0.01	99.76
ML219	feld in groundmass	54.50	0.03	27.54	0.04	1.11	0.02	0.14	11.31	5.00	0.10	0.01	99.80
ML219	feld in groundmass	57.83	0.21	22.69	0.05	3.59	0.05	2.07	7.20	6.78	0.20	0.02	100.70
MB6	feld in groundmass	51.61	0.02	29.79	0.02	0.77	0.01	0.26	14.08	3.54	0.06	0.04	100.20
MB6	feld in groundmass	51.29	0.07	28.61	0.01	0.82	0.03	0.18	13.12	3.71	0.08	0.02	97.94

slide	information	0	Si	Ti	AI	Cr	Fe2	Mn	Mg	Ca	Na	к	Ni	Total
KW18	feld centre	32	9 43	0.00	641	0.00	0 12	0.00	0.05	2 65	1 38	0.02	0.00	20.07
KW18	feld centre	32	9.65	0.01	6.15	0.01	0.13	0.00	0.04	2.45	1.63	0.02	0.00	20.08
KW18	feld centre	32	9.51	0.01	6.34	0.00	0.11	0.00	0.04	2.56	1.47	0.01	0.00	20.05
KW18	feld admass	32	10.16	0.01	5.79	0.00	0.06	0.00	0.03	1.55	2.66	0.02	0.00	20.28
KW18	feldspar	32	9.30	0.00	6.49	0.01	0.12	0.00	0.05	2.81	1.30	0.01	0.00	20.10
KW18	feldspar	32	9.44	0.01	6.38	0.01	0.12	0.00	0.04	2.65	1.41	0.02	0.00	20.07
KW18	feldspar	32	10.09	0.01	5.77	0.00	0.10	0.00	0.02	1.99	2.03	0.03	0.00	20.04
KW18	groundmass feld ????	32	11.04	0.00	4.63	0.01	0.53	0.00	0.10	0.72	2.99	0.21	0.00	20.24
KW14	altered feldspar in groundmass	32	10 78	0 00	5 /3	0.00	0.03	0.00	0.00	0.08	1 32	0.05	0.00	20.69
KW14	altered feldspar	32	9.36	0.20	5.91	0.00	0.79	0.00	0.09	2.22	1.52	0.25	0.00	20.37
	foldenar in subonbt texture	32	0 30	0.00	6 5 1	0.00	0 14	0.00	0.07	2 84	1 17	0.01	0.00	20.03
TLD25	foldenar in subopht texture	32	9.50	0.00	5.54	0.00	0.14	0.00	0.07	2.04	1 1/	0.01	0.00	20.28
TLD25	feld pheno in middle/rim	32	9.50	0.02	6.28	0.00	0.50	0.01	0.73	2.30	1 42	0.02	0.00	20.06
TLD25	feldspar pheno rim	32	9.80	0.00	5.94	0.01	0.10	0.00	0.04	2.00	1.42	0.02	0.00	20.06
TLD25	feld pheno core	32	9.70	0.01	5.83	0.01	0.10	0.00	0.00	2.34	1.07	0.02	0.00	20.11
TLD25	feld in groundmass	32	11 28	0.02	4 45	0.01	0.72	0.00	0.00	0.80	2.51	0.50	0.00	19.99
TLD25	alt dark rim of rdms feld	32	9.80	0.01	5 99	0.00	0.20	0.00	0.10	2.33	1.67	0.02	0.00	20.04
TLD25	feld in aroundmass	32	10.03	0.01	5 79	0.01	0.17	0.00	0.07	2.00	1.89	0.02	0.00	20.02
TLD25	ven small feld in onbt text	32	9.64	0.01	6.20	0.00	0.12	0.00	0.02	2 43	1.55	0.00	0.00	20.04
TI B25	feld enclosed in ophtic texture	32	9.20	0.01	6 64	0.00	0.09	0.00	0.06	2.87	1 17	0.01	0.00	20.06
TLB25	large pheno feldspar	32	9.54	0.01	6.30	0.00	0.11	0.00	0.05	2.53	1.50	0.01	0.00	20.06
TI B25	large phono feldspar-middle	32	9.80	0.01	6.06	0.00	0.13	0.00	0.02	2.27	1.69	0.02	0.00	20.01
TI B25	large pheno feldspar	32	9.72	0.01	6.12	0.00	0.13	0.00	0.04	2.35	1.71	0.02	0.00	20.09
TI B25	large pheno feldspar	32	9.76	0.00	6.08	0.01	0.12	0.01	0.03	2.31	1.75	0.02	0.01	20.08
TI B25	large pheno feldspar	32	9.85	0.01	6.02	0.00	0.11	0.00	0.02	2.19	1.85	0.02	0.00	20.06
TI B25	large pheno feldspar - v rim	32	10.37	0.02	5.46	0.01	0.13	0.00	0.05	1.65	2.30	0.10	0.00	20.08
TI 825	feld pheno in sub opth texture	32	9.52	0.01	6.32	0.00	0.15	0.00	0.06	2.52	1.45	0.01	0.00	20.04
TI B25	dark rim of feld	32	9.76	0.01	5.84	0.01	0.33	0.01	0.24	2.20	1.80	0.02	0.00	20.22
TLB25	feldspar in groundmass	32	10.39	0.01	5.51	0.00	0.09	0.00	0.01	1.71	2.26	0.03	0.00	20.00
TI B25	larger feld (middle)	32	9.21	0.01	6.63	0.01	0.09	0.00	0.04	2.88	1.19	0.01	0.00	20.07
TLB25	larger feld (dark rim)	32	10.48	0.01	5.40	0.01	0.11	0.00	0.03	1.68	2.16	0.04	0.00	19.9 1
ML219	feldspar	32	8.86	0.00	7.00	0.00	0.13	0.01	0.04	3.23	0.76	0.01	0.00	20.02
ML219	feld core	32	8.83	0.00	7.01	0.01	0.09	0.00	0.06	3.30	0.74	0.00	0.00	20.03
ML219	feld middle	32	8.93	0.01	6.90	0.00	0.11	0.00	0.05	3.17	0.87	0.01	0.00	20.05
ML219	smaller feld encased in cpx	32	9.14	0.00	6.70	0.01	0.11	0.00	0.04	2.97	1.02	0.01	0.01	20.01
ML219	small feld encased in cpx	32	9.05	0.00	6.78	0.00	0.11	0.00	0.05	3.06	0.97	0.01	0.00	20.04
ML219	small feld in cpx (core)	32	9.13	0.01	6.64	0.00	0.15	0.00	0.13	2.95	1.00	0.01	0.00	20.04
ML219	small feld in cpx (rim)	32	9.24	0.01	5.87	0.01	0.33	0.00	0.80	3.03	0.98	0.02	0.01	20.30
ML219	large feld edge cpx core	32	9.22	0.01	6.61	0.00	0.11	0.00	0.04	2.91	1.10	0.01	0.00	20.02
ML219	large feld core at edge of cpx	32	9.21	0.00	6.65	0.00	0.10	0.00	0.05	2.87	1.11	0.01	0.00	20.02
ML219	large feld rim at edge of cpx	32	10.04	0.01	5.30	0.00	0.53	0.01	0.70	1.81	1.75	0.04	0.00	20.20
ML219	large feld core/rim at edge of cr	32	10.06	0.02	5.78	0.00	0.15	0.00	0.03	2.00	1.94	0.02	0.01	20.01
ML219	feldspar	32	9.14	0.00	6.62	0.01	0.19	0.00	0.14	2.92	1.05	0.01	0.00	20.07
ML219	feldspar	32	9.19	0.01	6.67	0.00	0.12	0.00	0.0 6	2.90	1.03	0.01	0.00	19.99
ML219	feld core encased in cpx	32	8.90	0.01	6.95	0.00	0.09	0.00	0.06	3.23	0.77	0.00	0.00	20.01
ML219	feld rim encased in cpx	32	9.15	0.01	6.67	0.01	0.11	0.00	0.07	3.00	1.00	0.01	0.00	20.02
ML219	small feld in cpx	32	9.10	0.01	6.52	0.00	0.29	0.01	0.35	2.90	0.88	0.02	0.00	20.08
ML219	feldspar	32	9.10	0.00	6.74	0.00	0.10	0.00	0.07	3.01	0.97	0.01	0.00	20.01
ML219	feld (between two ophtic cpxs)	32	9.27	0.00	6.59	0.00	0.11	0.00	0.06	2.83	1.11	0.01	0.00	19.99
ML219	feld enclosed in cpx	32	9.16	0.01	6.69	0.00	0.11	0.00	0.06	2.93	1.02	0.01	0.00	20.00
ML219	feld enclosed in cpx	32	9.32	0.01	6.51	0.01	0.10	0.00	0.05	2.82	1.17	0.00	0.00	20.00
ML219	feld enclosed in cpx	32	9.24	0.01	6.60	0.01	0.10	0.00	0.06	2.91	1.08	0.01	0.00	20.01
ML219	feld in groundmass	32	9.91	0.00	5.90	0.01	0.17	0.00	0.04	2.20	1.76	0.02	0.00	20.03
ML219	feld in groundmass	32	10.47	0.03	4.84	0.01	0.54	0.01	0.56	1.40	2.38	0.04	0.00	20.29
MB6	feld in groundmass	32	9.41	0.00	6.40	0.00	0.12	0.00	0.07	2.75	1.25	0.01	0.01	20.02
MB6	feld in groundmass	32	9.54	0.01	6.28	0.00	0.13	0.00	0.05	2.61	1.34	0.02	0.00	19.98

slide	information	SiO2	TiO2	AI2O3	Cr2O3	FeO	MnO	MgO	CaO	Na2O	K20	NIO	Total
MB6	feld in groundmass	54.72	0.03	27.66	0.04	1.03	0.01	0.15	11.51	4.82	0.11	0.05	100.13
MB6	feld in groundmass	52.39	0.06	29.34	0.04	1.02	0.03	0.19	13.19	4.00	0.06	0.02	100.34
MB6	feld in groundmass	54.62	0.26	23.15	0.01	3.02	0.02	1.36	8.53	5.29	0.21	0.05	96.52
MB2	feldspar	51.05	0.05	29.50	0.01	0.57	0.02	0.16	13.34	3.82	0.06	0.08	98.6 6
MB2	feldspar	51.82	0.03	29.51	0.02	0.69	0.03	0.14	13.42	3.81	0.07	0.02	99.56
MB2	feldspar	52.00	0.03	29.48	0.04	0.79	0.00	0.24	13.38	3.93	0.06	0.03	99.98
MB8a	feld in groundmass	52.59	0.05	28.57	0.04	0.96	0.03	0.09	12.84	4.41	0.07	0.02	99.67
MB8a	feld (alt?) in subopht cpx	50.92	0.04	29.51	0.00	0.97	0.01	0.10	14.02	3.68	0.04	0.02	99.31
MB8a	feld (alt?) in subopht cpx	50.44	0.01	29.62	0.01	0.83	0.06	0.13	14.16	3.70	0.05	0.00	99.01
MB8a	feldspar phenocryst	54.28	0.06	27.52	0.04	0.84	0.03	0.15	11.59	4.94	0.31	0.01	99.77
MB8a	feldspar	57.75	0.05	25.24	0.03	0.78	0.01	0.02	8.33	6.70	0.40	0.00	99.31
MB8a	feldspar	53.79	0.05	27.53	0.03	1.10	0.02	0.14	10.97	5.21	0.24	0.02	99.10
MB8a	feldspar enclosed in opht text	53.30	0.04	28.09	0.05	0.81	0.04	0.14	12.36	4.75	0.07	0.01	99.66
MB8a	feldspar enclosed in opht text	52.93	0.0 9	28.10	0.04	0.79	0.03	0.11	12.59	4.53	0.09	0.04	99.34
MB8a	feldspar enclosed in opht text	51.73	0.07	29.13	0.04	0.75	0.02	0.13	13.36	4.03	0.06	0.02	99.34
MB8b	felspar-light oscill. zone	46.52	0.02	32.88	0.02	0.43	0.04	0.17	17.55	1.63	0.02	0.01	99.2 9
MB8b	felspar-dark oscill. zone	47.02	0.01	32.21	0.03	0.42	0.02	0.23	17.09	1.93	0.03	0.01	99.00
MB8b	felspar-light oscill. zone	45.73	0.00	33.16	0.04	0.43	0.03	0.19	18.02	1.39	0.03	0.02	99.04
MB8b	felspar-dark oscill. zone	47.04	0.06	32.47	0.02	0.42	0.00	0.22	17.23	1.82	0.03	0.02	99.33
MB8b	felspar-light oscill. zone	46.25	0.03	33.08	0.02	0.47	0.02	0.20	17.94	1.52	0.03	0.02	99.58
MB8b	felspar-dark oscill. zone	46.98	0.00	32.55	0.01	0.40	0.01	0.20	17.56	1.86	0.01	0.00	99.58
MB8b	felspar-light oscill. zone	46.45	0.05	33.18	0.04	0.46	0.02	0.16	17.97	1.47	0.01	0.02	99.83
MB8b	feldspar-light at rim	46.52	0.01	33.02	0.01	0.47	0.04	0.21	17.62	1.59	0.03	0.04	99.56
MB8b	feldspar-regrowth?	47.65	0.02	32.19	0.04	0.48	0.03	0.26	16.67	2.13	0.03	0.02	99.52
MB8b	feldspar-regrowth?	51.14	0.07	29.14	0.00	0.77	0.00	0.14	13.60	3.92	0.05	0.02	98.85
MB8b	feldspar-regrowth?	54.50	0.07	27.60	0.02	0.64	0.03	0.08	11.54	5.13	0.09	0.02	99.72
MB8b	feld in groundmass (rim)	60.29	0.08	24.06	0.03	0.47	0.02	0.03	6.69	7.69	0.26	0.01	99.63
MB8b	feld in groundmass (core)	51.53	0.03	29.51	0.04	0.76	0.02	0.11	13.64	3.94	0.07	0.02	99.67
MB8b	feld in groundmass	51.59	0.14	29.50	0.01	0.94	0.05	0.07	13.65	3.95	0.08	0.02	100.00

slide	information	0	Si	Ti	AI	Cr	Fe2	Mn	Mg	Ca	Na	к	Ni	Total
MB6	feld in groundmass	32	9.91	0.00	5.91	0.01	0.16	0.00	0.04	2.23	1.69	0.03	0.01	19.99
MB6	feld in groundmass	32	9.52	0.01	6.29	0.01	0.15	0.00	0.05	2.57	1.41	0.01	0.00	20.03
MB6	feld in groundmass	32	10.31	0.04	5.15	0.00	0.48	0.00	0.38	1.72	1.94	0.05	0.01	20.08
MB2	feldspar	32	9.43	0.01	6.43	0.00	0.0 9	0.00	0.04	2.64	1.37	0.01	0.01	20.04
MB2	feldspar	32	9.48	0.00	6.37	0.00	0.11	0.00	0.04	2.63	1.35	0.02	0.00	20.01
MB2	feldspar	32	9.48	0.00	6.34	0.01	0.12	0.00	0.06	2.61	1.39	0.01	0.00	20.04
MB8a	feld in groundmass	32	9.62	0.01	6.16	0.01	0.15	0.01	0.03	2.52	1.57	0.02	0.00	20.08
MB8a	feld (alt?) in subopht cpx	32	9.38	0.01	6.41	0.00	0.15	0.00	0.03	2.77	1.32	0.01	0.00	20.07
MB8a	feld (alt?) in subopht cpx	32	9.33	0.00	6.46	0.00	0.13	0.01	0.04	2.81	1.33	0.01	0.00	20.11
MB8a	feldspar phenocryst	32	9.89	0.01	5.91	0.01	0.13	0.01	0.04	2.26	1.74	0.07	0.00	20.06
MB8a	feldspar	32	10.46	0.01	5.39	0.00	0.12	0.00	0.01	1.62	2.35	0.09	0.00	20.05
MB8a	feldspar	32	9.87	0.01	5.95	0.00	0.17	0.00	0.04	2.16	1.85	0.06	0.00	20.11
MB8a	feldspar enclosed in opht text	32	9.74	0.01	6.05	0.01	0.12	0.01	0.04	2.42	1.68	0.02	0.00	20.08
MB8a	feldspar enclosed in opht text	32	9.70	0.01	6.07	0.01	0.12	0.01	0.03	2.47	1.61	0.02	0.01	20.06
MB8a	feldspar enclosed in opht text	32	9.50	0.01	6.31	0.01	0.12	0.00	0.04	2.63	1.44	0.01	0.00	20.06
MB8b	felspar-light oscill. zone	32	8.64	0.00	7.20	0.00	0.07	0.01	0.05	3.49	0.59	0.01	0.00	20.05
MB8b	felspar-dark oscill. zone	32	8.75	0.00	7.07	0.00	0.07	0.00	0.06	3.41	0.70	0.01	0.00	20.06
MB8b	felspar-light oscill. zone	32	8.53	0.00	7.29	0.01	0.07	0 .01	0.05	3.60	0.50	0.01	0.00	20.07
MB8b	felspar-dark oscill. zone	32	8.72	0.01	7.10	0.00	0.07	0.00	0.06	3.42	0.65	0.01	0.00	20.05
MB8b	felspar-light oscill. zone	32	8.58	0.00	7.23	0.00	0.07	0.00	0.06	3.57	0.55	0.01	0.00	20.07
MB8b	felspar-dark oscill. zone	32	8.70	0.00	7.11	0.00	0.06	0.00	0.06	3.49	0.67	0.00	0.00	20.08
MB8b	felspar-light oscill. zone	32	8.59	0.01	7.23	0.01	0.07	0.00	0.04	3.56	0.53	0.00	0.00	20.05
MB8b	feldspar-light at rim	32	8.62	0.00	7.21	0.00	0.07	0.01	0.06	3.50	0.57	0.01	0.01	20.06
MB8b	feldspar-regrowth?	32	8.81	0.00	7.02	0.01	0.07	0.01	0.07	3.30	0.76	0.01	0.00	20.06
MB8b	feldspar-regrowth?	32	9.45	0.01	6.35	0.00	0.12	0.00	0.04	2.69	1.40	0.01	0.00	20.08
MB8b	feldspar-regrowth?	32	9.91	0.01	5.91	0.00	0.10	0.01	0.02	2.25	1.81	0.02	0.00	20.04
MB8b	feld in groundmass (rim)	32	10.81	0.01	5.08	0.00	0.07	0.00	0.01	1.29	2.67	0.06	0.00	20.00
MB8b	feld in groundmass (core)	32	9.44	0.00	6.37	0.01	0.12	0.00	0.03	2.68	1.40	0.02	0.00	20.07
MB8b	feld in groundmass	32	9.43	0.02	6.36	0.00	0.14	0 .0 1	0.02	2.67	1.40	0.02	0.00	20.08

Appendix Giv: Electron microprobe da	ata - oxide mir	ierals.
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slide	information	SiO2	TiO2	AI2O3	Cr2O3	FeO	MnO	MgO	CaO	Na2O	К2О	NiO	Total
KW1	skeletal opaque	0.51	23.70	2.13	0.05	68.70	0.54	0.37	0.12	0.03	0.01	0.03	96.20
KW1	opque at rim	0.17	22.80	1.94	0.01	70.80	0.53	0.29	0.06	0.02	0.00	0.06	96.70
KW1	opaque in cpx	0.57	21.90	2.58	0.05	70.30	0.54	0.47	0.09	0.01	0.01	0.00	96.50
KW1	opaque	0.07	24.30	1.68	0.01	70.30	0.47	0.31	0.08	0.02	0.01	0.02	97.20
MB6	opaque	0.99	18.20	0.81	0.01	68.20	1.78	0.10	1.98	0.01	0.01	0.02	92.10
KW6	opaque in groundmass	0.16	21.40	1.52	0.00	70.90	0.44	0.18	0.14	0.03	0.01	0.03	94.90
KW6	opaque	0.36	22.60	1.96	0.03	67.60	0.67	0.27	0.12	0.02	0.01	0.02	93.60
TLB25	skeletal opaque	0.12	23.50	1.88	0.03	67.60	0.57	0.50	0.00	0.02	0.01	0.00	94.20
TLB25	large opaque	0.12	23.30	1.99	0.03	69.10	0.51	0.45	0.01	0.01	0.01	0.02	95.60
KW9	opaque	0.23	23.20	0.78	0.01	68.90	1.40	0.16	0.04	0.0 9	0.01	0.02	94.80
KW9	opaque	0.12	24.00	2.32	0.00	66.70	1.04	0.40	0.00	0.00	0.01	0.02	94.60
ML219	ilmentie (blade)	0.04	49.30	0.08	0.03	48.20	0.89	0.48	0.15	0.04	0.01	0.05	99.30
ML219	ilmentie (blade)	0.09	49.70	0.06	0.00	47.50	1.04	0.48	0.12	0.02	0.01	0.02	99.00
ML219	ilmentie (blade)	0.35	49.60	0.06	0.02	47.50	0.98	0.61	0.13	0.01	0.00	0.02	99.20
ML219	skeletal opaque	0.24	20.00	2.41	0.06	72.00	0.44	0.62	0.19	0.07	0.01	0.03	96.00
ML219	opaque	0.10	19.80	2.54	0.07	73.20	0.48	0.82	0.01	0.02	0.01	0.03	97.10
ML219	opaque	0.81	18.70	3.39	0.03	71.40	0.59	1.09	0.04	0.02	0.03	0.03	96.20
ML219	opaque	0.12	18.90	0.92	0.01	73.90	0.82	0.19	0.17	0.05	0.00	0.01	95.10
ML219	opaque	0.09	19.20	1.10	0.05	74.00	0.66	0.25	0.11	0.00	0.02	0.03	95.50
ML219	opaque	0.07	21.60	2.23	0.06	72.00	0.66	0.41	0.04	0.06	0.01	0.00	97.10

Appendix Giv:	Electron	microprobe	data ·	 oxide 	minerals.
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slide	information	0	Si	Ti	AI	Cr	Fe2	Mn	Mg	Ca	Na	к	Ni	Total
KW1	skeletal opaque	4	0.02	0.72	0.10	0.00	2.32	0.02	0.02	0.01	0.00	0.00	0.00	3.21
KW1	opque at rim	4	0.01	0.70	0.09	0.00	2.41	0.02	0.02	0.00	0.00	0.00	0.00	3.25
KW1	opaque in cpx	4	0.02	0.67	0.12	0.00	2.38	0.02	0.03	0.00	0.00	0.00	0.00	3.25
KW1	opaque	4	0.00	0.74	0.08	0.00	2.37	0.02	0.02	0.00	0.00	0.00	0.00	3.22
MB6	opaque	4	0.04	0.60	0.04	0.00	2.49	0.07	0.01	0.09	0.00	0.00	0.00	3.34
KW6	opaque in groundmass	4	0.01	0.68	0.08	0.00	2.49	0.02	0.01	0.01	0.00	0.00	0.00	3.28
KW6	opaque	4	0.02	0.71	0.10	0.00	2.36	0.02	0.02	0.01	0.00	0.00	0.00	3.23
TLB25	skeletal opaque	4	0.01	0.73	0.09	0.00	2.34	0.02	0.03	0.00	0.00	0.00	0.00	3.22
TLB25	large opaque	4	0.01	0.72	0.10	0.00	2.36	0.02	0.03	0.00	0.00	0.00	0.00	3.23
KW9	opaque	4	0.01	0.73	0.04	0.00	2.40	0.05	0.01	0.00	0.01	0.00	0.00	3.25
KW9	opaque	4	0.01	0.74	0.11	0.00	2.28	0.04	0.02	0.00	0.00	0.00	0.00	3.20
ML219	ilmentie (blade)	3	0.00	0.96	0.00	0.00	1.04	0.02	0.02	0.00	0.00	0.00	0.00	2.04
ML219	ilmentie (blade)	3	0.00	0.96	0.00	0.00	1.02	0.02	0.02	0.00	0.00	0.00	0.00	2.04
ML219	ilmentie (blade)	3	0.01	0.96	0.00	0.00	1.02	0.02	0.02	0.00	0.00	0.00	0.00	2.04
ML219	skeletal opaque	4	0.01	0.62	0.12	0.00	2.49	0.02	0.04	0.01	0.01	0.00	0.00	3.31
ML219	opaque	4	0.00	0.61	0.12	0.00	2.51	0.02	0.05	0.00	0.00	0.00	0.00	3.32
ML219	opaque	4	0.03	0.5 8	0.16	0.00	2.44	0.02	0.07	0.00	0.00	0.00	0.00	3.31
ML219	opaque	4	0.01	0.61	0.05	0.00	2.65	0.03	0.01	0.01	0.00	0.00	0.00	3.36
ML219	opaque	4	0.00	0.62	0.06	0.00	2.63	0.02	0.02	0.01	0.00	0.00	0.00	3.35
ML219	opaque	4	0.00	0.66	0.11	0.00	2.46	0.02	0.03	0.00	0.01	0.00	0.00	3.28

Appendix Gv: Electron microprobe data - alteration minerals.

slide	information	SiO2	TiO2	AI2O3	Cr2O3	FeO	MnO	MgO	CaO	Na2O	K20	NiO	Total
MI 219	olivine rim (darker altered)	46 53	0.04	5 4 1	0.02	18 85	0.30	16 95	1 73	0 10	0.43	0 02	90.38
ML219	alteration between olivine	49.37	0.04	3 73	0.02	16.31	0.00	18.98	1 22	0.10	0.45	0.04	90.44
ML219	olivine, bright BSE rim	45.47	0.05	3.08	0.03	22.53	0.49	17.74	1.80	0.05	0.31	0.02	91.57
ML219	altered rim of olivine	46.93	0.03	3.95	0.03	21.43	0.51	14.74	2.04	0.02	0.37	0.01	90.06
ML219	altered olivine rim	46.13	0.03	5.44	0.02	18.14	0.16	18.99	1.52	0.06	0.45	0.01	90.95
ML219	altered olivine	45.77	0.03	5.27	0.02	19.30	0.40	15.36	1.98	0.00	0.35	0.05	88.53
ML219	altered olivine	46.19	0.02	4.30	0.02	20.73	0.50	12.63	2.30	0.05	0.46	0.02	87.22
ML219	altered olivine	47.75	0.00	4.49	0.04	16.28	0.19	18.27	1.46	0.11	0.50	0.01	89.10
ML219	altered olivine	48.81	0.04	4.75	0.01	16.64	0.11	19.26	1.49	0.07	0.50	0.00	91.68
ML219	altered feld rim (next to fresh olivine)	39.21	0.01	5.86	0.01	17.14	0.22	17.10	1.30	0.15	0.28	0.04	81.32
ML219	altered feld rim (next to fresh olivine)	38.55	0.02	7.48	0.03	18.22	0.22	17.10	1.08	0.17	0.41	0.01	83.29
ML219	altered olivine	45.79	0.04	5.16	0.02	15.05	0.02	18.52	1.55	0.19	1.03	0.02	87.39
ML219	altered groundmass	51.25	0.12	10.05	0.03	11.12	0.09	13.87	2.43	2.76	0.62	0.02	92.36
ML219	altered vien	46.35	0.01	6.60	0.04	16.09	0.13	18.67	1.54	0.16	0.44	0.02	90.05
ML219	altered rim of olivine	47.46	0.04	4.16	0.01	20.56	0.50	13.03	2.30	0.01	0.13	0.05	88.25
ML219	altered rim of olivine	49.05	0.04	4.21	0.03	21.54	0.50	13.06	2.06	0.04	0.35	0.02	90.90
ML219	altered rim of olivine	49.35	0.01	4.28	0.02	13.97	0.08	20.70	1.51	0.07	0.47	0.02	90.48
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TLB25	pseudomorphed olivine	46.85	0.25	5.48	0.00	19.88	0.10	14.64	2.37	0.13	0.14	0.05	89.89
TLB25	pseudomorphed olivine	46.30	0.24	5.33	0.02	19.17	0.02	15.22	2.13	0.16	0.16	0.02	88.77
TLB25	pseudomorphed olivine	47.31	0.41	5.46	0.03	20.55	0.09	13.37	2.84	0.16	0.12	0.05	90.39
TLB25	pseudomorphed olivine	47.44	0.41	5.44	0.01	20.50	0.06	13.55	2.91	0.10	0.10	0.02	90.01
TLB25	groundmass alteration	42.84	2.48	4.80	0.02	21.73	0.18	13.17	1.52	0.10	0.14	0.00	00.90
TLB25	pseudomorphed olivine to brown clays	44.59	0.29	5.45	0.03	15.94	0.11	13.57	1.87	0.22	0.15	0.03	02.20
TLB25	pseudomorphed olivine to brown clays	46.77	0.27	5.68	0.03	20.48	0.09	12.94	2.30	0.11	0.15	0.02	00.92
TLB25	pseudomorphed olivine to brown clays	46.18	0.17	5.68	0.02	20.30	0.11	13.38	2.08	0.04	0.21	0.03	00.20
TLB25	altered groundmass	43.00	0.34	5.26	0.00	10.09	0.06	14.02	2.14	0.14	0.42	0.00	02.07
TLB25	Clay pseudomorphed olivine	43.04	0.01	5.20	0.02	10.92	0.04	14.04	1.01	0.12	0.14	0.02	01.00
TLB25	finegrained alteration in cracks of feld	45.83	0.35	14.89	0.02	14.17	0.04	14 00	1.02	2.00	0.32	0.00	91.29
TLB25	Olivine replacement	47.35	0.12	5.69	0.03	19.15	0.09	14.00	1.93	0.19	0.30	0.02	09.00
MB6	secondary vien	49.99	0.04	21. <b>21</b>	0.01	0.07	0.03	0.00	11.20	0.42	1.32	0.00	84.29
MB6	secondary vien	51.20	0.01	21.44	0.01	0.04	0.03	0.00	11.88	0.14	0.73	0.01	85.49
MB6	altered olivine	45.90	0.30	7.21	0.02	18.17	0.08	15.83	2.43	0.09	0.84	0.07	90.94
MB6	altered olivine	46.13	0.28	7.35	0.02	18.33	0.07	15.86	2.52	0.09	0.8 <del>9</del>	0.06	91.60
MB6	altered olivine	43.95	0.24	6.74	0.03	17.49	0.06	15.36	2.68	0.13	0.93	0.05	87.66
MB6	filled amygdale/replaced olivine?	43.04	0.09	6.83	0.04	16.73	0.04	15.39	2.23	0.16	0.66	0.00	85.21
MB2	altered feldspar	44.72	0.01	27.60	0.01	0.12	0.04	0.00	8.87	6.21	0.01	0.04	87.63
MB2	secondary vien	41.01	0.01	28.45	0.04	0.16	0.02	0.01	11.47	4.70	0.01	0.00	85.88
MB2	dark inclusion (feld altering???)	48.75	0.01	28.63	0.02	0.05	0.03	0.01	3.14	12.12	0.03	0.02	92.81
MB2	needle like alteration products	41.82	0.04	8.06	0.03	23.50	0.22	12.01	4.61	0.18	0.19	0.05	90.71
MB2	feid (v smali)	43.25	0.01	27.61	0.02	0.08	0.03	0.00	10.84	4.43	0.02	0.02	86.31
MB2	feld	49.51	0.02	28.77	0.00	0.02	0.01	0.00	2.45	12.97	0.05	0.02	93.82
MR16	feld (darker BSE bits, altered bits)	50.97	0.02	21.18	0.03	0.03	0.03	0.01	11.93	0.11	0.57	0.01	84.89
MB16	feld	50.67	0.04	21.08	0.01	0.27	0.03	0.00	11.18	0.30	0.71	0.01	84.30
MB16	buggered stuff????	49.87	0.01	20.78	0.02	1.00	0.01	0.47	11.02	0.38	0.62	0.03	84.21
KW6	pseudomorphed olivine in grounmass	46.21	0.72	4.76	0.03	21.45	0.08	12.74	1.55	0.03	0.23	0.02	87.82
KW6	clay pseudomorphed olivine	44.01	0.61	4.86	0.03	21.73	0.07	12.04	1.73	0.23	0.37	0.02	85.70
KW9	altered olivine rim	45.79	0.08	4.79	0.03	24.07	0.08	10.62	1.97	0.13	0.14	0.02	87.72
KW9	altered olivine middle	40.16	0.04	4.62	0.03	22.04	0.08	10.77	1.66	1.08	0.51	0.01	81.00
KW9	altered olivine in groundmass	44.08	0.16	4.22	0.01	24.60	0.17	7.35	1.83	0.08	0.06	0.00	82.56
KW9	centre of altered olivine enclosed in fe	45.52	0.11	5.32	0.01	23.94	0.01	11.99	1.92	0.19	0.30	0.02	89.33
KW9	rim of altered olivine enclosed in feld	45.14	0.12	4.94	0.00	24.24	0.11	10.57	1.93	0.15	0.24	0.03	87.47
KW9	brown altered cenre of pyroxene	40.39	0.60	4.14	0.02	21.94	0.10	11.34	1.68	0.24	0.42	0.04	80.91
KW9	brown altered cenre of pyroxene	44.77	0.62	4.21	0.03	23.27	0.09	11.37	3.19	0.21	0.24	0.02	88.02
KW9	pseudomorphed olivine	45.06	0.11	4.72	0.01	23.77	0.09	10.49	1.98	0.07	0.17	0.03	86.50

Ap	pendix	Gv:	Electron	micropro	obe da	ata - a	lteration	mineral	S
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slide	information	0	Si	Ti	AI	Cr	Fe2	Mn	Mg	Ca	Na	к	Ni	Total
MI 219	olivine rim (darker altered)	22	7 00	0.00	0.96	0.00	2 37	0.04	3 80	0.28	0.03	0.08	0.00	14.57
MI 210	alteration between olivine	22	7.28	0.00	0.65	0.00	2.07	0.04	1 17	0.20	0.00	0.00	0.00	14.07
ML213	olivine bright BSE rim	22	6.03	0.00	0.55	0.00	2.01	0.02	4.03	0.13	0.00	0.00	0.00	14.83
MI 210	altered rim of olivine	22	7 10	0.01	0.55	0.00	2.07	0.00	3 36	0.23	0.02	0.00	0.00	14.50
MI 210	altered olivine rim	22	6.88	0.00	0.71	0.00	2.74	0.07	4 22	0.00	0.02	0.07	0.00	14.69
WL215	altered oliving	22	7.06	0.00	0.90	0.00	2.20	0.02	3 53	0.24	0.02	0.03	0.00	14.50
ML219	altered olivine	22	7.00	0.00	0.90	0.00	2.43	0.05	2.03	0.33	0.00	0.07	0.01	14.36
ML219	altered olivine		7.29	0.00	0.00	0.00	2.73	0.07	2.97	0.39	0.02	0.09	0.00	14.00
ML219	altered olivine	22	7.17	0.00	0.79	0.00	2.04	0.02	4.09	0.24	0.03	0.10	0.00	14.50
ML219	altered fold rim (post to freeb olivino)	22	6.61	0.00	1 1 6	0.00	2.03	0.01	4.19	0.23	0.02	0.05	0.00	14.86
ML219	altered feld rim (next to fresh olivine)	22	6.01	0.00	1.10	0.00	2.42	0.03	4.23	0.24	0.05	0.00	0.00	14.00
ML219	altered leid him (next to fresh olivine)	22	7.00	0.00	0.02	0.00	1.02	0.03	4.22	0.19	0.00	0.03	0.00	14.50
ML219	altered olivine	22	7.02	0.00	1.67	0.00	1.90	0.00	4.23	0.20	0.00	0.20	0.00	14 37
ML219	altered vice	22	6 90	0.01	1.07	0.00	2.00	0.01	2.91 A 1A	0.37	0.75	0.11	0.00	14.59
ML219	allered vien	22	7.25	0.00	0.76	0.00	2.00	0.02	2.01	0.20	0.00	0.00	0.00	14.00
ML219	allered rim of oliving	22	7.00	0.00	0.76	0.00	2.00	0.07	2.01	0.30	0.00	0.03	0.01	14.20
ML219	altered rim of olivine	22	7.39	0.00	0.75	0.00	1 70	0.00	2.93	0.00	0.01	0.07	0.00	14.27
ML219	altered rim of olivine	22	7.19	0.00	0.73	0.00	1.70	0.01	4.50	0.24	0.02	0.03	0.00	14.45
TLB25	pseudomorphed olivine	22	7.11	0.03	0.98	0.00	2.52	0.01	3.31	0.39	0.04	0.03	0.01	14.42
TLB25	pseudomorphed olivine	22	7.09	0.03	0.96	0.00	2.46	0.00	3.48	0.35	0.05	0.03	0.00	14.45
TLB25	pseudomorphed olivine	22	7.16	0.05	0.97	0.00	2.60	0.01	3.02	0.46	0.05	0.02	0.01	14.35
TLB25	pseudomorphed olivine	22	7.16	0.05	0.97	0.00	2.59	0.01	3.05	0.47	0.05	0.02	0.00	14.36
TLB25	groundmass alteration	22	6.85	0.30	0.90	0.00	2.91	0.02	3.14	0.26	0.03	0.03	0.00	14.44
TLB25	pseudomorphed olivine to brown clays	22	7.26	0.04	1.05	0.00	2.17	0.02	3.29	0.33	0.07	0.03	0.00	14.25
TLB25	pseudomorphed olivine to brown clays	22	7.19	0.03	1.03	0.00	2.63	0.01	2.96	0.39	0.03	0.03	0.00	14.31
TLB25	pseudomorphed olivine to brown clays	22	7.15	0.02	1.04	0.00	2.63	0.01	3.09	0.35	0.01	0.04	0.00	14.35
TLB25	altered groundmass	22	7.05	0.04	1.02	0.00	2.29	0.01	3.57	0.38	0.04	0.09	0.00	14.48
TLB25	Clay pseudomorphed olivine	22	7.10	0.00	1.01	0.00	2.33	0.01	3.65	0.27	0.04	0.03	0.00	14.44
TLB25	finegrained alteration in cracks of feld	22	6.72	0.04	2.57	0.00	1.74	0.00	1.52	1.04	0.59	0.06	0.00	14.29
TLB25	Olivine replacement	22	7.15	0.01	1.01	0.00	2.42	0.01	3.35	0.31	0.06	0.07	0.00	14.41
MB6	secondary vien	22	7.29	0.00	3.65	0.00	0.01	0.00	0.00	1.75	0.12	0.25	0.00	13.07
MB6	secondary vien	22	7.33	0.00	3.62	0.00	0.00	0.00	0.00	1.82	0.04	0.13	0.00	12.95
MB6	altered olivine	22	6.86	0.03	1.27	0.00	2.27	0.01	3.53	0.39	0.03	0.16	0.01	14.56
MB6	altered olivine	22	6.85	0.03	1.29	0.00	2.28	0.01	3.51	0.40	0.03	0.17	0.01	14.57
MB6	altered olivine	22	6.84	0.03	1.24	0.00	2.27	0.01	3.56	0.45	0.04	0.18	0.01	14.63
MB6	filled amygdale/replaced olivine?	22	6.85	0.01	1.28	0.00	2.23	0.00	3.65	0.38	0.05	0.13	0.00	14.59
MB2	altered feldspar	22	6.39	0.00	4.65	0.00	0.02	0.00	0.00	1.36	1.72	0.00	0.00	14.15
MB2	secondary vien	22	6.04	0.00	4.94	0.00	0.02	0.00	0.00	1.81	1.34	0.00	0.00	14.16
MB2	dark inclusion (feld altering???)	22	6.56	0.00	4.54	0.00	0.01	0.00	0.00	0.45	3.16	0.00	0.00	14.75
MB2	needle like alteration products	22	6.53	0.00	1.48	0.00	3.07	0.03	2.79	0.77	0.06	0.04	0.01	14.78
MB2	feld (v small)	22	6.29	0.00	4.73	0.00	0.01	0.00	0.00	1.69	1.25	0.00	0.00	13.97
MB2	feld	22	6.59	0.00	4.52	0.00	0.00	0.00	0.00	0.35	3.35	0.01	0.00	14.83
MR16	feld (darker BSE bits, altered bits)	22	7.34	0.00	3.59	0.00	0.00	0.00	0.00	1.84	0.03	0.11	0.00	12.93
MB16	feld	22	7.35	0.00	3.60	0.00	0.03	0.00	0.00	1.74	0.08	0.13	0.00	12.95
MB16	buggered stuff????	22	7.28	0.00	3.58	0.00	0.12	0.00	0.10	1.72	0.11	0.12	0.00	13.04
KANA V	needemombed oliving in groupment	20	7 11	م مه	0 97	0.00	2 77	0.01	2 03	96.0	0.01	0.05	0.00	14 12
KW6	pseudomorphed olivine in grounmass	22	7.14	0.00	0.07	0.00	2.77	0.01	2.30	0.20	0.07	0.00	0.00	14.06
KW6	ciay pseudomorphed olivine	22	7.04	0.07	0.92	0.00	2.31	0.01	£.01	0.00	0.07	0.00	0.00	17.20
KW9	altered olivine rim	22	7.27	0.01	0.90	0.00	3.20	0.01	2.52	0.34	0.04	0.03	0.00	14.32
KW9	altered olivine middle	22	7.00	0.01	0.95	0.00	3.21	0.01	2.80	0.31	0.37	0.11	0.00	14.77
KW9	altered olivine in groundmass	22	7.49	0.02	0.84	0.00	3.49	0.02	1.86	0.33	0.03	0.01	0.00	14.11
KW9	centre of altered olivine enclosed in fe	22	7.11	0.01	0.98	0.00	3.13	0.00	2.79	0.32	0.06	0.06	0.00	14.46
KW9	rim of altered olivine enclosed in feld	22	7.22	0.01	0.93	0.00	3.24	0.02	2.52	0.33	0.05	0.05	0.00	14.37
KW9	brown altered cenre of pyroxene	22	7.02	0.08	0.85	0.00	3.19	0.01	2.94	0.31	0.08	0.09	0.01	14.58
KW9	brown altered cenre of pyroxene	22	7.13	0.07	0.79	0.00	3.10	0.01	2.70	0.54	0.06	0.05	0.00	14.47
KW9	pseudomorphed olivine	22	7.26	0.01	0.90	0.00	3.20	0.01	2.52	0.34	0.02	0.04	0.00	14.32

#### Appendix Gv: Electron microprobe data - alteration minerals.

slide	information	SiO2	TiO2	A12O3	Cr2O3	FeO	MnO	MgO	CaO	Na2O	К2О	NiO	Total
KW18	altered feld in groundmass	44.82	0.00	18.40	0.00	14.41	0.02	4.66	3.65	3.50	0.06	0.02	89.54
KW18	altered groundmass	32.92	0.02	14.55	0.03	31.32	0.18	5.98	2.12	2.41	0.10	0.02	89.65
KW18	altered groundmass	36.17	0.21	9.80	0.03	34.05	0.37	5.02	6.72	0.37	0.10	0.03	92.87
KW18	altered groundmass	39.53	0.04	15.22	0.02	24.89	0.12	5.00	2.50	1.59	1.55	0.04	90.50
KW18	altered groundmass	47.11	7.24	16.45	0.02	20.99	0.72	0.10	2.01	7.45	1.16	0.02	103.27
KW18	chlorite filled amygdale	27.77	0.03	13.44	0.03	42.62	0.20	5.02	0.43	0.01	0.00	0.05	89.60
KW18	chlorite filled amygdale	27.49	0.03	13.37	0.01	42.13	0.18	5.00	0.49	0.01	0.04	0.02	88.77
KW18	chlorite filled amygdale	27.20	0.04	13.11	0.03	42.80	0.14	4.65	0.36	0.02	0.01	0.02	88.38
KW13	feld in middle of cpx	43.05	0.08	22.57	0.04	12.85	0.04	3.97	9.14	1.21	0.75	0.00	93.70
KW13	inclusion in above feld	37.80	0.21	11.17	0.05	25.29	0.02	8.90	2.75	0.40	1.21	0.02	87.82
KW13	alteration groundmass xtals	42.84	0.86	21.91	0.03	10.53	0.03	0.54	8.05	4.12	0.37	0.12	89.40
KW13	alteration groundmass xtals	44.07	0.56	6.38	0.03	22.27	0.13	11.75	2.56	0.10	0.99	0.01	88.85
KW13	groundmass xtal	36.93	1.01	7.56	0.03	29.35	0.08	7.75	2.33	0.17	1.38	1.03	87.62
KW13	altered xtals	44.90	0.62	6.63	0.02	22.74	0.05	12.81	2.77	0.18	0.96	0.03	91.71
KW13	buggered edge of large feld	38.13	0.16	21.25	0.04	6.04	0.05	1.26	12.20	2.06	0.21	0.02	81.42
KW13	altered groundmass	54.94	0.24	17.31	0.00	10.61	0.02	2.84	4.51	5.12	0.67	0.00	96.26
KW13	altered groundmass	45.70	2.04	9.40	0.04	21.83	0.13	0.00	3.70	1.99	0.94	0.04	92.32
KW13	altered groundmass	43.00	0.75	0.14	0.02	20.90	0.07	6.66	5.00	1.10	0.97	0.01	88 42
KW13		43.50	0.00	0.51	0.05	20.70	0.14	0.00	1.52	1.10	0.50	0.00	00.72
KW1	Groundmass (replaced olivine?)	48.55	0.60	4.37	0.03	20.69	0.11	14.73	1.57	0.01	0.21	0.01	90.88
KW1	Groundmass (replaced olivine?)	48.20	0.57	4.48	0.00	21.28	0.10	14.70	1.61	0.09	0.31	0.02	91.36
KW1	brown alteration blobs in groundmass	46.94	0.85	4.12	0.03	21.44	0.13	12.49	2.00	0.05	0.24	0.02	88.31
KW1	brown alteration blobs in groundmass	45.62	0.71	11.55	0.00	13.64	0.08	14.72	2.41	3.55	1.24	0.02	00.00
	brown alteration blobs in groundmass	40.00	0.40	4.42	0.02	21.20	0.13	14.73	1.42	0.10	0.22	0.02	90.05
	brown altered stal middle	47.09	0.40	4.54 3.89	0.02	10 17	0.00	12.60	1.50	0.05	0.16	0.07	81 82
KW1	aroundmass cran	66 42	0.45	9.85	0.01	8.90	0.01	4.90	2.89	3.39	0.25	0.01	97.08
KW1	altered xtal in groundmass (once cpx?)	50.95	0.78	3.70	0.04	19.11	0.17	10.55	1.80	0.09	0.19	0.04	87.42
KW1	altered xtal. next to feld in cpx	48.54	0.65	4.28	0.03	20.70	0.09	14.38	1.47	0.14	0.41	0.06	90.75
KW1	altered xtal, next to feld in cpx	48.23	0.60	6.61	0.00	18.62	0.13	8.18	2.11	0.76	0.41	0.02	85.67
KW7	altered olivine	43.73	0.73	4.95	0.03	22.56	0.08	10.51	1.90	0.15	0.14	0.02	84.80
KW7	altered fine groundmass	53.08	0.60	12.55	0.04	13.00	0.03	5.46	4.27	4.20	0.34	0.05	93.62
KW7	altered fine groundmass	41.64	3.73	7.32	0.03	25.86	0.27	7.06	2.72	1.98	0.27	0.02	90.90
KW7	altered olivine	45.41	0.67	4.85	0.04	22.43	0.11	11.72	2.07	0.03	0.17	0.03	87.53
KW7	altered groundmass	55.01	0.49	14.56	0.02	9.92	0.01	3.33	3.71	4.36	2.48	0.02	93.91
KW7	pseudomorphed olivine	47.21	0.27	4.70	0.04	23.02	0.08	11.98	1.75	0.17	0.14	0.01	89.37
KW7	pseudomorphed olivine	46.33	0.40	4.71	0.04	22.36	0.05	11.64	1.75	0.11	0.14	0.02	87.55
KW7	pseudomorphed olivine	46.84	0.35	4.43	0.03	21.54	0.08	11.78	1.77	0.08	0.13	0.01	87.04
KW7	pseudomorphed olivine	47.95	0.49	4.80	0.01	22.80	0.11	12.18	1.70	0.20	0.12	0.02	90.38
KW7	pseudomorphed olivine	46.59	0.47	4.73	0.01	22.65	0.03	11.80	1.68	0.18	0.12	0.02	88.28
KW7	altered olivine in groundmass	45.04	0.76	4.70	0.02	22.63	0.12	11.44	1.78	0.07	0.15	0.07	86.78
KW7	v. white rim of olivine	47.23	0.41	4.80	0.00	22.31	0.09	11.95	1.47	0.04	0.12	0.02	88.44
MB8a	altered feldspar	53.79	0.05	27.53	0.03	1.10	0.02	0.14	10.97	5.21	0.24	0.02	99.10
MB8a	brown clay alteration of amygdale/oliv.	35.53	0.03	20.00	0.03	14.70	0.07	7.79	6.89	1.53	0.21	0.05	86.83
MB8a	green clay alteration of amygdale/oliv.	29.05	0.03	14.44	0.00	32.47	0.18	9.03	1.70	0.13	0.03	0.01	87.07
MB8a	green clay alteration of amygdale/oliv.	29.06	0.08	16.16	0.01	30.09	0.18	8.15	2.69	0.24	0.02	U.00	86.68
MB8a	green clay alteration of amygdale/oliv.	28.06	0.05	14.98	0.02	32.74	0.24	8.03	1.93	0.08	0.03	0.00	86.16
MB8a	brown clay alteration of amyg/olivine	35.64	0.06	11.86	0.00	22.48	0.10	14.02	2.63	0.29	0.21	0.01	07.30
MB8a MB8a	brown clay alteration of amyg/olivine altered olivine enclosed by plag	36.19 31.50	0.04 1.38	11.48 11.60	0.00 0.03	21.34 20.80	0.05	14.72 12.77	∠.38 3.32	0.31	0.28 0.24	0.00	82.02

App	pendix	Gv:	Electron	micro	probe	data -	alteration	minerals.
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slide	Information	0	Si	Ti	AI	Cr	Fe2	Mn	Mg	Ca	Na	к	Ni	Total
KW18	altered feld in groundmass	22	6.64	0.00	3.21	0.00	1.78	0.00	1.03	0.58	1.01	0.01	0.00	14.26
KW18	altered groundmass	22	5.50	0.00	2.87	0.00	4.38	0.03	1.49	0.38	0.78	0.02	0.00	15.46
KW18	altered groundmass	22	5.94	0.03	1.90	0.00	4.68	0.05	1.23	1.18	0.12	0.02	0.00	15.15
KW18	altered groundmass	22	6.23	0.00	2.82	0.00	3.28	0.02	1.17	0.42	0.49	0.31	0.00	14.75
KW18	altered groundmass	22	6.37	0.74	2.62	0.00	2.37	0.08	0.02	0.29	1.95	0.20	0.00	14.65
KW18	chlorite filled amygdale	22	4.96	0.00	2.83	0.00	6.36	0.03	1.34	0.08	0.00	0.00	0.01	15.62
KW18	chlorite filled amygdale	22	4.95	0.00	2.84	0.00	6.35	0.03	1.34	0.09	0.00	0.01	0.00	15.63
KW18	chlorite filled amygdale	22	4.95	0.00	2.81	0.00	6.51	0.02	1.26	0.07	0.01	0.00	0.00	15.64
KW13	feld in middle of cpx	22	6.14	0.01	3.79	0.00	1.53	0.00	0.84	1.40	0.33	0.14	0.00	14.19
KW13	inclusion in above feld	22	6.19	0.03	2.16	0.01	3.47	0.00	2.17	0.48	0.13	0.25	0.00	14.89
KW13	alteration groundmass xtals	22	6.35	0.10	3.83	0.00	1.31	0.00	0.12	1.28	1.19	0.07	0.02	14.26
KW13	alteration groundmass xtals	22	6.92	0.07	1.18	0.00	2.93	0.02	2.75	0.43	0.03	0.20	0.00	14.53
KW13	groundmass xtal	22	6.28	0.13	1.52	0.00	4.18	0.01	1.96	0.42	0.06	0.30	0.14	15.00
KW13	altered xtals	22	6.84	0.07	1.19	0.00	2.90	0.01	2.91	0.45	0.05	0.19	0.00	14.61
KW13	buggered edge of large feld	22	6.15	0.02	4.04	0.00	0.82	0.01	0.30	2.11	0.64	0.04	0.00	14.15
KW13	altered groundmass	22	7.36	0.02	2.73	0.00	1.19	0.00	0.57	0.65	1.33	0.11	0.00	13.97
KW13	altered groundmass	22	6.88	0.23	1.68	0.00	2.75	0.02	1.49	0.60	0.58	0.18	0.00	14.42
KW13	altered groundmass	22	6.93	0.09	1.52	0.00	2.77	0.01	1.59	1.03	0.34	0.20	0.00	14.49
KW13	altered groundmass	22	6.95	0.08	1.55	0.00	2.74	0.02	1.57	1.01	0.34	0.20	0.00	14.47
KW1	Groundmass (replaced olivine?)	22	7.27	0.07	0.77	0.00	2.59	0.01	3.29	0.25	0.00	0.04	0.00	14.30
KW1	Groundmass (replaced olivine?)	22	7.21	0.06	0.79	0.00	2.66	0.01	3.28	0.26	0.03	0.06	0.00	14.37
KW1	brown alteration blobs in groundmass	22	7.30	0.10	0.75	0.00	2.79	0.02	2. <del>9</del> 0	0.33	0.02	0.05	0.00	14.25
KW1	brown alteration blobs in groundmass	22	7.13	0.08	2.13	0.00	1.78	0.01	1.52	0.40	1.08	0.25	0.00	14.38
KW1	brown alteration blobs in groundmass	22	7.23	0.05	0.78	0.00	2.68	0.02	3.30	0.23	0.03	0.04	0.00	14.36
KW1	brown altered xtal core	22	7.27	0.05	0.78	0.00	2.63	0.01	3.26	0.26	0.01	0.04	0.00	14.31
KW1	brown altered xtal middle	22	7.29	0.06	0.77	0.00	2.68	0.02	3.14	0.30	0.02	0.03	0.00	14.30
KW1	groundmass crap	22	8.48	0.04	1.48	0.00	0.95	0.00	0.93	0.40	0.84	0.04	0.00	13.17
KW1	altered xtal in groundmass (once cpx?)	22	7.80	0.0 <del>9</del>	0.67	0.00	2.45	0.02	2.41	0.29	0.03	0.04	0.00	13.81
KW1	altered xtal, next to feld in cpx	22	7.29	0.07	0.76	0.00	2.60	0.01	3.22	0.24	0.04	0.08	0.01	14.32
KW1	altered xtal, next to feld in cpx	22	7.57	0.07	1.22	0.00	2.44	0.02	1.91	0.35	0.23	0.08	0.00	13.91
KW7	altered olivine	22	7.17	0.09	0.96	0.00	3.09	0.01	2.57	0.33	0.05	0.03	0.00	14.31
KW7	altered fine groundmass	22	7.43	0.06	2.07	0.00	1.52	0.00	1.14	0.64	1.14	0.06	0.01	14.08
KW7	altered fine groundmass	22	6.58	0.44	1.36	0.00	3.42	0.04	1.66	0.46	0.61	0.05	0.00	14.64
KW7	altered olivine	22	7.18	0.08	0.90	0.00	2.97	0.01	2.76	0.35	0.01	0.03	0.00	14.32
KW7	altered groundmass	22	7.59	0.05	2.37	0.00	1.14	0.00	0.68	0.55	1.17	0.44	0.00	13.99
KW7	pseudomorphed olivine	22	7.29	0.03	0.86	0.00	2.97	0.01	2.76	0.29	0.05	0.03	0.00	14.30
KW7	pseudomorphed olivine	22	7.29	0.05	0.87	0.00	2.94	0.01	2.73	0.30	0.03	0.03	0.00	14.26
KW/	pseudomorphea olivine	22	7.38	0.04	0.82	0.00	2.84	0.01	2.77	0.30	0.02	0.03	0.00	14.21
	pseudomorphed olivine	22	7.30	0.06	0.80	0.00	2.90	0.01	2.77	0.28	0.06	0.02	0.00	14.27
	eltered eliving in groundmass	22	7.20	0.00	0.07	0.00	2.90	0.00	2.75	0.20	0.03	0.02	0.00	14.20
KW7	v. white rim of olivine	22	7.33	0.05	0.88	0.00	2.90	0.02	2.73	0.24	0.02	0.03	0.00	14.22
MB8a	altered feldspar	22	6.79	0.00	4.09	0.00	0.12	0.00	0.03	1.48	1.27	0.04	0.00	13.83
MB8a	brown clay alteration of amygdale/oliv		5.61	0.00	3.72	0.00	1.94	0.01	1.83	1.16	0.47	0.04	0.01	14.80
MB8a	green clay alteration of amygdale/oliv.		5.06	0.00	2.96	0.00	4.73	0.03	2.34	0.32	0.04	0.01	0.00	15.50
MB8a	green clay alteration of amygdale/oliv.		5.02	0.01	3.29	0.00	4.35	0.03	2.10	0.50	0.08	0.00	0.00	15.38
MB8a	green clay alteration of amygdale/oliv.		4.97	0.01	3.12	0.00	4.84	0.04	2.12	0.37	0.03	0.01	0.00	15.50
MB8a	brown clay alteration of amyg/olivine		5.80	0.01	2.27	0.00	3.06	0.01	3.40	0.46	0.09	0.04	0.00	15.14
MB8a	brown clay alteration of amyg/olivine		5.88	0.00	2.20	0.00	2.90	0.01	3.56	0.41	0.10	0.06	0.00	15.11
MB8a	altered olivine enclosed by plag	22	5.50	0.18	2.39	0.00	3.04	0.01	3.32	0.62	0.09	0.05	0.01	15.21

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## Appendix Aii Kwaleunga River (1/2)

Tanya Louise Babbs Doctor of Philosophy June 1997

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