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# The Volcanic History and Geochemical Evolution of the Hveragerði Region, S. W. Iceland.

by

# Cherry L. Walker

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A Thesis submitted in partial fulfilment of the requirements for the degree of Doctor of Philosophy

**Department of Geological Sciences** 

**University of Durham** 

**1992** 



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2 DEC 1992

I dedicate this thesis

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to

Joseph Christian Pandeli

# ABSTRACT

# The Volcanic History and Geochemical Evolution of the Hveragerði Region, S. W. Iceland.

#### Cherry Walker, University of Durham, 1992

The Hveragerŏi Region is situated at the Hengill Triple junction, SW Iceland, where there are three volcanic systems. The crust in the area is constructed from both fissure (elongate) and lava shield (conical) eruptive units. Hengill is the presently active spreading zone with the Hengill Central Volcano, whereas the Hveragerŏi region is inactive with the extinct Grensdalur Central Volcano. Recent geophysical research indicates the presence of high and low density volumes within the upper 5 km of the crust in this area. The location of the density anomalies coincides with surface geological features, such as Recent lava shields, and the extinct and active central volcanoes.

A geological map of the Hveragerði Volcanic System has been constructed, and approximately 450 basaltic samples have been examined petrographically and analysed for wholerock, volcanic glass and mineral chemical data from this region. Observations from these data, coupled with the geological and geophysical observations, suggest that the lava shields are fed straight from the base of the crust, whereas fissure eruptions originate from shallow crustal reservoirs The character of the crustal reservoir has been highly variable in the past 1 Ma, and has varied from a melt-dominated reservoir, to a crystal mush-dominated one.

Each lava shield is compositionally distinct and is thought to preserve the mantle-melting signature. The compositional variation amongst the lava shields suggests that instantaneous melts are able to segregate from the mantle without complete mixing with accumulated melts from the entire length of the melting column. The depleted instantaneous melts from the crest and top of the melting column will either form picritic lava shields or they may interact with more fractionated crustal reservoirs and undergo quench crystallisation of megacrystic plagioclase (An 80-90). These crystals with associated pyroxene and olivine become flotation cumulates. There are episodes within the stratigraphy where off-axis lava shield and highly megacrystic fissure eruptions dominate, and such periods may represent periods of low magma supply.

On the submerged mid-ocean ridges, linear and conical features are also observed, and these may be analogous to the fissure and lava shields, respectively. The basalt types reported here from the Hveragerði region have also been reported off-shore, and they may therefore represent basalts derived from similar magmatic processes in a similar magmatic plumbing system. However, an initial observation of the relationship of 12 dredged basalts from 63°10'N on the Reykjanes Ridge suggests that this is not the case at this locality.

# DECLARATION

I declare that this thesis, which I submit for the degree of Doctor of Philosophy at the University of Durham, is my own work and is not substantially the same as any which has previously been submitted for a degree at this or another university.

Cherry L. Walker

University of Durham

August 1992

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#### Maps

A1 & A2 - Geological maps of the Hveragerŏi Volcanic System B1, B2 & B3 - Sample location maps for whole-rock samples C - Sample location map for volcanic glass samples

#### Publications

- Parson, L. M., Murton, B. J., Searle, R. C., Field, P., Keeton, J., Walker, C. L., Shor, A., Smith, D., McAllister, E., Spencer, S., Redbourne, L., Evens, J., Millard, N., Rouse, I., Summerhayes, C. & Laughton, A. (submitted to EPSL) 1992. En echelon volcanic ridges at the Reykjanes Ridge: A life cycle of volcanism and tectonics.
- Walker, C. L. 1991. North Atlantic ocean crust and Iceland. In: Floyd P. (ed.). Oceanic Basalts, Blackie & Sons, Glasgow, 311-352.

# Chapter 1

# Introduction

# **1.1) Tectonics of Iceland**

Iceland is an island in the North Atlantic Ocean at approximately 65°N and 20°W. The land mass is a result of the coexistence and interaction of two phenomena: a constructive plate boundary; and a mantle plume, expressed at the surface as the Icelandic hotspot. Excessive magma production from the hot plume has produced the Icelandic plateau (McKenzie, 1984) on which the constructive plate boundary is expressed as a complicated arrangement of active volcanic and seismic zones (Figure 1.1) (Ward, 1971; Pálmason & Sæmundsson, 1974; Sæmundsson, 1974; 1979; Einarsson, 1991).

The three main actively spreading volcanic zones are the Western, Eastern and Northern volcanic zones (Figure 1.1). The Western Volcanic Zone (WVZ) is the onland continuation of the Reykjanes Ridge and extends inland becoming progressively wider and trending dominantly NE. The Northern Volcanic Zone (NVZ) has a more northerly trend, and extends from the north coast of Iceland to the Vatnajökull glacier. The Eastern Volcanic Zone (EVZ) is the widest (72 km) and trends NE, subparallel to the WVZ from Vatnajökull, as far south as the Westman Islands. A comparatively small volcanic zone, the Öræfajökull Volcanic Zone (OVZ) crosses the east corner of Vatnajökull (Steinthórsson *et al.*, 1985). There is possibly a fifth zone trending E-W on the Snæfellsnes Peninsula in west Iceland, the Snæfellsnes Volcanic Zone (SVZ).

The term "neovolcanic zone" in Icelandic literature refers to all the regions where volcanic material has erupted during Postglacial times and includes all the volcanic zones mentioned above. The WVZ and NVZ, together with the northern half of the EVZ, form the axial rift zone and are regarded as similar to a typical midocean ridge environment. The southern half of the EVZ, and the OVZ and SVZ are



referred to as flank or lateral zones (Sæmundsson, 1978), as many of their volcanotectonic properties differ from those of the other volcanic zones.

Figure 1.1: Tectonic map of Iceland and the surrounding region. NVZ = Northern Volcanic Zone; WVZ = Western Volcanic Zone; EVZ = Eastern Volcanic Zone; SVZ = Snæfellsnes Volcanic Zone; OVZ = Öræfajökull Volcanic Zone; TFZ = Tjörnes Fracture Zone; SISZ = South Iceland Seismic Zone; The axial and flank zones are differentiated accordingly. Names referred to in the text include:<math>K=Krafla; L=Laki & Eldgjá; St=Skaftafell; S=Skógar; RP=Reykjanes Peninsula; SP=Snæfellsnes Peninsula.

There are two regions of high seismicity on the Iceland plateau which are comparable to transform zones found on mid-ocean ridges (Einarsson, 1991). The Tjörnes Fracture Zone (TFZ) on the north coast of Iceland connects and offsets the subaerial NVZ from the submarine Kolbeinsey Ridge. In the south of Iceland there is an E-W trending zone of high seismicity known as the South Iceland Seismic Zone (SISZ) (Einarsson *et al.*, 1981). It connects the WVZ and EVZ without laterally offsetting either of them, but it does influence the tectonic fabric within each. In the southern part of the EVZ the topographic fabric is deflected from the normal NE trend to an E-W orientation, and in the WVZ, faults and fissures deviate from the normal NE trend to a N-S orientation (Sæmundsson, 1967). The southernmost tip of

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the EVZ is estimated to have propagated from north to south at a rate of 3.5-5 cm yr<sup>-1</sup>, bringing the transient SISZ with it (Einarsson, 1988).

The axial rift and flank zones are segmented to form volcanic systems (Sæmundsson, 1978) (Figure 1.2). These are composed of a set of parallel eruptive ridges and fissures and represent a grouping of eruption sites that are spatially and temporally related. In this thesis, the term "eruption unit" refers to a single eruption event with a clear beginning and an end (usually defined by sediment deposition and/or erosion). Non-orthogonal spreading has resulted in an en echelon arrangement of the volcanic systems within the axial rift zones which facilitates recognition of each system (Figure 1.2). When the systems are arranged orthogonally it is harder to define their longitudinal extent.



Figure 1.2: A map of Iceland indicating the locations of the volcanic systems (stippled) and central volcanoes (black) of the neovolcanic zone (after Jakobsson, 1979b).

Volcanic systems vary in length from 17 to 30 km, in width from 5 to 10 km (Jakobsson, 1979a), and may persist for 0.3-1 Ma (Sæmundsson, 1979). They usually contain a central volcano (G. P. L. Walker, 1963; Sæmundsson, 1978) which displays a combination of the following features: a high-temperature geothermal system; a radial component of stress, evident from the late-stage eruptive fissures; an increase in the percentage of intrusives; repeated fissure eruptions at one site; a caldera; the presence of intermediate and acidic rocks. The diameter of the central volcano approximately equals the width of the fissure swarm. These volcanoes have magma reservoirs in the order of 10's of cubic km in volume that exist at shallow

#### Chapter 1

levels in the crust (<5 km) (Pálmason, 1971; G. P. L. Walker, 1974; Björnsson *et al.*, 1977; Einarsson, 1978; Flóvenz, 1980; Foulger & Toomey, 1989). These volcanoes are not equivalent to the "central volcanoes" described from of the East Pacific Rise (EPR) (Searle, 1983; Macdonald, 1986), or those on the South Atlantic mid-ocean ridge (MOR) at 26°S (Batiza *et al.*, 1989), but appear to be peculiar to Iceland.

The sub-surface structure of Iceland has been investigated using a number of geophysical methods, including gravity (e.g. Thorbergsson et al., 1990), various seismic techniques (Pálmason, 1971; Reykjanes Ridge Iceland Seismic Project [RRISP], 1980; Foulger & Toomey, 1989) and electrical studies (Beblo & Björnsson, 1978; 1980; Hersir et al., 1984). Collectively, they infer that the lithosphere is 8-10 km thick at the axial rift zone, and thicker on the margins of the plateau (20-30 km). The crustal part of the lithosphere is of variable thickness over the plateau, and is thickest under the EVZ flank zone, and older crust in NW Iceland (summary in Flóvenz & Gunnarsson, 1991). The structure within the crust is found to be heterogeneous in its physical properties at the axial zone and away from it. Eruptive products exposed at the surface, and observed in seismic reflection profiles, dip towards the neovolcanic zone by 10-15° (Sæmundsson, 1979; Zverev et al., 1980) and, in older crust, small synclines are seen. These are thought to represent previous locations of the spreading axis and indicate jumps of the order of tens of km (Helgason, 1985; 1989). This process has been occurring for at least 27 Ma (e.g. Helgason, 1984; Jancin et al., 1985; Steinthórsson et al., 1985; summary in C. L. Walker, 1991).

# **1.2) Geology of Iceland**

### 1.2.1) Rock types and ages

The Icelandic crust is thought to be composed of at least 2-6 km of extrusive material (Pálmason, 1971; Sæmundsson, 1979; Zverev *et al.*, 1980; Flóvenz & Gunnarsson, 1991), which is generally more than the 2 km of extrusives in oceanic crust (e.g. White, 1989). The exposed part of the Icelandic crust is composed of 80-85% basalt and 10-15% of intermediate and acidic rocks. Most of the latter occurs as intrusions (Sæmundsson, 1979). The igneous rocks observed on Iceland include both alkali and tholeiitic varieties, which are found in distinct parts of the island. The alkali series is restricted to the flank zones, whereas the tholeiitic series is found in the axial rift zone (Jakobsson, 1979a). The tholeiitic basalts occur as three petrographic types: 1) olivine tholeiites; 2) aphyric tholeiites; and 3) highly

#### Introduction

porphyritic tholeiites (Sæmundsson, 1979). The volcanic products are divided into four stratigraphic sub-divisions: i) the Postglacial (<9,000-13,000 years); ii) the  $U_{pper}$  Pleistocene (younger than 0.7 Ma, excluding the Postglacial); iii) the Pleistocene (0.7-3.1 Ma); iv) the Tertiary (older than 3.1 Ma) (Sæmundsson, 1979). The oldest rocks are 16 Ma and occur in NW Iceland (Pálmason & Sæmundsson, 1974).

#### **1.2.2) Eruptive environments**

The eruptive environment dominates the morphology of Iceland's volcanic zones (Thorarinsson, 1974), and this has fluctuated a great deal during the last few million years. Three environments have existed: subaerial, subglacial and shallow submarine. Subaerially erupted material forms lava flows. Each flow is guided by pre-existing topography, such as valleys, or it may cover flat plains, burying previously erupted material. Subglacial eruptions form hyaloclastite deposits that build steep to vertical-sided conical mountains or elongate ridges. These are referred to as table mountains (tuyas) and moberg ridges (tindas) respectively. During the Tertiary, thick piles of subaerial lavas were erupted until the onset of glaciation during the Pliocene. From this period to the Holocene, the axial rift zone was intermittently exposed above the ice, so that the hyaloclastite deposits are intercalated with subaerial flows. The ice retreated 9,000-13,000 years ago and most of the axial rift zone has since experienced subaerial volcanism (Sæmundsson, 1979).

Erosional processes, due to the action of ice, wind, rivers and coastal seas, have modified the original shape of the volcanic landscape. The erosional elements exploit weaknesses in volcanic edifices such as faults and hydrothermally altered areas. In addition, the heterogeneous nature of volcanic successions leads to differential erosion of the topography. The Tertiary rocks of Iceland form resistant lava plateaus, whereas the hyaloclastite edifices of the Plio-Pleistocene are easily eroded down, even between successive eruptions within a single volcanic system.

#### **1.2.3)** Volcanic eruption types

During the Postglacial era, two types of volcanic eruption event seem to have taken place producing either conical or linear volcanic morphologies. In the subglacial deposits, conical and linear topographic features also occur. These two morphological types are referred to as **lava shield** eruptions and **fissure** eruptions respectively. There is however a continuum between these two end member morphological features within the axial rift zone (G. P. L. Walker, 1965). Figure 1.3

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is a schematic representation of profiles for each of these morphologically distinct volcanic features, and an impression of how the volcanic products appear in plan view. Figure 1.4 illustrates examples of each eruption type from the WVZ (Kjartansson, 1967; Landmælingar Íslands, 1988).



Figure 1.3: Schematic representation of profiles and plan views of the Icelandic morphologically distinct volcanic features as they appear when erupted either subaerially or subglacially; a) lava shields; b) fissure eruptions. The scale for each eruption type is highly variable.

Lava shields are conical edifices with an aspect ratio in plan (i.e. length/width) usually of <2. The subaerial examples are characterised by an accumulation of petrographically monotonous lavas erupted from a single, centrally positioned vent, essentially as one continuous eruption (Sæmundsson, 1992). The volume of erupted material may be small or large (<1-17 km<sup>3</sup>: Jakobsson, 1979a; Sæmundsson, 1992). The edifices have low angles of slope on their flanks (1-10°) forming domes, believed to be due to low effusion rates of the lavas (G. P. L. Walker, 1971). The subglacial variety form flat-topped, steep-sided table mountains, conical or slightly ellipsoidal in shape. Skjaldbreiður and Hlöðufell are classic subaerial and

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subglacial examples, respectively, of lava shields (Figure 1.4). The spatial and temporal relationship of the lava shields to the volcanic systems varies within the axial rift zone, and they are absent from the flank zones. Away from the axial rift zone, in the Tertiary Icelandic crust, lava shields have also been identified in the lava profiles (Rutten, 1964; J. Helgason, pers. comm., 1990).



Figure 1.4: Morphologically distinct types of volcanic features that occur in Iceland: (a) field sketch of a portion of the WVZ south of the Langjökull glacier (Kjartansson, 1967). The summit if Hlööufell is 3 km across; (b) a topographic map of the region illustrated in (a) with a contour interval of 200 m (after Landmælingar Íslands, 1988).

The eruptive products from **Fissures** form typically elongate volcanic structures, the character of which is also influenced by the eruptive environment. If subaerial, a small number of lava flows may erupt from the fissure that usually travel

some distance from the vent (G. P. L. Walker, 1971). The volume of erupted material may also be small or large (<1-13.5 km<sup>3</sup>, Vilmundardóttir, 1977; Jakobsson, 1979b). The resulting shape of the edifice at the vent is dependent on the composition of the magmas and duration of the eruption, but is usually represented by a line of cinder and/or spatter cones. The lavas do not form thick dome-shaped piles that typify subaerial lava shields but are more laterally extensive (Figure 1.3b), even though they may be of equivalent volume (e.g. the Laki fissure eruption and the Skjaldbreiður lava shield eruption are both large-volume eruptions). This is believed to be the result of higher effusion rates for the fissure eruptions (G. P. L. Walker, 1971). The subglacial variety of fissure eruptions form elongate undulating hills of varying magnitude and of limited lateral extent (Figure 1.3b). Examples of subaerial and subglacial fissure eruptions are  $b \delta r \delta l f s f e l$ .

# **1.3)** Geology and tectonics of the Hengill-Ingolfsfjall region

The study area of this thesis is approximately 400 km<sup>2</sup> and lies at 64°N and 21°W, within the WVZ. The previous research is extensive and is largely summarised in Table 1.1. The area can be divided into three regions (Figure 1.5): i) the Hengill region, which includes the presently active central volcano (reported to be the Hengill Mountain; e.g. Björnsson & Hersir, 1981; Foulger & Toomey, 1989) within the Hengill Volcanic System; ii) the Hveragerði region, which includes the extinct Grensdalur Central Volcano within the Hveragerði Volcanic System; and iii) the Hrómundartindur region, which lies between the Hengill and Hveragerði regions.

The area contains numerous eruptive units, at least four of which are lava shields (Árnason *et al.*, 1986; 1987). These are, in order of decreasing age, Ingolfsfjall, Hæðir, Húsmúli and Skálafell (Figure 1.5). There are other lava shields in the area that have been either eroded or buried so that they are no longer morphologically distinct features. However, they may still be recognised by the geometry of the eruptive products (Rutten, 1964).

The Hengill Volcanic System is the most south-westerly in Iceland reported to contain a central volcano (Jakobsson *et al.*, 1978; Árnason *et al.*, 1987) (Figure 1.2). The lava shields of Hæðir, Húsmúli and Skálafell are contemporaneous with this volcanic system (Árnason *et al.*, 1987). The Hrómundartindur Volcanic System has very few extensional tectonic features, and the fissure-erupted ridges have lower aspect ratios than those of the Hengill system (Figure 1.5a). No lava shields occur within this system. The Hveragerði Volcanic System consists of smooth linear hills

and a horse-shoe shaped topographic-low area which is inferred to be the position of the extinct Grensdalur Central Volcano (Björnsson *et al.*, 1974; Foulger & Long, 1984; Foulger & Toomey, 1989) (Figure 1.5). Ingolfsfjall is contemporaneous with the Hveragerði Volcanic System. It is a prominent, flat-topped, slightly elongate table mountain representing a lava shield situated 10 km SE of the Hveragerði village.

The Hveragerõi Volcanic System is the oldest of the three systems in the region (Sæmundsson, 1967). When it became inactive at approximately 200 ka, the volcanic and tectonic activity of the area was shared between the Hengill and Hrómundartindur systems (Árnason *et al.*, 1987). The volcanism in the whole region has alternated between fissure and lava shield type eruptions during the last 1 Ma. The local tectonics and geology record spatial oscillations of the locus of spreading and volcanism within the area, although the dominant trend is westward migration (Árnason *et al.*, 1987). There have been up to 13 km of crustal extension within this 24 km broad zone since the middle of the Pleistocene (700 ka), assuming a full spreading rate of 1.88 cm yr<sup>-1</sup> derived from the plate-movement model NUVEL (DeMets *et al.*, 1990).

The tectonics of the area are further complicated by the fact that the region is a triple junction at present, and has been influenced by the southerly migration of the SISZ (Einarsson, 1988). This seismic zone is thought to be currently located to the south of the study area. The orientation of faults and recent eruptive fissures in the zone is N-S, reflecting a change from an extensional-fault to a shear-fault mode of deformation (Foulger, 1988b). A study of local seismicity also suggests that deformation occurs on N-S orientated fault planes (Foulger, 1988b).

Magnetotelluric measurements suggest that the lithospheric thickness beneath the Hengill Volcanic System is 7.5-8 km, below which there is a 3 km thick layer of high conductivity (Hersir, 1980; Björnsson & Hersir, 1981; Hersir *et al.*, 1984). This high-conductivity layer is interpreted as the melt-rich top of the asthenosphere at the crust-mantle boundary (Beblo & Björnsson, 1978; 1980; Hersir *et al.*, 1984). Earthquakes are absent deeper than 10 km (Foulger, 1988a) which complies with this interpretation. Seismic and gravity studies show that the crust is laterally heterogeneous (Flóvenz, 1980; Thorbergsson *et al.*, 1984; Foulger & Toomey, 1989). The zone with a high velocity gradient, which may be the extrusive layer, is only 2 km thick here (Foulger & Toomey, 1989). a)



Figure 1.5: The study area: a) LANDSAT image of the region (Landmælingar Íslands, 1990); b) a summary map of Hengill-Ingolfsfjall region as shown in (a); the main tectonic features are shown, including the active Hengill and extinct Grensdalur central volcanoes (after Foulger & Toomey, 1989). The heavy dashed lines indicate the approximate boundaries between the three volcanic systems (Hengill Hrómundartindur and Hveragerði, from left to right). The morphologically distinct lava shields are labelled, as are other place-names referred to in the text. The exact location of the Hengill Central Volcano is challenged in this thesis.

Introduction



Figure 1.5b: continued.

Drilling in the Nesjavellir valley, to the north of the Hengill mountain, has mapped the upper 2 km of crust in that region (Franzson *et al.*, 1986; Franzson, 1988). At 1800 m below sea level the crust is composed of 80-100% intrusives and the alteration is in the epidote facies.

Earth Science Field	Technique	Region	References
Geology	Mapping	Hengill &	Sæmundsson, 1967.
		Hrómundartindur	Árnason et al., 1987.
			McFarlane, 1990.
			Postlethwaite, 1990.
			Pullinger, 1991.
		Hveragerði-Ingolfsfjall	Einarsson, 1951; 1962.
			Eiriksson, 1973.
			Jónsson, 1989.
Petrology and	Petro-	Hengill and	Jónsson, 1977.
Geochemistry	geochemical	Hrómundartindur.	Hardardóttir, 1983; 1986.
			Trønnes, 1990.
			Hansteen, 1991.
Geophysics	Electrical	SW Iceland	Hersir, 1980.
	1	2	Hersir et al., 1984.
		Study area	Björnsson et al., 1974.
			Björnsson & Hersir, 1981.
		All of Iceland	Hermance & Grillot, 1970; 1974.
			Grillot, 1973.
			Hermance, 1973.
	Seismic	SW Iceland	Pálmason, 1971; 1973.
	Refraction/		Zverev et al., 1980.
	reflection		Gebrande et al., 1980.
			Flóvenz, 1980.
	Earthquake	Study area	Foulger & Einarsson, 1980.
	Seismology		Foulger, 1984; 1988a & b.
			Foulger & Long, 1984.
			Einarsson, 1988; 1991.
			Toomey & Foulger, 1989.
			Foulger & Toomey, 1989.
	Gravity	Study area	Arnason <i>et al.</i> , 1986.
			Thorbergsson et al., 1984.
			Hersir et al., 1990.
Other	Alteration	Hengill and	Árnason et al., 1969.
		Hveragerði	Sigvaldason, 1963.
	Palaeomagnetic	Ingolfsfjall	Sæmundsson & Einarsson, 1980.
			Kristjánsson et al., 1988.
	Drilling	Hveragerði	Sæmundsson & Árnason, 1971.
		Hengill	Franzson et al., 1986.
			Franzson, 1988.
		SW Iceland	Pálmason et al., 1979.

**Table 1.1:** A summary of previous studies within the area, defined by Figure 1.5, and other directly relevant investigations.



Figure 1.6: A map of the study area with the density structure of the crust indicated. The Bouguer gravity anomalies (heavy contours) are indicated. The Bouguer density is assumed to be 2.6 g/cm<sup>3</sup> (from Hersir et al., 1990). The seismic velocity structure of the crust is also indicated (after Foulger and Toomey, 1989), an explanation for which is provided in the Key. (See also Figure 1.7). The locations of the lava shields are indicated.





Figure 1.7: Three-dimensional image of tomography results (Toomey & Foulger, 1989). The colour scale denotes the percentage difference in velocity from the regional background velocity. Each cubic block is 0.25 km. The view is from the NE, looking south along the rift zone. The location of Húsmúli (H), Hrómundartindur (Hr), and the Grensdalur (GCV) and Hengill (HCV) central volcanoes at the surface are indicated.

A tripartite geothermal system encompasses the area (Foulger, 1984). It is subdivided, in order of decreasing temperature maxima, into the Hengill field within the Hengill Volcanic system, the Ölkelduháls field hosted by the Hrómundartindur system, and the Hveragerði field in the vicinity of the Grensdalur Central Volcano (Árnason *et al.*, 1987).

Tomographic, gravity and electrical investigations (Thorbergsson et al., 1984; Foulger & Toomey, 1989; Hersir et al., 1990) have identified heterogeneity within the crust under the Hengill-Hverageroi region (Figures 1.6 & 1.7). The crust below the extinct Grensdalur Central Volcano is more dense (i.e. higher seismic velocities and Bouguer gravity anomalies) than the surrounding region, and tomographic evidence indicates that this is from the surface down to 3 km depth, occupying a volume of greater than 40 km<sup>3</sup>. The crust below the lava shields is also denser than the surroundings (Thorbergsson et al., 1984) and tomographic investigation has imaged a 5 km long high density cylindrical conduit below the Húsmúli lava shield (Figure 1.7) (Foulger & Toomey, 1989; Toomey & Foulger, 1989). Underneath and to the south of the Hengill mountain there is a lower density region (i.e. lower seismic velocities and Bouguer gravity anomalies) between 2 to 4 km depth, which may indicate a shallow magma reservoir. It is most persistent to the north of the mountain, under the Nesjavellir valley. Some earthquakes have been identified from this lowdensity region, indicating that melt probably exists in pockets within a crystal mush (Foulger & Toomey, 1989).

# **1.4) The research project**

Prior to the commencement of the work described in this thesis, the Hengill and Hrómundartindur regions had been mapped in detail as part of the geothermal development project in the Nesjavellir region (Árnason *et al.*, 1986; 1987). Petrological studies had also been carried out on various eruptive units of the Hengill and Hrómundartindur regions (Hardardóttir, 1983; 1986; Trønnes, 1990; Hansteen, 1991). Mapping of the Hveragerði-Ingolfsfjall region had been at a reconnaissance level (Sæmundsson, 1967; Eiriksson, 1973; Jónsson, 1989), and geochemical analyses of rocks were non-existent.

The primary objective of this research was to establish the stratigraphy of the Hveragerði-Ingolfsfjall area and to examine the geochemical variation of the volcanic products there and in the whole region. A geological map was compiled

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#### Chapter 1

from data accumulated during six months of fieldwork. The mapped area is 72 km<sup>2</sup>, recognises 81 eruptive units, and includes the Grensdalur Central Volcano and the Ingolfsfjall lava shield (Maps A1 & A2). The field relationships of the volcanic products and sediments were also studied. A suite of 560 samples was collected from the whole area, using the existing maps of the Hengill and Hrómundartindur regions and the new map of the Hveragerði region as sampling guides. Petrographic features were observed, and major oxide and trace element compositions were determined for a subgroup of approximately 425 whole-rock samples and 24 glass samples. The sample locations are illustrated on Maps B1, B2, B3 and C1.

The following topics were addressed:

Stratigraphy

i) Stratigraphy of the Hveragerði-Ingolfsfjall region.

ii) Absolute K/Ar age determinations.

## Petrology

i) Petrographic nature of the rocks encountered.

ii) Compositions of minerals, volcanic glass and whole-rock samples within the region.

iii) Regional geochemical variation.

iv) Magma compositions in relation to crystallization processes, nature of the magmatic plumbing system, mantle melting dynamics, and the compositional nature of the mantle source region.

# Iceland in relation to mid-ocean ridges

i) Morphological character of the neovolcanic zone on the Reykjanes Ridge and a comparison of Icelandic subglacial topography with the bathymetry of the seafloor.

ii) Geochemistry of basalts collected 150 km south of Iceland and a comparison of these with the samples from Icelandic study area.

In this thesis, Chapters 2 to 6 contain descriptive accounts of observations made during this research. A summary is provided at the end of each chapter listing the main observation. In Chapter 7, the information is collated, interpreted and discussed in terms of a model for the geochemical evolution of the basalts from the region. The implications of this model to magmatism at mid-ocean ridges is also considered.

## **1.5)** Summary

Iceland results from the interaction of a hot mantle plume with a mid-ocean ridge plate boundary. The latter is expressed on Iceland in the form of axial volcanic zones and areas analogous to transform zones. The axial volcanic zone is segmented in the form of volcanic systems that predominantly contain central volcanoes. Here, the lithosphere is 8-10 km thick and it becomes progressively thicker off-axis.

The Icelandic extrusive igneous crust ranges in age from 16 Ma to Recent, and has been produced in diverse eruptive environments. It has subsequently been subjected to variable erosional processes. The recently erupted products form two morphologically distinct volcanic features, conical and elongate. These represent two types of eruption event, lava shield and fissure eruptions.

The study area is composed of three volcanic systems (Hengill, Hrómundartindur, and Hveragerði) and contains the active Hengill and the extinct Grensdalur central volcanoes. There are examples of lava shields and fissure eruptions produced during subaerial and subglacial periods. The region is a ridge-ridge-transform triple junction, and the transform branch, the SISZ, has influenced the tectonic lineaments of the spreading zone. The crust is <1 Ma and 7-10 km thick at the spreading axis, with active geothermal manifestations. The density structure of the crust comprises high and low density areas which correlated with surface tectonic and geological features, such as the volcanic systems, the central volcanoes and lava shields.

The primary objective of the research was to establish the volcanic history and geochemical evolution of the area. This has been achieved by the production of a geological map and the chemical analysis of approximately 450 samples. The observations made are compared to those on the Reykjanes Ridge and other submarine spreading ridges.

# Chapter 2

# Icelandic Volcanology

# **2.1) Introduction**

In this chapter, the eruptive products and their relationships, as observed in the field in Iceland, are described and illustrated in some detail. An understanding of such relationships is essential for construction of the geological map, and for a comparison with marine eruptive products. The study area is dominantly volcanic in origin consisting of both primary and reworked eruptive material in various forms. Intrusive igneous material in the form of dykes and sills also occurs.

# 2.2) Primary igneous lithologies

# 2.2.1) Extrusives

#### 2.2.1.1) Hyaloclastite deposits

Icelandic glacial ridges (tindas) and flat-topped mountains (table mountains or tuyas) have a generalised constructional model that has been established from field observations in Iceland and elsewhere (Mathews, 1947; Sæmundsson, 1967; Sigvaldason, 1968; Jones, 1968). The constituents are collectively referred to as the "hyaloclastite formation" (or moberg formation) (Figures 2.1 & 2.2). A subglacial eruptive origin for this formation is widely accepted (Kjartansson, 1943; Mathews, 1947; Jones, 1968). After a subglacial eruption under Vatnajökull in 1934, Nielsen (1936) suggested that the variations observed within a single hyaloclastite formation can be explained as a complicated interplay between magma, water, ice, existing deposits and the landscape.

The prefix "hyalo" means "glass" and "clast" means "fragment". A hyaloclastite deposit should therefore, by definition, contain a high percentage of volcanic glass fragments (Cas & Wright, 1987). The deposits are neither pyroclastic

nor epiclastic in origin, but are produced by the quench-fragmentation of magma exposed to water (Mills, 1984; Kokelaar, 1986). Hyaloclastite is used as a genetic term for a fragmental glassy aggregate where such an origin has been established (Rittmann, 1962; Pichler, 1965; Honnorez & Kirst, 1975; Cas & Wright, 1987).



Figure 2.1: Model of the hyaloclastite formation: a) simplified example; b) longitudinal section through a hyaloclastite ridge, illustrating some of the more complex relationships that are observed; c) a cross section through (b).



Figure 2.2: An example of an edifice that typifies the hyaloclastite formation, from the EVZ in south Iceland, at Skógar. The edifice has 400 m of relief, above sea-level, but only the upper 200 m are visible in this photograph.



Figure 2.3: Photomicrographs of hyaloclastite tuffaceous matrix: a) basaltic fragments and abundant crystals (plain polarised light - PPL); b) angular fragments of glass (sideromelane) with altered glass (palagonite) in the interstices (PPL). The large vesicles are still intact and have not exploded. Field of view in both is approximately 15 mm.

In the field, hyaloclastite is orange-tan-brown in colour (representing the palagonite constituent) with additional black unaltered fragments of glass (referred to as sideromelane) (Sigvaldason, 1968). The shapes of the glassy fragments are dominantly blocky or splintery with sharp edges and corners, and planar margins (Figure 2.3); collectively these are thought to characterise the quench-fragmental

origin. Other possible ways of producing glassy fragmental volcanic rocks are by explosive volcanism in volatile-rich magmas, but the glass fragments produced would be expected to be Y-shaped from exploded vesicles (Honnorez & Kirst, 1975). No such texture is observed in the hyaloclastite glass in Figure 2.3, and the vesicles are still intact. In addition to this hyaloclastite matrix, there can be any number and size of glassy and basaltic clasts. The deposit, as a whole, is usually compositionally homogeneous (with few exceptions), and may be very crystal-rich if the magma was porphyritic (Figure 2.3).



Figure 2.4: Field photographs of the constituents of a hyaloclastite edifice: a) pillow lavas and hyaloclastite breccias grading outwards into hyaloclastite tuffs; b) a clast-rich variety of hyaloclastite breccia; c) a clast-poor variety of hyaloclastite breccia.

The constituents of the hyaloclastite formation are illustrated in Figure 2.1, and exemplified in Figure 2.2. In the simplest example, the core of the edifice contains a feeder zone (Section 2.2.3.1) that grades laterally into pillow lavas. These subaqueous deposits are blanketed by hyaloclastite lithologies (Figure 2.4). A gradation from hyaloclastite pillow breccias to bedded hyaloclastite tuffs occurs with distance from the feeder zone (Figures 2.2 & 2.4a).

**Pillow breccias** consist of a hyaloclastite matrix containing occasional whole pillows, pillow fragments with glassy rinds still attached, and basalt blocks without glassy rinds. **Hyaloclastite breccias** consist of basalt blocks that tend to lack glassy pillow rinds, but still with abundant glass within the tuffaceous matrix. A gradual decrease in the number of basalt blocks in the matrix leads to the end member known as **hyaloclastite tuff**. This may be bedded peripherally with a dip of up to 30° (Sæmundsson, 1979). The hyaloclastite components are all gradational with each other.

#### 2.2.1.2) Subaqueous sheet flows

Subaqueous sheet flows were first identified in the South Pacific by Bonatti (1967), and later by Bellaich *et al.* (1974) in the FAMOUS area of the Mid-Atlantic Ridge (MAR), but are generally more abundant on fast-spreading and intermediate-spreading ridges (Ballard *et al.*, 1979). The term "sheet flow" originally described the surface form of a submarine sheet-like structure where the origin and thickness were unknown. It was intended as a provisional term only to be replaced when more information on the structure was known (Ballard *et al.*, 1979). The term "sheet flow" is now applied to sheet-like structures of subaqueous eruptive origin.

The interiors of sheet flows, as observed in onland examples (e.g. Schmincke & Bednarz, 1990), are usually massive with vertical or partly radial columnar jointing, but the upper and lower surfaces show much variation. Numerous surface textures occur, of which several may be present on a single sheet. These include lobate, smooth, rippled, wrinkly, hackey, jumbled, pillowed and brecciated textures (Ballard *et al.*, 1979; Cas & Wright, 1987). Despite this diversity, two broad divisions into simple and complex varieties are recognised by vertical profiles through the interiors (Figure 2.5a) (Cas & Wright, 1987).

Sheet-like bodies occur towards the base of the Grensdalur volcanic pile (Figures 2.5b & 2.6). They each have similar internal structure, regardless of their

thicknesses, which range from 1-30 m. They lack the usual volcanological features characteristic of subaerial lava-flows (as summarised in Figure 2.7), or the chilled surfaces typical of intrusive sills. They are sheet flows of the simple type, with folds, whorls, and poorly developed pillow structures at the upper surfaces, and chilled, structureless lower margins. Sheet flows are identified elsewhere in Iceland, e.g. in N Iceland (Sigvaldason, 1968).



Sharp chilled basal contact



#### Icelandic Volcanology



Figure 2.6: Sheet flows of variable thickness, but with similar internal structures (illustrated in Figure 2.5b). These flows are exposed on the west bank at the mouth of the Grænsdalur river.

## 2.2.1.3) Subaerial lavas

Within the study region, subaerial lavas have been erupted at various times, including the Postglacial period (<9,000-13,000 years), interglacial periods (intervals during the Upper Pleistocene) and occasionally during glacial periods when an eruption breached the surface of glacial melt-water. In the field, the Postglacial lavas are recognised by such features as: a red/purple tinge to the rocks; under-surfaces with chilled drips hanging downwards; ropy flow texture; an abundance of scoriaceous cinder-like rubble. All of these features are seen in the Postglacial lavas of the Hengill region (e.g. Nesjavellir) (Figure 2.7), and they may be used to identify older subaerial lavas. The latter are generally more compacted and usually with any loose rubbly surface either lithified or absent.

There are two types of lava flow observed in the Postglacial and older deposits. These are the pahoehoe (or smooth) and the aa (or blocky) varieties (G. P. L. Walker, 1971; Cas & Wright, 1987). These are schematically illustrated in Figure 2.8. The pahoehoe lava is smooth, often with ropy texture on the upper surface (Figure 2.7d). Internally, it has a well defined base, with or without pipe vesicles, and the upper portion of the lava has undeformed circular vesicles. By contrast, the aa lava is blocky, often with rubbly and frothy upper and lower surfaces. Internally it is massive, usually with columnar jointing, and with much contortion of the vesicles and other flow direction indicators.



Figure 2.7: Collection of field photographs showing textures that characterise subaerial lava flows in Iceland: a) boulder of spatter agglomerate; b) relict spatter agglomerate vent; c) underside of subaerial lava flow that has chilled as the surface was in the process of dripping; d) subaerial ropy texture on surface of a subaerial lava flow. The photograph in (b) is from the Eldgjá region (EVZ). The photograph in (d) is from north of the lake Thingvallavatn and was kindly donated by P. Field.

The term "flow-unit" was introduced to describe lava flows in terms of cooling units (Nichols, 1936). It is defined as a body of lava that has visible signs of cooling on the upper and lower surfaces. Compound lava refers to lava that is divisible into flow-units. Simple lava refers to one that is not (G. P. L. Walker, 1971). This means that if an eruption produces more than one lava flow (or flow-unit) they are referred to as compound, and if it produces only one, it is simple. Compound lava is the most common type in the study area.





# 2.2.2) Intrusives

## 2.2.2.1) Dykes

One of the criteria used to identify a central volcano in eroded Icelandic crust is a local increase in the percentage of intrusives (K. Sæmundsson, pers. comm. 1988) (Section 1.1). During this research, 270 dykes were mapped in a 72 km<sup>2</sup> area, and 76% of these outcrop in the Grensdalur Central Volcano (Figure 2.9). There is another concentration of dykes (and an area of hydrothermal alteration) at Grafningsháls, and this may represent another volcanic centre (Figure 2.9).



Figure 2.9: Map illustrating the location of dykes and a solidified lava lake within the region. The lava lake is labelled accordingly.

#### Icelandic Volcanology

On a regional scale, the petrological variation of the dykes closely follows that of the eruptive units that they intrude, and is particularly great within the Grensdalur Central Volcano. The dykes mainly occur as isolated, <1 m wide and <500 m long, curvi-linear single dykes, and they may occur in swarms less than 100 m across, trending parallel to the regional dyke trend (036°). Composite dykes, where one intrudes the centre of another, are rare. Thinner and shorter dykes composed entirely of basaltic glass are abundant in the hyaloclastite deposits. The glassy nature may indicate that they are approximately contemporaneous with the material they intrude, if it had been water-saturated at the time. However, the glassy dykes analysed here are not of the same composition as the hyaloclastite fragments of the breccia, but are andesites (Figure 2.10).



Figure 2.10: A glassy dyke intruding moderately altered hyaloclastite breccias within the Tindar high of the Grensdalur Central Volcano. The composition of the dyke is drastically different from that of the surrounding hyaloclastite, and is one of the most evolved rocks (andesites) collected during this research.

The dykes in the area may be divided into four main types (A-D), based on field and petrological character. These types are illustrated and explained in Figure 2.11. Approximately 90% of the dykes exposed within the Grensdalur Central Volcano are of one of these four types. However, there is no consistent cross-cutting relationship between these types. In the central volcano, the orientation of the dykes
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of types A and B deviate the most from the regional fabric direction (036°), but no clearly defined radial arrangement is observed amongst the dykes within the central volcano.



Figure 2.11: Four types of dyke observed in the Grensdalur Central volcano.

## 2.2.2.2) Sills

Very few sills outcrop in the region. In the Grensdalur Central Volcano, examples include single sills that intrude concordantly between eruptive units and horizontal, multiple sills that intrude disconcordantly through eruptive units and sediments. Subaerial lavas of the same type of basalt that forms the sills can sometimes be found overlying them (Figures 2.12 & 2.13), but this is not always the case. At Ingolfsfjall, two sills intrude a thick pile of glacial sediment. The upper one is petrographically similar to the basalt of the Ingolfsfjall lava shield, whereas the lower sill is different (Appendix 1D; Figure 1D.2). The former was emplaced within 5 m of the base of the lava shield.

In general, the sills in the area all appear to have intruded at a very shallow level in the crustal pile and close to the surface where, in some instances, there is erupted lava of similar composition. In other areas, the contemporaneous nature of the sills and extrusives is not clear because of the lack of petrological or geochemical diversity in that area.



Figure 2.12: Transgressive sills outcropping in the hillside in the Djúpagil valley; a) Field photograph; b) line-drawing of a), indicating the various eruptive units there (see also Figure 2.13).

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Figure 2.13: Sketch map of the Djúpagil valley, within the Grensdalur Volcano, where multiple transgressive sills are intruded through a series of older hyaloclastite units and sedimentary rocks. The lavas that onlap the sediments are of the same type of basalt as the sills. This map includes the area shown in Figure 2.12.

#### 2.2.3) Pseudo-intrusives

## 2.2.3.1) "Feeders" of the Hyaloclastite Formations

Feeders are not hyaloclastite but are a component of the hyaloclastite formation (Section 2.2.1.1). They are intermediate between an intrusive and extrusive, because they are often surrounded by material derived from themselves or nearby companion feeders. They occur as variably sized and shaped bodies, characteristically displaying radial columnar jointing which becomes less pronounced towards the margins, where vesicularity and flow texture may be developed (Figure 2.14). There are no sharp enclosing boundaries but, instead, feeders grade into one of the hyaloclastite lithologies (usually pillow lavas or pillow breccias) (Figure 2.15). There is no grainsize difference between the basalt in the feeders and the basalt of the pillow lavas and hyaloclastite breccias. The similarity in internal structure between the feeders and the sheet flows may give rise to mis-identification when continuity of outcrop is poor. Icelandic Volcanology



Figure 2.14: Field photograph of a circular feeder displaying radial columnar jointing. The feeder is approximately 3 m in diameter. (Photograph kindly donated by G. Friðleifsson, Orkustofnun).



Figure 2.15: Field photograph of a feeder tube (upper left hand-side of the photo) and its gradational contact with the contemporaneous pillow lavas (right). The outcrop is approximately 2.5 m high. (Photograph kindly donated by G. Friðleifsson, Orkustofnun).

2.2.3.2) "Lava Lake" lithology

On the top of the Ingolfsfjall subglacial table mountain there are scattered exposures of gabbro of variable grainsize (Figure 2.16) that collectively outline an elliptically-shaped outcrop (1.2 km, with an aspect ratio in plan = 1.74) (Figure 2.9). The gabbro exposures are not located at the highest point of the mountain, or in the centre. Adjacent to the gabbro body is a relict subaerial vent indicated by an accumulation of spatter agglomerate. The position of the gabbro outcrop high in the eruptive pile and adjacent to well preserved subaerial vent structure suggests that it is not an intrusion exposed by erosion, but that it has acquired its coarse grainsize by slow cooling, as might occur in a lava lake. Some of the Postglacial lava shields are reported to have summit lava lakes (Rutten, 1964; Sæmundsson, 1992). In Hawaii, cooling lava lakes are monitored and sampled (Heltz, 1987) and coarse gabbro layers that are almost identical in description to those observed on the Ingolfsfjall mountain have been reported (Heltz *et al.*, 1989).





Figure 2.16: Photomicrograph of gabbro that outcrops on the Ingolfsfjall table mountain; a) PPL; b) XP. The field of view is 18 mm.



Figure 2.16: Continued.

# 2.3) Reworked igneous lithologies

## 2.3.1) Undifferentiated sediment

Reworked volcanic deposits may be recognised by the presence of hyaloclastite breccia and tuffaceous lithic clasts, rounded basalt clasts, and a reduction in fresh glassy fragments in the matrix. The constituents of the sediments, such as basalt clast type and phenocryst content within the matrix, may be used to identify the source of the material in terms of particular eruptive units in the surrounding region. Some eruptive units with exceptionally distinct mineral assemblages erode to produce sedimentary layers that may be used as marker horizons over a broad region.

Sediments accumulate along the flanks of the central volcanoes, and often the vertical variation in detritus within these piles can be used as a guide to the stratigraphy of the surrounding regions. Some sediment traps appear to have persisted for considerable periods (tens of thousands of years), spanning numerous eruptions, as indicated by the diversity in layered detritus through a single section. In this research, logging such sections in the stratigraphy of the Grensdalur Volcano and the

adjacent Reykjafjall region has enabled correlation of the stratigraphic units there (Appendices 1A & 1B).

## 2.3.2) Glacial deposits

By definition, **Tillite** is synonymous with "boulder clay" and is a variety of drift deposit (Whitten & Brooks, 1972). Two types exist: the fluvio-glacial variety that is water-lain; and the material dumped by the retreat of a glacier or icesheet. The former is likely to be stratified with some degree of sorting, and the latter unstratified and poorly-sorted (Whitten & Brooks, 1972). In Iceland, a distal moraine is considered to be poorly-sorted, inhomogeneous, and to have a distinctive grey matrix which characteristically lacks abundant glass fragments (K. Sæmundsson pers. comm., 1991) (Figure 2.17). Difficulties in field mapping may arise when a hyaloclastite grades laterally into reworked glacial material. This may occur during an eruption when melt-water at the periphery of the hyaloclastite edifice mobilizes existing unconsolidated tillite moraine and mixes it with freshly erupted hyaloclastite products (Geirsdóttir, 1991).



Figure 2.17: Field photograph of glacial moraine in south Iceland, at the foot of Vatnajökull (Skaftafell). Note the grey coloured rock-flour matrix that supports the varied and poorly-sorted clasts.

Tillites in the study area have sparse, matrix-supported clasts of variable angularity, colour and lithology. There are frequently laminations in the matrix, but unstratified tillites also occur. In this thesis, a reworked deposit with a dull grey matrix is described as tillite, and this implies that it originated near a glacier. Difficulty in identifying tillites occasionally arose because of their similarity to / sediments, but they were easily distinguished from hyaloclastite deposits. The tillites are valuable in that they separate successive eruptive units, providing marker horizons in much the same way as distinctive sedimentary layers.

# 2.4) Lithological relationships

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In Iceland, the environment of volcanic eruption and sedimentary deposition governs the lithological relationships. In the study area, subglacial hyaloclastite deposits dominate the volcanic succession and are intercalated with subaerial deposits. In the following section, the *relationship* of the various lithologies and their *processes of formation* are described.

The relationship between the constituents of hyaloclastite formations is complex and rarely as simple as the model in Figure 2.1 suggests. The relative proportions of the deposits may vary and one or more of the constituents may be absent. For example, the feeder zone may grade directly into a hyaloclastite breccia (Figure 2.14), or the edifice may lack hyaloclastite breccias and consist only of feeders, pillow lavas and hyaloclastite tuffs. Similarly, a resistant cap of subaerial lavas (if present) may vary in thickness and distribution and there may be subaqueous sheet flows at the base of the succession.

Unconformities between hyaloclastite units may easily be seen when angular (Figure 2.18), but they are harder to detect when concordant. The situation is further complicated by minor unconformities within a single hyaloclastite formation (i.e. a single eruption). This type of unconformity may be produced by current action, fluctuations in the depositional environment, and by magmatic pulses. These produce discontinuities in the facies that are most pronounced in the distal parts of the formation. They may be difficult to distinguish from true unconformities between two separate eruptions with hyaloclastite of similar overall appearance. However, in the case of a true unconformity, erosion, weathering, or sediment deposition is expected.

The large volume lava shields (e.g. Ingolfsfjall) often have thick hyaloclastite breccia beds (with or without lavas) of almost identical basalt. These are directly superimposed with no sediment or erosion between them, and they are likely to represent multiple magmatic pulses during the one eruption event, coupled with fluctuations in the ice water level.

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When examining the field relationships of the hyaloclastite formations it is important to consider what sequence of events occurred under the ice that resulted in the observed geology. Witnessed eruptions and field observations suggest the following model (Nielsen, 1936; Sæmundsson, 1967; Jones, 1968). During a subglacial eruption, magma extrudes from a vent and melts the lower layers of ice to form a subglacial lake. This enlarges as the eruption continues, and the melt-water causes the magma to form subaqueous deposits (pillow lavas, sheet flows and hyaloclastites). The volume of ice melted by the magma is calculated to be sufficient to provide room for the growing volcanic pile (Allen, 1980). The dispersion of the erupted material is therefore not necessarily restricted by the icecave walls, at least, not at first. If the cavern becomes drained of melt-water, the environment becomes subaerial, and then the shape of the volcanic edifice will differ from a purely subaqueous one.





Figure 2.18: b) and d) field photographs of unconformities between two older hyaloclastite units (labelled 2 & 3) and a younger Postglacial lava flow (labelled 1); a) and c) are line drawings of (b) and (d) respectively. (Photographs are from the Hrómundartindur Volcanic System and are kindly donated by E. Postlethwaite).

Field observations show that a feeder zone of variable dimensions may form around the core of the edifice. The feeder grades into pillow lavas - the "spalling off" process - and the pillow lavas grade into hyaloclastite pillow breccias. These may form by: i) the pillow lava flow encountering a steep slope, causing discontinuity in the pillows (Sigvaldason, 1968); ii) by gravitational collapse down-slope in varying states of cooling and crystallization (Jones, 1970; Lonsdale & Batiza, 1980). Pillow fragments are likely to result from the latter process, and isolated pillows from the former. Both processes probably occur during an eruption. Granulation of pillow lava rind adds to the hyaloclastite matrix. As the edifice grows, the feeder zone may drive the pillow flow-front through previously erupted hyaloclastite debris so that it intrudes hyaloclastite material that has only just been erupted. Abundant melt-water ensures frequent lateral and vertical facies changes, particularly around the periphery of the edifice. Fluctuations in the ice-water level will cause temporal and/or spatial alternation between subaerial and subglacial depositional environments. Glass-poor boulder breccia (or flow-foot breccia) form by the interaction of a subaerial lava with water as it transgresses from one environment to the other (Jones, 1970).

## **2.5)** Summary

The study area is composed of both primary and reworked igneous material. The extrusive primary igneous products include the subaqueous constituents of the hyaloclastite formation, sheet flows and subaerial lava flows. Individual compound lava sequences are composed of either aa or pahoehoe flow-units, but the simple lava is always composed of aa type flows. The intrusives occur as dykes and sills, and the reworked deposits include sediments and tillites. The relationships of the deposits are complex and provide information on the volcanic history of the region.



# **Chapter 3**

# Geology of the Hengill-Ingolfsfjall Region

# 3.1) Introduction

In this chapter, the methodology behind the mapping is described, followed by a description of some details and complications that exist, and were discovered, within the stratigraphy of the region. Radiometric age determinations were attempted for twelve samples from the area and the results are presented here (courtesy of R. A. Duncan, Oregon State University). Appendix 1 (in parts A-E) contains additional data and information used to constrain the stratigraphy, including petrographic and geochemical lines of evidence.

Eruptive products in the area date from the Late Tertiary (<1 Ma) to Recent in the Postglacial era (<10,000 years) (Sæmundsson & Einarsson, 1980). The crust produced during this period is made up of many individual eruptive units. Prior to this work, several geological maps of the study region existed (Table 1.1). These are various maps from Sæmundsson (1967); two published 1:25,000 maps, one of the Hengill mountain and the area to the north of it, and one of the adjacent Hrómundartindur region (Árnason *et al.*, 1986; 1987); and one unpublished map of the Reykjafjall-Ingolfsfjall region (Eiriksson, 1973). Therefore, it was of primary importance to this research to construct a map and establish the stratigraphy of the eruptive units in the Hveragerði to Ingolfsfjall area. This fills the gap in the previous map coverage of the region, and also reveals the existence, and spatial and temporal distribution of lava shields in the area.

From the geological map produced during this research the stratigraphy of the Hveragerði-Ingolfsfjall region has been established by tracing marker horizons within the area. These horizons divide the eruptive units into five main stratigraphic groups, which are, in some cases, subdivided further according to regions when lateral correlation between units of a group is not possible. The distribution of these groups is summarises in Figure 3.1. Within the Hveragerði Volcanic System, 81



Volcanic System

G sub-groups

T sub-groups

B sub-groups

Figure 3.1: Geological map of the Hveragerői Volcanic System with the main stratigraphic divisions. In the key, the groups are listed in approximate stratigraphic order, but a more detailed account of the stratigraphy is given in Figure 3.2. The location of the geological boundaries for the Hengill products in the SW of the region are from Sæmundsson (1967) and Pullinger (1991).





eruptive units have been mapped and these are detailed in Maps A1 and A2. The resulting stratigraphy is shown in the form of a chart in Figure 3.2. Oxygen isotope data, summarised by Harland *et al.* (1990), indicates that interglacial episodes occurred approximately every 100 ka during the last 1 Ma period. On deglaciation, tillite deposition occurs, and when two eruptive units are separated by a substantial tillite horizon, it is inferred that an interglacial episode has occurred in the interval between their formation. This constraint has been used here to position the eruptive units in Figure 3.2. Using the information in Figures 3.1, 3.2 and Maps A1-A2, the locus of magmatic activity can be traced within the region over the past 1 Ma.

## 3.2) Mapping

Mapping was carried out on the scale of 1:10,000 using standard "Greenline Mapping" (Barnes, 1981). Since all rock types exposed in the area are broadly basaltic, the usual lithological mapping approach proved unsuitable. Instead, the different "forms" of the basalt, as described in Chapter 2, were recorded. The resulting genetic map is a suitable means to record intricate field detail and relationships, but is a poor way to represent the overall geology and volcanic history of the area. From the information contained in the field maps, a second geological map was developed which illustrates the boundaries between successive eruptive units (Maps A1 & A2), and from which the volcanic stratigraphy of the area was derived (Figure 3.2).

Mapping eruptive units is greatly assisted where the landscape has suffered little erosion, like the Hengill and Hrómundartindur regions (Figure 1.5), since the topography is created by volcanism and can therefore be used as a guide to distinguishing individual eruptive units. In such regions, the lava shield and fissure eruptions are morphologically distinct. Unfortunately, in the Grensdalur Central Volcano, there has been pervasive hydrothermal activity resulting in highly altered basalt that is now considerably eroded, and in most of the Hveragerði Volcanic System glacial action and weathering have smoothed the volcanic landscape (Figure 1.5a).

In all cases, the most reliable distinguishing features of the different eruptive units are: i) the percentage and types of phenocrysts; ii) observed unconformities between units; iii) compositional differences between eruptive units; iv) marker tillite and other sedimentary horizons. In the absence of diverse phenocryst assemblages or non-eroded topography, other variations such as vesicularity, differential

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hydrothermal alteration and weathering had to be used. Aerial photography, supported by ground-level examination, was valuable for detecting subtle variations between successive units.

The vertical cliff sections exposed in the region, such as those in the walls of the Grensdalur Volcano, on the south and west faces of Ingolfsfjall, and in the many deep gorges, provide opportunity for detailed sedimentological-type logging of the vertical variation within and between eruptive units (Appendix 1A). Subsequent correlation of the logs establishes the stratigraphy of the region (e.g. Appendix 1B). Thin section petrographic examinations revealed substantial differences between the eruptive units of the whole region. However, it was often found that the variation is minimal within a given region, and there may be as much petrographic variation in a single eruptive unit as there is in that particular region. Alternatively, an eruptive unit may be so petrologically distinct that thin section examination cannot advance stratigraphical understanding beyond that gained by field observation.

The sampling strategy for the geochemical work (3 samples per eruptive unit) enabled thorough coverage of the stratigraphic column. In general, it was found that the petrographically similar eruptive units in close spatial and temporal proximity tend to be of very similar composition. However, some geochemically distinct units were discovered that could be used as marker horizons (Appendix 1C). Such units were invaluable in the construction of the geological map and stratigraphy of the study area. In conclusion, all the techniques described above are necessary for the best interpretation of the geology of the area.

The principal of uniformitarianism, where the present is the key to the past, was applied to the investigation of the eruptive units of the Hveragerði Volcanic System in order to attribute the units to either a lava shield or a fissure type of eruption. A similar approach has been applied to areas of old crust in Iceland (Rutten, 1964) and elsewhere (Noe-Nygaard, 1968). In the present neovolcanic zone, the morphologically distinct volcanic features of lava shield and fissure display consistent differences in their lava types; simple, compound, aa and pahoehoe (Section 2.2.1.3) (Rutten, 1964; G. P. L. Walker, 1971; Sæmundsson, 1979).

The Postglacial subaerial lava shields are built up from compound pahoehoe flow-units that are individually usually thin (1-10 m), and collectively they take on the form of a low-angled dome (1-10°). No aa lavas are known from shields in Iceland (G. P. L. Walker, 1971). The young subglacial lava shields form table

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mountains that are laterally extensive in more than one direction, are usually thick and, in the study region, are observed to form "beds" or sub-units. The Postglacial subaerial fissure eruptions produce either simple or locally compound lavas which are dominantly of the aa variety. Individual lavas may flow considerable distances and, despite the locally compound nature of the flows, they do not form well-defined, lowangled domes (G. P. L. Walker, 1971) (Figure 1.3, p. 6). The young subglacial variety of fissure eruptions form hyaloclastite ridges that are laterally restricted in one dimension, and they are usually thinner than subglacial lava shields, with bedding restricted to the tuffs in the periphery of the deposit. In the study area, buried smallvolume morphologically distinct subglacial lava shields were not observed. It is envisaged that, with limited exposure, it would be very difficult to distinguish conclusively between buried subglacial small-volume shields and buried subglacial fissure eruptions.

# 3.3) Stratigraphy of the Hveragerði Volcanic System

## 3.3.1) Marker horizons

The stratigraphy of the Hveragerði Volcanic System has been constrained by six marker horizons. These are described below, starting with the oldest first. Their location within the field is illustrated here with maps and field photographs (Figures 3.3-3.9), and further details on these markers are provided in Appendix 1 (A-D).

## Ingolfsfjall Marker 1 [IM1]

This marker consists of bedded tillite and other sediments that outcrop on the south face of Ingolfsfjall (Figures 3.3 & 3.4b-d). It appears orange at a distance but, apart from this, it does not have any distinguishing features and only forms a marker because it is laterally continuous. On the south face of Ingolfsfjall, eruptive units below this sedimentary horizon belong to the Jaramillo subchron (Kristjánsson *et al.*, 1988) indicating that they are approximately 900 ka (Harland *et al.*, 1990). The bedded nature, thickness and variety of sediments indicates that they represent deposition over a considerable period of time (at least 150 ka). It is possible that the sediments observed on the west face of Ingolfsfjall and at Grafningsháls may also belong to the IM1 marker (Figures 3.3 & 3.4b-d).

## Brunhes-Matuyama boundary (700 ka)

This palaeomagnetic reversal boundary has been located in the area by Sæmundsson & Einarsson (1980) and Kristjánsson *et al.* (1988). It is indicated in



Figure 3.3: Stratigraphic logs of the Sognar to Ingolfsfjall region detailing eruptive units of the Gljúfur and Bjarnarfell groups. The Ingolfsfjall marker horizons (IM1 & IM2) and the Brunhes Matuyama reversal boundary (700 ka: Sæmundsson & Einarsson, 1980; Kristjánsson et al., 1988) are indicated in the logs The location of these logs is provided in the map, and more detailed accounts of these logs are given in Appendix 1A.



b)

a)



Figure 3.4: Geology of the western and southern flanks of the Ingolfsfjall table mountain with accompanying line drawings, on which the geological information, in terms of eruptive units and marker horizons, is displayed. For the locations of these profiles, see Figure 3.3. The details of each profile are given in Appendix 1A: a) profile on the west face of the mountain (referred to as IF); b) profile on the west face (referred to as IL); c) profile on the south face (referred to as IU); d) profile on the south face (referred to as IX). (See also Appendix 1D; Figure 1D.2)

Figure 3.4: continued.





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Figures 3.3 and 3.4. This boundary provides some constraint on the absolute age of the volcanic products and also indicates that the volcanic products are not systematically younger towards the active zone in the west (Figure 3.2). For example, Ingolfsfjall is within the same magnetic reversal period as the last eruptive units from the Grensdalur Central Volcano, indicating that they could have been contemporaneous (approximately at 300 ka; K. Sæmundsson pers. comm., 1991). This means that the volcanically active zone could have been over 10 km wide at that time, or that Ingolfsfjall is an off-axis volcano.

## Ingolfsfjall Marker 2 [IM2]

This marker consists of tillite which outcrops at the base of the Ingolfsfjall lava shield (Figures 3.3 & 3.4). It is of a reasonably constant thickness (usually <5 m) and, like IM1, is traced by lateral continuity. This separates the Ingolfsfjall lava shield from the underlying eruptive units which, in some cases, are of very similar basalt (e.g unit BA3; Figure 3.4a).

## Sauða lava shield [eruptive unit DA1]

This is the lowest marker within the Grensdalur Central Volcano. It occurs in the Sauða valley, forming a laterally continuous plateau within the core of the eroded volcano (Figures 3.5-3.7). The base is composed of subaqueous pillow lavas which are overlain by subaerial pahoehoe compound lava (up to 8 flow-units). The latter is laterally extensive and can be traced to the southwest of Dalafell at Djúpagil (Figures 3.5 & 3.6). It is possible that the outcrops here at Djúpagil may belong to another eruptive unit further south (unit DB7b) (Sæmundsson pers. comm., 1992) but the evidence gathered during the course of this research favours the former interpretation. A small outcrop of similar basalt occurs on the east side of the volcano near Reykjafjall where it is exposed by the landslide (Figures 3.5 & 3.8). The laterally extensive nature of the DA1 unit and the small succession of compound pahoehoe lavas suggest that it may have been a lava shield-type eruptive event.

## Reykir lavas [eruptive unit S1]

The compound lavas of aa type that belong to this marker horizon are not entirely laterally continuous in the field. They are similar in handspecimen and thin section to other lavas in the Grensdalur Central Volcano. However, samples from the S1 unit are compositionally similar to each other and dissimilar to other eruptive units with which they are intercalated (Appendix 1C). The core of the S1 unit occurs in the NE corner of the Grensdalur Volcano in the flanks of Selfjall (Figure 3.5), where it occurs as hyaloclastite breccias with large feeder sheets and tubes. These deposits are compositionally similar to some lava flows that occur between Klóarfjall and Álútur, (up to 6 aa flows) (Figure 3.9). Still further west, the lava flows appear to extend into the west flank of Klóarfjall (up to 3 aa flows). However, the composition of these three lavas indicates that this is not the case, and only the top flow of the three is a continuation of the six aa lavas of the S1 marker (Figure 3.9 & Appendix 1C).



Figure 3.5: Geological map illustrating the location of the three marker horizons within the Grensdalur Central Volcano. The older Gljúfur group and the younger Hengill and Hrómundartindur eruptive products are also indicated. The positions of the boundaries of the other Grensdalur eruptive units are illustrated for reference (see Map A1). The location of the stratigraphic logs in Figure 3.6 are indicated, and log titles are underlined.





Figure 3.6: Stratigraphic logs from the area of the Grensdalur Central Volcano, indicating the locations of the three stratigraphic marker horizons. The key for these logs is the same as in Figure 3.5. The positions of the logs are indicated in Figure 3.5, and further details on the logs are provided in Appendix 1A.

Elsewhere, in the cliff to the west of Reykjafjall (Reykir cliff with profile Reykir 1; Figures 3.5 & 3.6), there is a relict valley that has preserved 7 aa lava flows intercalated with sediment and tuffaceous horizons which dip to the NW (225/025°) (Appendix 1A & B). The lavas are all petrographically similar, but the six upper lavas can be compositionally matched with the S1 marker unit. The lowest lava (of unit TB6; Figure 3.6), separated from those above by a sedimentary layer, is dissimilar in composition to the S1 marker (Appendix 1B & C). This S1 marker unit enables the stratigraphy of the western and eastern halves of the volcano to be correlated, but this would not have been possible without the geochemical analyses of these basalts. Geology



Figure 3.7: Field photograph of the Grænsdalur valley and the Sauða plateau. The location of the DA1 plateau is labelled. (Photograph kindly donated by Omar Friðleifsson).



Figure 3.8: The Reykjaffall cliff showing a landslide and the location of an outcrop of DA1 marker basalt. This small outcrop provides a constraint on the age of the Reykjaffall region relative to the rest of the central volcano. (Photograph kindly donated by Omar Friðleifsson).

## Selfjall deposits [eruptive unit S12]

This eruptive unit occurs on the Selfjall peak, and sediments derived from it occur to the west at Klóarfjall and to the south at Reykjafjall (Figures 3.5, 3.6 & 3.9). The unit is petrographically distinct because it is highly porphyritic and the only one of its type in the Grensdalur area (with abundant olivine, plagioclase and pyroxene phenocrysts). Sediments produced nearby during and/or after this eruption are distinctive in the same way with clasts of this basalt-type and crystals of all three phases present in the matrix of the sediment. Although the clast-type in the sediments



Figure 3.9: The Klóarfjall-Álútur area: a) field photograph (kindly donated by Omar Friðleifsson); b) line drawing of a) indicating the geology of the area. The lava flows and eruptive units discused in the text are indicated.

is mixed, the S12 basalt is by far the dominant constituent. Many of the sediments are water-lain, and show features such as stratification and relict channel structures with steeply inclined sides (varied between 20-40°). These suggest that the deposits were dumped from water travelling via rivers through existing topography, and indicates that the Selfjall eruption (unit S12) occurred late in the history of the central volcano.

#### 3.3.2) Stratigraphic Groups

## 3.3.2.1) The Gljúfur Group [GA, GB and GC sub-groups]

This group contains 18 eruptive units, all of which are of reversed magnetic polarity (Matuyama epoch, >700 ka) (Sæmundsson & Einarsson, 1980; Kristjánsson *et al.*, 1988). The group is subdivided on a regional basis into sub-groups A-C: A for the region in the west (abbreviated to GA); B for the south face of Ingolfsfjall (abbreviated to GB); and C for the west face of Ingolfsfjall (abbreviated to GC) (Figure 3.1).

The GA sub-group comprises 12 eruptive units, all of which are essentially aphyric. This means that successive eruptive units can only be differentiated in the field by vesicularity, degree of weathering and alteration, and observed sedimentary horizons separating them. The terrain in this region is smoothed off by erosion and consists of undulating linear hills locally cut by deep gorges, believed to represent faults orientated dominantly N-S (Sæmundsson, 1967). The hyaloclastite units are of the order of 30-50 m thick and of limited lateral extent perpendicular to the regional spreading direction (099°). The area encompassed by this sub-group is thought to represent the site of repeated fissure eruptions.

The GB sub-group consists of 5 eruptive units that outcrop in the lower slopes on the south face of Ingolfsfjall (Figure 3.4c-d). Units (GB1-3) are exposed in a single profile (IX) (Figure 3.4d & Appendix 1A). Unit GB4 outcrops to the west, in profile IU, and forms a small peninsula (Appendix 1D; Figure 1D.2). It can be matched on compositional grounds with a pahoehoe lava to the north of Ingolfsfjall, on the NE flank, which is also of reversed magnetic polarity (Sæmundsson & Einarsson, 1980). The GB4 unit may represent a small-volume lava shield, because of the pahoehoe nature of this lava in the north. Overlying the thick IM1 marker tillite on the south face of Ingolfsfjall are the lavas that mark the Brunhes-Matuyama boundary (unit GB6) (Kristjánsson *et al.*, 1988). This apparently laterally continuous

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lava (Kristjánsson et al., 1988) is divided here into two flows (both aa type) based on petrographic and supporting compositional grounds (Appendix 1D).

The GC sub-group consists of only two eruptive units that are exposed on the west face of Ingolfsfjall (Figures 3.3 & 3.4a). They are thin hyaloclastite units similar to those of the GA sub-group, and are separated, and overlain by, sedimentary horizons. Absolute correlation with the GA or GB units is not possible. Units GB1 to GB4 are the oldest in the study area (Jaramillo subchron; Kristjánsson *et al.*, 1988: 900 ka; age after Harland *et al.*, 1990). Units of the GA, GC sub-groups and the GB6 unit maybe approximately contemporaneous in age (between 800-700 ka) (Figure 3.2).

3.3.2.2) The South Bjarnarfell Group [BA and BB sub-groups]

This group contains 8 eruptive units, (9 including the Ingolfsfjall lava shield eruption) all of which are of normal magnetic polarity (Brunhes Epoch, <700 ka) (Sæmundsson & Einarsson, 1980; Kristjánsson et al., 1988) (Figure 3.1-3.3). As with the previous group, they are divided according to spatial distribution into sub-group A, for units in the west (BA), and sub-group B for those exposed in the western and southern cliff faces of Ingolfsfjall (BB). One eruptive unit (BA3) can be tentatively traced between the two areas.

The BA sub-group consists of 3 eruptive units stacked on top of each other in the flanks of Bjarnarfell (Map A2). The top unit (BA3) occurs in four separate outcrops. Three of these can be grouped together here on a compositional and petrographic basis, and include the inliers at Sognar and South Bjarnarfell, and a thin veneer of hyaloclastite breccia at Reykir. The fourth outcrop is intercalated within the Ingolfsfjall stratigraphy at Grafningsháls and in the west face of Ingolfsfjall (Figures 3.3 & 3.4a-b). Here, the BA3 unit is a prominent resistant horizon bounded by IM2 above and sediment below, allowing it to be traced laterally into the Ingolfsfjall region. The BA3 unit, therefore, displays the characteristics of a laterally extensive subglacial lava shield-type eruption.

The BB sub-group includes five units that outcrop in the lower half of the Ingolfsfjall table mountain, below the main Ingolfsfjall lava shield (sixth unit), and in the Grafningsháls valley. Unit BB3 is the oldest and it is only exposed in the Grafningsháls valley (Figure 3.3). Unit BB4 occurs on both sides of Grafningsháls as hyaloclastite deposits, and also as a series of compound lavas on the west face of

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Ingolfsfjall (Figure 3.4b). Units BB5-7 occur as lavas in the more southerly flanks of Ingolfsfjall (Figures 3.4b-d). The lavas of these four eruptive units, BB4 to BB7, represent a lava sequence that is repeated twice here in the stratigraphic succession. In a single sequence, pahoehoe compound lavas (units BB4 & BB6) are overlain by one or two very fine-grained aa basalt flows (units BB5 & BB7). It may just be coincidence that they are intercalated with each other, or it may indicate that these units are related. The compound series of pahoehoe flow-units of BB4 & BB6 suggest lava shield eruptive events, but aa lavas (like those of BB5 and BB7) have not been observed in Recent lava shields (G. P. L. Walker, 1971). Alternatively, the aa lavas of BB5 and BB7 may have come from elsewhere. The top lava of the second sequence (BB7) represents the last unit to be erupted prior to the main Ingolfsfjall lava shield eruption.

3.3.2.3) The Tindar Group [TA, TB and TC sub-groups]

This is the first group considered to be part of the Grensdalur Central Volcano. All the units are normally magnetized (Sæmundsson & Einarsson, 1980; Kristjánsson et al., 1988), and they are all older than the DA1 marker horizon (Figure 3.2). The group has been divided into three sub-groups on the basis of a combination of age and regional distribution. None of the eruptive units possess features indicative of lava shield-type eruptions.

The TA sub-group is composed of 11 units, most of which represent individual eruption events, but some may represent more than one event. For example, the oldest unit (TA1) includes outcrops of basalt on the banks of the Varmá river (excluding Hengill postglacial lavas) (Figure 3.1, Map A1), and although these outcrops are probably not of the same eruptive unit, they have been grouped together because they represent basalt that occurs below the main Tindar sub-groups. In the north of the Grensdalur Volcano depression there is an intensely altered area that occupies the valleys of Sauða and Grænsdalur (Figure 3.7). In parts of this area, outcrop is minimal, with only resistant dykes flanked by coloured clays, forming steep linear ridges. Landslide deposits cover much of the area within the depression (Figure 3.5). The tuffs and basalts exposed in this region have been examined, and some poorly defined heterogeneity exists, but for the purpose of this research they have been grouped as one unit (TA2a). Sheet flows exist along the Sauða river that cannot easily be related to the altered TA2a basalt or the overlying Tindar units, and so they have been separated as TA2b (Map A1). The remainder of the TA units outcrop on the central Tindar high. Mapping of this region was severely hindered by pervasive smectite alteration affecting basaltic units that were initially very similar. Unconformities were mapped using a combination of variation in the character of the hyaloclastite, handspecimen petrological diversity (which is minimal), observed "apparent" unconformities in the field (usually from the degree of alteration) and observed unconformities in the form of sedimentary horizons. In addition, subtle compositional differences were identified. The Tindar high is composed of a combination of hyaloclastite formations, sheet flows and compound aa lava horizons. The lavas that cap Tindar (unit TA9), are similar to the basalt within unit TA10 at the southern foot of the Reykjafjall region (Map A1), but while they may be contemporaneous, direct correlation is not possible.

The TB sub-group occurs immediately to the east of the Tindar high in the flanks of Reykjafjall and the Reykir cliff (Map A1). The TB units that occur in the Reykir cliff have been correlated with each other, and with other units in the region, by thin, locally continuous and distinctive tillite marker horizons (Appendix 1B), as well as the DA1 marker to the north of the cliff (Figure 3.8). Unit TB6 is indistinguishable from unit TB1 in handspecimen but is compositionally distinct (Appendix 1C). The TB6 unit cannot be correlated directly with the DA1 marker, but it is estimated from the state of erosion of the TB6 eruptive unit and the number of tillite horizons present in the Reykir Cliff Profile that it is probably marginally older than the DA1 marker.

The TC sub-group consists of three eruptive units that outcrop as inliers in various locations on the Tindar region. They are younger than the TA units, but cannot be directly correlated with the TB sub-group or the rest of the Grensdalur stratigraphy. Unit TC1 occurs as a veneer of fresh, aphyric hyaloclastite and tuffs. Unit TC2 occurs in the same region, and is composed of fresh porphyritic hyaloclastite. The third unit, TC3, is situated on high ground and consists of scattered blocks of rock protruding through, or positioned on the clay scree. In thin section, the rock is completely replaced by hematite, which is reflected in its bulk composition (Appendix 1E). Vesicles, which display flow texture, suggest that the rock is eruptive material and not a primary alteration product. The colour and texture are similar to the Postglacial scoria observed in the Hengill region (Nesjavellir) and this suggests that it is of Postglacial origin (Figure 2.7). The location of this material suggests that other environmental elements during the Postglacial era could have transported the blocks

to their present location. The most abundant occurrence of blocks is on a gentle slope towards the Grensdalur river (Sæmundsson, pers comm., 1990) (Appendix 1E).

## 3.3.2.4) The Dalafell Group [DA and DB sub-groups]

All the eruptive units within this group are also of normal polarity (Sæmundsson & Einarsson, 1980), and the Group is divided into sub-groups on the basis of spatial distribution. The DA units occur in the eastern half of the Grensdalur Central Volcano, in the Sauða region, and the DB units occur to the west in the Dalafell and Klóarfjall regions (Figure 3.1 & Map A1). Other than the DA1 marker unit and one other (unit DB7b), none of the eruptive units represent lava shield eruptions and any lavas are of compound or simple aa type.

The DA sub-group is composed of 3 eruptive units. The lowest is the laterally extensive DA1 marker (Section 3.3.1). Unit DA2 is situated north of the Tindar high, on top of the plateau formed by DA1 (Figure 3.9 & Map A1). This is thought to be older than the DB eruptive products to the west because of the relative altitude of the units. However, the actual contact is not exposed. Unit DA3 occurs at Álútur and is directly overlain by the S1 marker. The contact between DB10 and this unit is not exposed (Figure 3.9 & Map A1).

The DB sub-group consists of twelve eruptive units, all erupted after the DA1 marker and before the S1 marker. The lower units are subglacial, whereas the stratigraphically higher ones are dominantly subaerial. The lavas dip towards the active volcanic zone in the NW, and sediments are observed within the sequence of eruptive units. The older unit (DB2) has a NE trending linear outcrop pattern, mimicking a ridge, that runs through the base of the Dalafell high (Map A1). It shows varying degrees of alteration, ranging from intense chlorite to very minor smectite alteration in the distal parts of the hyaloclastite mound. Many of the units are composed of hyaloclastite breccias that are glass poor, rich in small angular cream-coloured basalt clasts and have a minimal tuffaceous constituent, in which spindle bombs locally occur. These are indicative of very shallow water conditions. Unit DB8a shows a transgression between subaqueous and subaerial erupted products, whereas unit DB9, restricted to the southern regions of Dalafell (Map A1), is entirely subaerial.

To the south of the main road, in the flanks below the Skálafell lava shield at Núpafell, there are two eruptive units exposed in the cliff face (DB7b and DB8b) that





Figure 3.10: The cliff face of Núpafell, on the east side of the Skálafell mountain. The cliff is 100 m in relief. (See also Sæmundsson, 1967; p. 13).

were sampled during this research (Figures 3.1 & 3.10). The lower unit (DB7b) is composed of hyaloclastite beds and a series of compound pahoehoe lavas, and resembles a buried lava shield (Pullinger, 1991). The succession can be divided into three sub-units. The lower one (unit 1) is a horizontal bed of hyaloclastite (20 m thick), and the second (unit 2, of a similar thickness) dips to the south by 30-40° and is interpreted as a flow-foot breccia (Sæmundsson, 1967; Jones & Nelson, 1970; Pullinger, 1991). The top sub-unit (unit 3) comprises a series of compound pahoehoe lavas intercalated with thin hyaloclastite horizons. The overlying DB8b unit is composed of a thin veneer of hyaloclastite breccia. The geology of the area was studied in detail by Sæmundsson (1967) and Pullinger (1991) who observed that, to the north of this location, but still south of the main road (Road 1), unit DB7b and DB8b are separated by tillite and the DB8a aa lavas (Pullinger, 1991; Sæmundsson pers. comm., 1992). The tillite between units DB7b and DB8a suggests that there has been a deglaciation event between these two eruptions. The stratigraphic position of unit DB8b is poorly constrained with respect to the rest of the stratigraphy within the Grensdalur Central Volcano, and it could be considerably younger (Figure 3.2).

The basalt that forms unit DB10, to the west of Klóarfjall (Figure 3.9), can also be subdivided into three sub-units (DB10a-c) based on petrographic textural differences. The oldest and youngest sub-units of these are composed of course basalt with subophitic texture. The middle sub-unit (DB10b) is the most extensive and is a much finer grained basalt. These DB10 sub-units are all compositionally indistinguishable (Appendix 1C).

## 3.3.2.5) The Selfjall Group [S units]

This group consists of 17 eruptive units which occur mainly in the eastern half of the Grensdalur Central Volcano (Figure 3.1). The base of the group is marked by the S1 unit (Section 3.3.1). This, and the S12 marker, constrain the stratigraphy of the Selfjall to Klóarfjall region (Figures 3.5 & 3.6). The eruptive units in the Reykjafjall region are more loosely constrained by the S12 marker which, in the form of sediments and glacial tillites, could have been deposited at any time after the S12 eruption.

The absolute age relationship of units above the S12 marker is often nondeterminable, as many of the units occur as inliers (Map A1). Where this occurs, the freshness of the deposits and robustness of the topography have been used to infer the relative ages. Unit S17 is an extreme example of this (Figure 3.1), where the curvilinear ridge of very fresh hyaloclastite onlaps onto the units of the Dalafell high, but the unit cannot be correlated with the other S units of equally fresh appearance in the eastern half of the Grensdalur Central Volcano.

## 3.4) The Hengill and Hrómundartindur volcanic systems

The geology of the Hengill and Hrómundartindur volcanic systems has been described in detail by Árnason *et al.* (1986; 1987). Using the geological maps and stratigraphy from these authors, several specific eruptive units were selected for sampling and petrological investigation. The location of these units within the region and the stratigraphic column are summarised in Figures 3.11 and 3.12. Several of the younger eruptive units from the Hengill and Hrómundartindur volcanic systems onlap the units of the Grensdalur Central Volcano (Árnason *et al.*, 1987) (Figure 3.1). The following summary of the geology is from field observations made during sampling and mapping by the author, unless otherwise stated.

The first unit associated with volcanic activity in the Hengill Volcanic System are the Hæðir and Húsmúli lava shields (Árnason *et al.*, 1987) (see below). Drilling in the Nesjavellir valley indicates that there is another 600 m of eruptive material belonging to the Hengill Volcanic System (Franzon, 1988). The first unit reported to be associated with the Hrómundartindur Volcanic System is composed of aa lava flows in the Krókur gorge, and is referred to as the Stekkás grágryti basalts (Árnason *et al.*, 1987) (Figure 3.11). These basalts are petrographically and compositionally similar to those of unit DB10 from the Hveragerði Volcanic System (Appendix 1C), and it is suggested here that may belong the this eruptive unit.

Later eruptive units sampled here occur on either side of the Nesjavellir valley (Figure 3.11), and the youngest samples collected were from main Hengill eruptive unit (Figures 3.11 & 3.12). This eruption event produced about 6-8 km<sup>3</sup> of basalt that was erupted under subglacial conditions to form the Hengill mountain. There are some lava flows that cap the edifice (Árnason *et al.*, 1987). There are traces of hyaloclastite derived from the Hengill eruption to the north of the mountain that indicate that the initial phase of eruption was located on a fissure-type vent (Árnason *et al.*, 1987; Sæmundsson pers. comm., 1989). The morphology of this table mountain is similar to subglacial lava shields and may represent an eruption event of this sort (Figure 1.5a, p. 10). It was originally referred to as a large table mountain by Sæmundsson (1967), but has since been described as the Hengill Central Volcano (e.g. Hersir *et al.*, 1990). Within some of the units from the Hengill volcanic system, there is a heterogeneous accumulation of phenocryst phases in the basalt (e.g. the Lómatjarnarháls and Hengill eruptive units).





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Geology



Figure 3.12: Stratigraphic chart for the Hengill and Hrómundartindur Volcanic Systems (after Árnason et al., 1987). The eruptive units labelled are those that have been sampled and investigated during this research. The position of the other units in the area are marked as open squares.

## 3.5) Lava shield eruptive events; Ingolfsfjall, Hæðir and Húsmúli

The three lava shields from the study area are morphologically pronounced features that appear to be quite distinct. They are, in order of decreasing age, Ingolfsfjall, Hæðir and Húsmúli (Figure 1.5). They are described here to exemplify the geology of the lava shield type eruption events and to give some indication of the type of variation that may occur.

The Ingolfsfjall lava shield is the only shield that has been mapped in detail during this research. It is situated in the extreme east of the area, forming the upper 200-300 m of the steep-sided Ingolfsfjall table mountain (Figure 3.1 & Map A2). It covers an area of 25 km<sup>2</sup> and has an estimated volume of 5 km<sup>3</sup>. It is composed of subaqueous hyaloclastite deposits and subaerial compound pahoehoe lavas. The distribution of these hyaloclastites and lavas is illustrated Figure 3.13. Both volcanic products are unevenly distributed within the shield, but in general, hyaloclastite dominates the northern end of the mountain. The subaerial lavas can be subdivided into monotonous compound lavas and a single, particularly coarse lava exposed on the plateau of the table mountain (Figure 3.13). In addition, there is the ellipsoidal-shaped solidified lava lake and the nearby relict spatter agglomerate cone. There are some outcrops of tillite covering the deposits of the lava shield, both on the plateau and on the northern flanks.

Volcanic successions are well exposed on the vertical flanks of Ingolfsfjall (Figure 3.4, & Appendices 1A and 1D). On the south face, hyaloclastite pillow lavas are overlain by hyaloclastite breccias which give way to compound lavas (Figure 3.4b & d) which are, themselves, intercalated with thin hyaloclastite horizons (Appendix 1A). There are several linear dykes cutting the succession and abundant feeder tubes occur within the breccia beds, both of which are contemporaneous with the lava shield. There are no active or extinct geothermal manifestations within the shield, and the basalt is unaltered. The edifice is dissected by several non-linear faults that trend dominantly in a N-S direction and downthrow to the east (Figure 3.13). It should be noted that fault planes are not evident in the field, and that these faults have been mapped from aerial photographs.

The absolute height at which the base of the Ingolfsfjall shield occurs (i.e. top of IM2) varies from 100-275 m and is thought to indicate the undulating nature of the topography prior to eruption of the lava shield. The NE corner of the shield was
topographically lower than the southern end, and the highest point, prior to eruption, was at the NW corner at Grafningsháls.



Figure 3.13: Geological map of the Ingolfsfjall lava shield. The location of the IM2 marker horizon is indicated.

The Hæðir lava shield is situated to the NW of the Hengill Volcanic System and forms a flat, partly dissected dome (Figure 1.5). It has a radius of 3.4 km measured from the summit to the edge of the topographic expression to the west. The lavas cover an area of 36 km<sup>2</sup> and has an estimated volume of  $1.15 \text{ km}^3$  of lava exposed at the surface. The edifice is sharply truncated by faults to the east of the summit (Figure 1.5). The half-dome has very low slope angles (less than 2°). It is the oldest exposed eruptive event associated with the Hengill Volcanic System and is partly buried by younger eruptive units in the fault grabens to the east (Árnason *et al.*, 1987) (Figure 3.11). It is entirely subaerial in origin and composed of compound pahoehoe lavas of variable thickness. The section of the lava pile studied here is composed of at least 8 lava flows from the deeper regions of the presently exposed shield (Appendix 1A).

The **Húsmúli** lava shield is situated on the western perimeter of the Hengill Volcanic System (Figure 3.11) and forms a dome structure, half of which is buried by younger hyaloclastite eruptions (Figure 1.5) (Sæmundsson, 1967). It covers an area of 7 km<sup>2</sup>, but is estimated to have covered at least 9 km<sup>2</sup> prior to burial, and the volume of exposed material is 0.64 km<sup>3</sup>. It is similar in appearance to the Hæðir lava shield, except the dome is more pronounced with the flanks sloping at 9°. Hyaloclastite occurs on the northern flank (Sæmundsson, 1967), but most of the dome is composed of subaerial compound pahoehoe lavas. The shield is reported to have been erupted directly after Hæðir (Figure 3.12) (Árnason *et al.*, 1987). The section studied here is composed of five flow-units probably from the mid-region of the shield (Appendix 1A).

The lava shields examined here are composed of monotonous sequences of eruptive products of variable volume that are believed to have been produced during one volcanic eruption event, that may have lasted for less than 50 years (after Sæmundsson, 1992). The larger volume shield (Ingolfsfjall) appears to have more complex geology than Hæðir and Húsmúli, but each lava shield is unique in some way. The small-volume lava shields are similar, but with variable slopes of their domes. The larger shield is composed of sub-units which may be the result of variations in the environment of eruption, and/or may represent pulses within the eruption. The latter is exemplified by the transition between the compound lavas and the coarse lavas of Ingolfsfjall.

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# **3.6) Radiometric dates and regional stratigraphy**

The opportunity to date twelve samples was offered by Professor R. A. Duncan (Oregon State University). The most analytically suitable samples were chosen from the author's collection from strategic positions within the stratigraphy. The objectives of dating these specific sample were: 1) to date the commencement of activity within the Grensdalur Central volcano; 2) to date at least some of the marker horizons, or eruptive units directly above or below them; 3) to solve specific problems that exist within the stratigraphy. The method used was K/Ar dating, using  $K_2O$  wt. % oxide values obtained during this research by XRF (Appendix 3).

Sample no	Stratigraphic location	Reason for dating	Result (ka)
A406	GA10, Hv	A sample within the Matuyama epoch (>700 ka).	592 +/- 117
A524	BA3, Hv	This unit is intercalated within the stratigraphy of Ingolfsfjall and the GCV, B.	642 +/- 133
A398	BA3, Hv	Part of the same eruption event as A524 (unit BA3) which is intercalated with the GCV, B.	Excess <sup>40</sup> Ar
A36	BB7, Hv	K-rich lavas directly below the IM2, B.	363 +/- 17
A65	Ingolfsfjall lava shield, Hv	To date the lava shield eruption event and to correlate it with the GCV, B.	446 +/- 64
A247	TA3, Hv	The freshest material that forms the core and base of the GCV, B.	Excess <sup>40</sup> Ar
A133	DB7a, Hv	Example of western half of the GCV, B.	Excess <sup>40</sup> Ar
A474	DB7b, Hv	Buried lava shield beneath Skálafell, B.	73 +/- 56
A301	S4, Hv	Example of the eastern half of the GCV, B.	Excess <sup>40</sup> Ar
A384	Hæðir lava shield, He	Date the shift of activity to He, B; (115-120 ka; *1).	80 +/-109
A383	Nesjaskógur, He	Lowest fissure eruption in He, B; (100 ka; *1).	59 +/- 38
A374	Hengill eruption, He	Date the top of the stratigraphic section sampled here, B; (50 ka; *1).	Excess <sup>40</sup> Ar

**Table 3.1:** The samples selected for K/Ar age determination, with the reasons for selection, the stratigraphic location and the age determination in thousands of years (ka). The samples are listed approximately in order of age, as determined from field relationships. The error bars here, and in Figure 3.14, represent 1 $\sigma$  (R. A. Duncan pers. comm., 1991). Samples that had huge amounts of atmospheric <sup>40</sup>Ar are indicated accordingly. GCV=Grensdalur Central Volcano; B=Brunhes epoch (<700 Ka); Hv=Hveragerõi Volcanic System; He= Hengill Volcanic System; (\*1)= dates according to Árnason et al. (1987).

The sample numbers, their stratigraphic location and the reasons for selection are summarised in Table 3.1, and the results are shown in Figure 3.14. Overall, the results satisfy few of the objectives. Nearly half of the samples had excessive atmospheric <sup>40</sup>Ar, which is probably the result of minor calcite disseminated in the groundmass of these basalts. The large error bars on all the determinations (except sample A36) are due to the low  $K_2O$  content of the samples (Figure 3.14). The only information provided by these age determinations that could not be deduced independently from the fieldwork is the age of the BB7 lavas at around 400 ka. The Ingolfsfjall lava shield was erupted after this time, and at least 1 glacial period later, indicated by IM2 tillite. This indicates that they are separated by at least 100 ka (after Harland *et al.*, 1990). The absolute time that activity began in the Grensdalur Central Volcano is not determined by these results.



Key

(\*) Arnason et al. (1987)

(+) Pullinger (1991)

- (Ø) Sæmundsson & Einarsson (1981)
- (Δ) Sæmundsson pers. comm. (1991)

(=) Kristjansson et al. (1988)

Figure 3.14: The results from the K/Ar age determinations (by R. A. Duncan). Error bars represent 10 (R. A. Duncan pers. comm., 1991). Also indicated for each sample is the name of the eruptive unit; the magnetic polarity (Sæmundsson & Einarsson, 1980; Kristjánsson et al., 1988); the estimated ages from field observations, using tillite horizons and stratigraphy, as determined here and from other field studies (references indicated in the Key). The actual age determinations from the K/Ar dating are given in brackets. The age acquired for BB7 indicates that the Ingolfsfjall lava shield was erupted after 400 ka. Overall, the age determinations agree well with the constraints provided by field mapping. However, there are some minor discrepancies, most of which can be explained by the large error bars for these samples, due to such low K abundances. For example, the relationship between A406, A524 and the Brunhes-Matuyama boundary, and the relationship between A65, A36 and the IM2 tillite; in both cases the stratigraphy has been overturned completely by the K/Ar dates. The only outstanding problem, where the difference in age estimation is well in excess of analytical error, is for a lava from the DB7b lava shield (sample A474). This problem cannot be resolved without further field investigation.

# **3.7) Migration of the locus of magmatic activity**

From Figures 3.1 and 3.2, the magmatic activity can be traced both spatially and temporally. Generally, it appears that the volcanism has oscillated within the region with a tendency to migrate westwards with time. Within the stratigraphic column there are natural breaks where the volcanic activity has apparently shifted elsewhere. For example, from the region occupied by TA sub-group the activity moved west to the region occupied by DB sub-group, and then east again to produce the S units. Similar oscillations seem to have occurred in the east between the older units of GA & BB. It is possible that such "shifts" are a result of an incomplete stratigraphic section due to erosion and burial. The stratigraphy observed here may represents only a small fraction of the 2 km thick crustal pile, and not the full volcanic history of the area over the last 1 Ma. Nevertheless, the evidence suggests that there may have been volcanic activity in both the Grensdalur and Ingolfsfjall areas at the same time and when this occurred, the volcanism seems to have been dominated by lava shield-type eruption events (e.g. the period around 400 ka; Figure 3.2).

# 3.8) Summary

Volcanology is used, in conjunction with other lines of evidence, to produce a geology map that illustrates the distribution of the various eruptive units. This map contains 81 eruptive units which can be divided into five main stratigraphic groups defined by six marker horizons. The stratigraphy of the Hveragerði Volcanic System spans approximately 600 ka, at which time (200 ka) the volcanic activity began in the Hengill and Hrómundartindur volcanic systems. Each of the morphologically distinct lava shields in the region has some unique aspect of its geology. Six buried lava shields are recognised and are related temporally with the Hveragerði Volcanic

System. Absolute ages were acquired for selected samples using K/Ar methods; these date the BB7 eruptive event at around 400 ka. This indicates that the Ingolfsfjall lava shield was produced after activity had commenced in the Grensdalur Central Volcano and suggests that the locus of volcanic activity has oscillated over this region for at least the last 900 ka.

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# **Chapter 4**

# **Petrography and Mineral Chemistry**

# 4.1) Introduction

In this chapter, the petrography and mineral chemistry of basalts collected from the Hengill to Ingolfsfjall region are described. The aims of this petrological study were: 1) to explore the *petrographic diversity*; 2) to examine the petrographic diversity in terms of *regional distribution*; and 3) to establish the extent of *equilibrium* between phenocryst and host rock. Petrographic observations were made from 330 thin sections of samples which are diverse in both texture and mineral assemblage. The mineral chemistry of a carefully selected subset of 16 samples was investigated using a Cambridge Instruments Geoscan microprobe fitted with a Link 860 energy-dispersive detector system (Appendix 3). The results of these investigations are presented below, and the mineral compositions are tabulated in Appendix 2.

Mineral phases that crystalise from a<sup>7</sup>primary magma<sup>8</sup> do so either prior to, or during and after eruption. These environments of crystallisation are referred to as "intratelluric" (prior to eruption in a reservoir) and "quench" (during and after eruption) (Cox *et al.*, 1979). In general, intratelluric crystals may grow to be large phenocrysts, whereas quenching produces the smaller crystals of the groundmass. This relationship is complicated when large crystals form on eruption, such as skeletal crystals, and when the intratelluric crystals are small (e.g. some opaque minerals). Additionally, a continuum of crystal sizes from large to small may exist (seriate texture), and then it is difficult to determine with confidence the environment of crystallisation of the individual crystals. With these complexities in mind, the term "phenocryst" is used here for crystals larger than the grainsize of the groundmass, and if the site of crystallisation has been confidently established, it is described accordingly as intratelluric or quench. In the majority of samples examined here, this distinction is possible.

# 4.2) Petrological observations

#### 4.2.1) Intratelluric environment

#### 4.2.1.1) Intratelluric single-crystal textures

#### Olivine phenocrysts of intratelluric origin

These are found as macro and microphenocrysts in a large proportion of the porphyritic rocks encountered, and range in composition from Fo 86-60 (Figure 4.1). They are especially abundant in the less-evolved basalts but also occur in more fractionated eruptive units. In the less-evolved basalts, the olivines are usually euhedral to subhedral with inclusions of euhedral chrome-spinel, whereas in the evolved rocks, anhedral crystals dominate and opaque inclusions are virtually absent.



Figure 4.1: Compositional variation in the crystals formed in the intratelluric and quench environments: a) for olivine, expressed as Fo %; b) for feldspar, expressed as An %.

#### Plagioclase feldspar phenocrysts of intratelluric origin

These are highly variable in both texture and composition (An 54-89) (Figure 4.1). The composition may vary by up to 25 absolute molar An % within a single sample (e.g. sample A495, Appendix 2A, p. 270-271). They range in form from euhedral to anhedral, show variation in the type of optical and chemical zoning and have various combinations of fluid/glass inclusions. From this abundant variation, several textural types of feldspar phenocryst can be distinguished. The shape of the crystals observed in thin section is dependant on the orientation of the thin section relative to the crystallographic axes of those crystals. Therefore the variation in

crystal textures in terms of shape is often displayed as a continuum, and with some overlap between the textural groups defined below.

The first and most common variety of feldspar is referred to here as megacryst (Figure 4.2). This crystal type is characterised by its shape, composition and large size range. The crystals are usually large (1-10 mm) and subequant with a low aspect ratio. They have calcic crystal-core compositions (An >80) which are unzoned, or poorly zoned, and often in the shape of an anhedral crystal. The core is surrounded by a mantle and then a thin rim of feldspar. These are both optically discontinuous and usually of different composition to each other, as well as to the core. The An content of the mantle is often similar to that of the smaller euhedral plagioclase phenocrysts in the sample, whereas the An content of the rim is usually similar to that of the quench feldspar. The amount of mantle overgrowth and core resorbtion are variable, resulting in variably-shaped crystals. The history of an individual crystal is preserved in its optical and chemical zoning, which often indicates an alternation between growth and resorbtion. However, for the megacrysts, the core compositions are always highly calcic. Although the mantles and rims are usually more sodic that the cores, in some cases the reverse is true (e.g. sample A416, Appendix 2A, p. 269).

The cores and mantles of the megacrysts usually have abundant glass inclusions. Two types of inclusions are observed: 1) small dot-like inclusions, generally restricted to the crystal cores; and 2) larger, elongate, devitrified glass inclusions, aligned parallel to the cleavage of the crystal and found in both the cores and the mantles (Figure 4.2). It is generally the case that the crystals (or parts) with dot-like inclusions are the most anorthitic (An >80) whereas the compositions of the crystals with the larger elongate inclusions are more varied (An 68-89).

Megacrystic feldspar of this kind is found at numerous spreading centres but is particularly abundant in the North Atlantic (Aumento, 1968; Melson & Thompson, 1971; Bougault & Hekinian, 1974; Blanchard *et al.*, 1976; Donaldson & Brown, 1977; Natland *et al.*, 1983). Similar plagioclase crystals have also been reported from Greenland (Larsen *et al.*, 1989). They were first described as xenocrysts (e.g. Aumento, 1968; Melson & Thompson, 1971) but later, the non-genetic term "megacryst" was adopted (e.g. Bougault & Hekinian, 1974; Natland *et al.*, 1983). This terminology is followed here, but it should be noted that the term "megacryst" does not necessarily imply a crystal of a phenomenal size. Table 4.1 summarises the characteristic features of feldspar megacrysts described here and from the Atlantic

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Figure 4.2: Photomicrographs of a typical feldspar megacryst (from within sample A368, Bæjarháls unit, Hrómundartindur Volcanic System): a) PPL; b) crossed polars (XP). The core (C) has some dotlike inclusions and minimal optical zoning, and the mantle (M) has elongate devitrified glass inclusions, aligned parallel to the twin plane. A thin white rim of feldspar surrounds the mantle, but is only visible when the mantle is in complete extinction. This is not the case in (b). The field of view is 4.5 mm. Spot analyses from this crystal are presented in Appendix 2A, p. 275 (crystal #=6). The locations of the spots are indicated with open circles.

Crystal Variety	Mega	acrysts	Pheno	ocrysts
Reference	Donaldson & Brown (1977)	The Hengill- Ingolfsfjall area	Donaldson & Brown (1977)	The Hengill- Ingolfsfjall area
Size (mm)	1-3	1-10	<0.8	<5 long
Shape	Round; subequant	Round; subequant or tabular	Euhedral; bladed	Euhedral; tabular or bladed
Composition	An 85-92	An 80-90	An 73-82	An 54-80

(Donaldson & Brown, 1977), and highlights the differences between these crystals and other feldspar phenocrysts.

**Table 4.1:** Comparison of the features of plagioclase feldspar megacrysts and plagioclase phenocrysts from the Mid-Atlantic Ridge as summarised in Donaldson & Brown (1977) and those observed from the Hengill-Ingolfsfjall area.

A second variety of feldspar crystals are the **tabular** intratelluric microphenocrysts (Figure 4.3). These crystals have a moderate to low aspect ratio. They often display oscillatory or step zoning, with sharp optical discontinuities, or they may display gentle sweeping optical zoning throughout the whole crystal. There is considerable compositional variation within and between crystals of a single sample (up to 13 absolute molar % of An), and some parts of the crystal may be as calcic as the megacrysts (up to An 86). Tabular crystals also usually possess a thin rim of feldspar that is in optical discontinuity with the interior of the crystal. The tabular variety of feldspar may be similar in appearance to the smaller megacrysts. However tabular crystals can be distinguished when they display oscillatory zoning, as this is not a characteristic of the megacrysts.

The third variety of feldspar is much less abundant and forms a more poorly defined group of **euhedral** crystals (Figure 4.3). They are texturally and compositionally varied but are characterised by a high aspect ratio on a prismatic crystal section. Basal sections would mimic tabular crystals in shape. Collectively, the euhedral variety of feldspars show features such as elongate glass inclusions, strong oscillatory zoning, and the ends of crystals are often incompletely formed (Figure 4.3). They lack a mantle of overgrowth, as seen in the megacrystic plagioclase, but may display a thin rim of feldspar of a more sodic composition (Figure 4.3).



Figure 4.3: Photomicrograph of euhedral and tabular intratelluric feldspar phenocrysts (labelled accordingly) from sample A495 (Unit S12). Note the high aspect ratio of the euhedral crystal. The field of view is 4.5 mm, and is the same as in Figure 4.2.

### Pyroxene phenocrysts of intratelluric origin

Clinopyroxene phenocrysts are rare in the samples studied, and are compositionally classified as augite  $(En_{49}Wo_{40}Fs_{11})$ to En<sub>44</sub>Wo<sub>44</sub>Fs<sub>11</sub> and  $En_{40}Wo_{30}Fs_{20}$  (Figure 4.4). The clinopyroxene phenocrysts analysed here from the Hengill to Ingolfsfjall region show a wide range of augitic compositions, both within individual crystals and within single eruptive units. Such chemical zoning is common in pyroxene (Thompson et al., 1980). Clinopyroxene phenocrysts typically occur in the more fractionated rocks and are absent from the less evolved basalts. The crystals show a variety of textures, broadly subdividing them into euhedral and anhedral phenocryst types. The anhedral crystals are usually large resorbed macrophenocrysts that may have strain extinction and be up to 10 mm in size. By contrast, the euhedral crystals are generally much smaller (<5 mm), and are thin and bladed in a prismatic crystal section.



Figure 4.4: Clinopyroxene compositions projected onto the Di-Hd-En-Fs quadrilateral (intratelluric crystals =  $\bullet$  and quench crystals =  $\bullet$ ).

Large clinopyroxene macrophenocrysts, of the anhedral type, are documented from the Hrómundartindur region (the Mælifell picrite, Figures 3.10 & 3.11; Hardardóttir, 1983; 1986; Risku-Norja, 1985; Trønnes, 1990; Hansteen, 1991). Some of these crystals are endiopside and have higher  $Cr_2O_3$  and  $Al_2O_3$  wt. % contents than is usual for mid-ocean ridge basalts (MORB) (e.g. Hansteen, 1991). In the field and in handspecimen, the clinopyroxenes that occur in the Hrómundartindur region are petrographically indistinguishable from the anhedral type of clinopyroxene observed in many of the eruptive units within the Hveragerði Volcanic System.

## 4.2.1.2) Intratelluric multi-crystal textures

Crystals produced in the intratelluric environment commonly interact with each other in glomerophyric clusters and in gabbroic xenoliths. Amongst the glomerophyric clusters present in these basalts there are generally two types observed, and feldspar is the dominant constituent in both (Table 4.2). Xenoliths occur in only a small proportion of the eruptive units where they may be locally abundant.

# Agglutinated glomerophyric clusters

These comprise macro and microphenocrysts of feldspar (with or without olivine and/or clinopyroxene phenocrysts) that are interlocked and strung together, often in a linear fashion (Figure 4.5a). The megacrystic and tabular varieties of feldspar form these clusters which are dominantly monomineralic. Olivine phenocrysts found in association with megacrysts in agglutinating glomerophyric clusters lack chrome-spinel inclusions. In handspecimen, a single megacryst may appear as large as a centimetre in length and very rounded. When examined in thin section, however, this "crystal" is composed of several smaller crystals intergrown, or of larger megacrysts partly or completely enclosing smaller ones. When olivine and clinopyroxene are observed in these agglutinating clusters they are anhedral and subophitically enclose feldspar. Occasionally megacrysts surround interstitial clinopyroxene, producing mesocumulate texture within the cluster.

Multi-crystal textures	Single-crystal textures	Petrological type (see Table 4.3)
Agglutinated clusters (Figure 4.5a)	Megacrysts and tabular	2
Radiating clusters (Figure 4.5b)	Tabular and euhedral	3

**Table 4.2:** Summary of intratelluric multi-crystal and single-crystal plagioclase feldspar textures, and how these relate to the petrological types described in the text below (Section 4.3).

# Radiating glomerophyric clusters

These clusters are composed of microphenocrysts of tabular and euhedral feldspar phenocrysts that radiate outward from a central point (Figure 4.5b). These may or may not have olivine macro and microphenocrysts concentrated in the core of the clusters. The radiating habit that is the characteristic feature of this cluster type is adopted by variable amounts within single samples and between different samples.

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Figure 4.5: Photomicrograph of the two varieties of glomerophyric clusters: a) agglutinating clusters with macro and microphenocrysts of megacrystic and tabular feldspar, partially intergrown with each other from sample G91 (unit DB9) (field of view is 18 mm); b) radiating clusters with microphenocrysts only of tabular and euhedral feldspar radiating outwards from a central point from sample A374 (the Hengill eruption event) (field of view is 4.5 mm). Note the difference in scale for these two types of glomerophyric clusters.

#### xenoliths

The xenoliths from the Hengill to Ingolfsfjall region are fragments of coarsegrained igneous material of a gabbroic nature (Figures 4.6 & 4.7). Such xenoliths have been described from the Mælifell eruptive unit from the Hrómundartindur Volcanic System (Figures 3.10 & 3.11) (Hardardóttir, 1983; 1986; Trønnes, 1990; Hansteen, 1991). A sample of basalt from the Mælifell eruptive unit, containing abundant xenoliths, was examined petrographically in this work (Figure 4.6). The abundant small xenoliths (<3 cm here) display orthocumulate texture with megacrystic feldspar poikilitically enclosed by either clinopyroxene or olivine. The composition of the feldspar is highly calcic (cores An 81-99; Hansteen, 1991). Towards the edge of the xenoliths, the crystals shows resorption along grain boundaries (Figure 4.6), which indicates that these xenoliths are in a state of partial disintegration (Trønnes, 1990; Hansteen, 1991).

The xenolith sample examined from the Hveragerői Volcanic System (unit S12) is a mesocumulate (Figure 4.7). It has interlocking, subhedral, megacrystic feldspar crystals with abundant glass inclusions in the cores of the crystal. Compositions of An 90 have been obtained from the core and rims of these feldspar crystals (Appendix 2A, p. 271). Olivine is a minor constituent that partly encloses the feldspar, and larger clinopyroxene crystals poikilitically enclose both the olivine and the feldspar. There is a large amount of intercumulus glass with vesicles, and feldspar crystals in contact with this glass have a thin, optically discontinuous rim. Xenoliths found in unit S12 are up to 7 cm in diameter in handspecimen, and are well rounded and highly spherical.

The xenolith samples examined from the Hrómundartindur and Hveragerði volcanic systems are thought to be magma chamber flotation cumulates, with megacrystic feldspar being the first phase to have crystallised, followed by olivine and clinopyroxene, the latter two presumably crystallising from the intercumulus glass. Only megacrystic feldspar occurs in the xenoliths examined here. This strongly suggests that the feldspar megacrysts represent crystal mush from a magma reservoir (i.e. cognate xenocrysts). Such crystals may not be in equilibrium with the rock in which they occur. Further discussion on the origin of plagioclase megacrysts may be found in Chapter 7.



Figure 4.6: Photomicrographs of xenoliths from basalt of the Mælifell eruptive unit (within the Hrómundartindur Volcanic System (Section 3.4)): a) pyroxene (Px) encloses macrophenocrysts of megacrystic plagioclase feldspar (F) that are in agglutinating glomerophyric clusters (field of view is 16 mm); b) olivine l) encloses smaller olivine crystals, as well as plagioclase (F) (field of view is 16 mm). The vesicular picritic groundmass is also indicated. Both olivine and pyroxene display strain extinction.



Figure 4.7: Photomicrograph of a xenolith from S12 eruptive unit (sample G22) (marker horizon within the Hveragerői Volcanic System (Section 3.3)), The xenolith is dominantly composed of macrophenocrysts of interlocking feldspar (F) (An 90; Appendix 2A, p.271). Olivine (Ol) has grown around these crystals and pyroxene (Px) encloses both olivine and feldspar. The continuity of the pyroxene is indicated by the optical continuity of this crystal here as seen in XP. The interstices contain vesiculated glass, with or without quench growth of crystals (olivine, pyroxene and feldspar are observed). The field of view is 18 mm.

# 4.2.2) Quench environment

# 4.2.2.1) Quench single-crystal textures

#### Olivine crystals of quench origin

Quench olivine ranges in composition from Fo 89-36 (Figure 4.1). This mineral phase is especially abundant in the less-evolved rocks but also occurs in some fractionated basalts. It may be present as rounded small grains, or as micro and macrophenocrysts showing a skeletal habit (Figure 4.8a).

# Feldspar crystals of quench origin

These usually occur as euhedral laths or needles and range in composition from An 85-53 (Figure 4.1). During quenching, feldspar grows on the edges of any existing intratelluric feldspar phenocrysts in the form of a thin rim (Figure 4.3). This may be intergrown with groundmass crystals and/or quench microphenocrysts and is indicative of simultaneous growth of the groundmass and the rim. This conclusion is supported by the fact that rim compositions are similar to the quench microphenocrysts within the same sample.

#### Pyroxene and opaque crystals of quench origin

The groundmass pyroxene is augite  $(En_{48}Wo_{40}Fs_{12} \text{ and } En_{44}Wo_{44}Fs_{12} \text{ to } En_{41}Wo_{37}Fs_{22})$ . There is little difference in composition between the quench pyroxene and the intratelluric clinopyroxene phenocrysts (Figure 4.4). The quench clinopyroxene occurs as small subhedral microphenocrysts showing strain extinction, or as tiny granular crystals in the groundmass. Comb-pyroxene also occurs. Opaque crystals occur either as euhedral rhombic grains or as wispy thin strands. Compositionally they are magnetites (Appendix 2D). Chrome-spinel compositions have been analysed for opaque inclusions within olivine phenocrysts (Appendix 2D).

#### 4.2.2.2) Quench multi-crystal textures

The crystals produced in the quench environment usually texturally interact with each other. Microphenocrysts represent the first minerals to crystalise as the magma enters a new pressure and temperature regime. Quench textures such as comb pyroxene and bow-tie intergrowths occur (Figure 4.8b). Such textures, and those for individual crystals described above, infer rapid crystal growth. The variable texture exhibited by the groundmass minerals is dominantly governed by the cooling rate during solidification. Slow cooling will result in textures such as seriate and ophitic (e.g. sample A512 from the Hæðir lava shield), whereas rapid cooling will produce glass devoid of microlites and crystallites. The groundmass of basalt collected from hyaloclastite formations usually displays quench textures, including interstitial glass, comb-pyroxene, spherulitic plagioclase or complete quench in the form of glass. Basalt groundmass of subaerial lava is usually equigranular and coarse grained. The grainsize of the groundmass, and the constituent minerals, may vary within a sample, and the transition from one region to the other may be well defined. This may be interpreted as evidence of incomplete mixing of magma in different cooling states or of distinct compositions.



Figure 4.8: Photomicrographs of quench textures: a) two skeletal crystals of olivine intergrown with feldspar needles in a fragment of altered hyaloclastite glass from sample T11 (unit GA2) (field of view is 1.75 mm); b) groundmass showing crystals of feldspar needles intergrown at 90° to each other in a "bow-tie" habit from sample A368 (Bæjarháls unit) (field of view is 1.75 mm).

# 4.3) Petrological types

In the present study, samples have been divided into three groups on the basis of crystal textures displayed by the feldspar phenocrysts (Figures 4.2, 4.3 and 4.5). These groups, or petrological types, are summarised in Table 4.3. Each group has been subdivided according to various mineral assemblages present amongst the samples. The mineral assemblages overlap, but each group has distinctly different textural types of feldspar within that assemblage (except petrological type 3b; see below). The petrological relationships within petrological types 1-3 are dominant in the study area, but minor exceptions and deviations from the pattern occur. The first group, referred to as **petrological type 1**, includes essentially aphyric rocks, which have been divided into two sub-groups by the presence or absence of quench olivine in the groundmass. This group of rocks appears aphyric in handspecimen but may contain a few small intratelluric microphenocrysts in thin section.

Petrological	Distinguishing features	Subdivisions
Types		
Type 1	Essentially aphyric basalt .	a) Aphyric + groundmass olivine.
		b) Aphyric - groundmass olivine.
Туре 2	Feldspar megacrysts with or without	a) Feldspar only
(Figure 4.9a)	clinopyroxene and olivine phenocrysts	b) Feldspar + clinopyroxene
	(minus spinel inclusions), some of which	c) Feldspar + clinopyroxene +
	may also be megacrystic in origin (Figure	olivine
	4.5a).	
Type 3	Radiating glomerophyric clusters of	a) Feldspar + Olivine
(Figure 4.9b)	intratelluric feldspar microphenocrysts	b) Olivine only
	usually with abundant intratelluric olivine	c) Feldspar only
	macro and microphenocrysts with spinel	
	inclusions (Figure 4.5b). Other variations	
	with only feldspar or only olivine exist.	

Table 4.3: Petrological types.

The second group includes basalts that have megacrystic feldspar and is referred to as **petrological type 2** basalt. An example of this type of basalt is shown in Figure 4.9a. The phenocryst content and abundances of this petrological type may be highly variable within a single eruptive unit. For example, in the Grensdalur Central Volcano, at Reykjafjall, the petrography of one particular eruptive unit (S9)



Figure 4.9: Examples of the two porphyritic petrological types: a) petrological type 2 basalt from Lómatjarnarháls unit from the Hrómundartindur Volcanic System (sample A169). This unit is a "mixed" unit, and this sample is an example of the highly porphyritic regions of the eruptive unit, and is from feeder basalt; b) petrological type 3 basalt from a sill intruded along the base of the Ingolfsfjall lava shield sample A40) (field of view is 18 mm for both).

varies within a single outcrop. A change is observed in a  $3\times5$  m outcrop of feeder basalt from aphyric at one end, through to highly phyric at the other, with no detectable contact between the two. Eruptive units like this, with a heterogeneous distribution of phenocrysts, are referred to as "mixed" in this thesis. Such units provide further support for the cognate nature of these plagioclase megacrysts. Euhedral and tabular feldspar phenocrysts may also be present in petrological type 2 basalt but usually in subordinate amounts. When clinopyroxene and olivine occur as significant constituents they are usually of the anhedral variety, and olivine lacks chrome-spinel inclusions. Clinopyroxene or olivine of this sort are never observed as a sole phenocryst phases in any of the eruptive units in the study area, and they are always associated with megacrystic feldspar. This relationship suggests that these crystals also represent cognate xenocrysts.

The third group, petrological type 3, includes basalts that have radiating clusters of feldspar microphenocrysts, with or without olivine phenocrysts, and no pyroxene phenocrysts, as exemplified in Figure 4.9b. The olivine phenocrysts in this basalt type dominantly have chrome-spinel inclusions. The majority of basalts that form this petrological type contain both olivine and feldspar phenocrysts, but there are exceptions. Many examples of basalt occur with only abundant olivine phenocrysts (with chrome-spinel inclusions). No basalt occurs with olivine phenocrysts as the sole phenocryst phase that lack chrome-spinel inclusions. These have been included with petrological type 3 basalt because of constraints provided by field observation. In a pile of cogenetic compound pahoehoe lavas (unit DB7b; Chapter 3, p. 59-60), a transition is observed from basalt with olivine phenocrysts only, at the base of the section, to basalt with olivine and radiating feldspar microphenocrysts towards the top. The transition occurs somewhere in the middle of the pile between sample A474 and A475 (see Appendix 1A, p. 238). Another exception exists in the form of a single occurrence of basalt dominated by radiating feldspar microphenocrysts alone. In this basalt type there are very rare macrophenocrysts of olivine, clinopyroxene and feldspar. The large feldspar crystals mimic the habit of megacrysts except they have well developed oscillatory zoning at the margins.

The petrological types defined here are analogous to descriptions of other basalt samples from MORs and elsewhere. Petrological types 1 and 2 are petrographically equivalent to aphyric and porphyritic tholeiites, respectively, and petrological type 3 basalts are equivalent to olivine tholeiites described in Iceland and

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from other MORs (e.g. Langmuir et al., 1977; Jakobsson et al., 1978; Sæmundsson, 1979; Batiza, 1982).

# 4.4) Regional and Stratigraphic distribution of the petrological types

## 4.4.1) Introduction

Samples that have been investigated petrographically, and those selected for mineral analyses, are representative of the whole spectrum of petrographic types discussed above (Section 4.3). They are also strategically positioned within the stratigraphy of the region (Chapter 3), and span the various eruption types encountered, i.e. the lava shield, fissure and central volcano (Chapter 1). In this section, the distribution of the petrological types are examined in terms of stratigraphy and spatial distribution within the study region. Figures 4.10 and 4.12 illustrate the petrographic variation within the stratigraphy of the Hveragerði Volcanic System, and the Hengill and Hrómundartindur Volcanic Systems, respectively.

### 4.4.2) Hveragerði Volcanic System

#### Ingolfsfjall region

The eruptive units of the Ingolfsfjall region are dominantly composed of basalt of petrological types 1 and 3, with only one occurrence of petrological type 2 basalt (Figure 4.10). Petrological type 3 basalt occurs in the lava shields and unit GB4. The Ingolfsfjall lava shield has a much coarser grained lava at the top of the succession, and the solidified lava lake is composed of gabbro with feldspar, clinopyroxene and an opaque phase (Figure 2.16). Some of the petrological type 3 units in this region lack feldspar phenocrysts, but the olivines have abundant chromespinel inclusions (e.g. unit BB4). Petrological type 2 basalt forms unit GB6, but the "megacrysts" in this unit have well developed oscillatory zoning at the margins. Nevertheless, the abundant olivine and pyroxene phenocrysts classify the basalt of this unit as petrological type 2 basalt.

# Gljúrfúr region

This region is dominated by petrological type 1 basalt which has very sparse microphenocrysts of olivine and feldspar present in agglutinating microglomerophyric clusters. There are only minor occurrences of petrological type 3 basalt (e.g. unit BA3; a lava shields eruptive event). No petrological type 2 basalt was observed within this region.



Figure 4.10: Distribution of petrological types (1-3) within the stratigraphy of the Hveragerői region (See Figures 3.1 & 3.2, p. 42-43). The recognised lava shield-type eruptive events are encircled. Dashed lines indicates an origin from shields, but where field evidence is not conclusive (see Chapter 3 for discussion on each particular example).

#### Grensdalur region

Within the Grensdalur Central Volcano, Petrological types 1 & 2 basalt dominate the stratigraphy. The older eruptive units are of petrological type 1 basalt (sub-group TA). The younger TB units consist of highly porphyritic petrological type 2 basalts (e.g. units TB1 & TB6) intercalated with aphyric units of petrological type 1. This pattern is repeated throughout the remainder of the stratigraphic column for the central volcano, where petrological type 2 basalt dominated. The DB sub-group has a low phenocryst content (<5% megacrysts), whereas crystal abundances increase within the porphyritic S units. Petrological type 3 basalt is poorly represented in the succession of the Grensdalur Central Volcano and is confined to the buried lava shields of DA1 and DB7b and a veneer of hyaloclastite breccia from unit BA3 that is intercalated within the stratigraphic section (Appendix 1B).

#### 4.4.3) Hengill and Hrómundartindur volcanic systems

The petrological variation within the Hengill and Hrómundartindur volcanic systems has been described by Árnason *et al.* (1987), who observe that the oldest eruptive units are generally aphyric and the percentage of phenocrysts increases with time within each volcanic system; a pattern similar to that described above for the Hveragerði Volcanic System. Porphyritic units, interdigitated with aphyric units, are dominantly of petrological type 2 basalt. A summary of the petrological type determined for the eruptive units sampled here is given in Figure 4.11. Mixed eruptive units, like those in the Hveragerði Volcanic system have also been observed (e.g. Lómatjarnarháls unit; Figure 4.9a). The lava shield Hæðir is composed of a coarse variety of petrological type 3 basalt, dominantly displaying seriate texture. Scarce clinopyroxene phenocrysts have been reported from this lava shield (Hardardóttir, 1983), but none was observed in the samples collected here (eight successive lavas in total), or in the field outcrops, indicating that they are probably a minor constituent.

Húsmúli is noticeably different from the other lava shields in that it lacks abundant olivine phenocrysts and contains occasional clinopyroxene and feldspar macrophenocryst. These phases, plus olivine, are present in very small amounts (<1% modal abundance). The radiating clusters of feldspar microphenocrysts are indistinguishable in texture and composition from those in other petrological type 3 and lava shield samples. The only other occurrence of petrological type 3 basalt is in the Hengill eruptive unit. In this unit , there is localised accumulation of olivine phenocrysts observed in the field. The similarity in petrographic character of basalt from the Hengill unit to basalt from lava shields (e.g. Ingolfsfjall) provides further supporting evidence which suggests that the Hengill mountain represents a lava shield type of eruptive event.



Figure 4.11: Petrological variation in a) the Hengill Volcanic System; b) the Hrómundartindur Volcanic System. The petrological information is from this research, whereas the stratigraphic order is taken from Árnason et al. (1987). The lava shield eruptive events are underlined. Hengill, proposed to be a lava shield in this research is underlined by dashes.

# 4.4.4) Summary of regional and stratigraphic distribution

Each volcanic system is composed of eruptive units that are highly variable in terms of the petrological type of basalt produced there. However, periods of the stratigraphy exist where successive eruptive units are dominated by aphyric basalt (e.g. TA & GA sub-groups). In all three volcanic systems, as the volcanism persists, the petrological type 2 basalt, with abundant megacrysts, becomes progressively more abundant. This is well represented in the Grensdalur Central Volcano. The lava shields recognised within the area are all composed of petrological type 3 basalt, although there are subtle differences within the mineral assemblages between each eruptive unit of this type. The fissure eruptions generally erupt petrological type 1 and 2 basalt. In general, the two types of feldspar texture (e.g. Figure 4.5) are mutually exclusive within eruptive units, and the fact that they are generally restricted to distinct eruption types suggests that these two types of volcanic eruption event may each have mutually exclusive magmatic plumbing systems.

# **4.5)** Crystallisation processes

### 4.5.1) Principles of equilibrium

Crystallisation mechanisms are broadly divided into two extreme end member concepts of equilibrium and fractional crystallisation (Cox *et al.*, 1979). During *equilibrium crystallisation*, compositional equilibrium between the liquid and the solid is continuously maintained. This means that, as a crystal grows under equilibrium conditions, material of continually changing composition is added to its rim but the crystal remains homogeneous by solid-state diffusion within itself. The resulting crystals are unzoned and the bulk composition of the end product is the same as that of the starting magma composition. *Fractional crystallisation* occurs where equilibrium between the crystals and the liquid is not maintained. This is achieved when crystals become separated from the melt or when solid-state diffusion is insufficiently fast to maintain crystal homogeneity.

In a closed system undergoing fractional crystallisation, the earlier formed parts of the crystal are prevented from re-equilibrating with the evolving melt by subsequent crystal growth. The cores of the crystals will then be out of equilibrium, while the rims will be in equilibrium with the surrounding liquid, resulting in what is commonly referred to as "normal zoning". Similarly, the first formed crystals will have a different composition to the last formed crystals. The bulk composition of the remaining liquid plus phenocrysts in this closed system, will still equal that of the initial primary magma.

In an open system, fractional crystallisation is inevitable. In addition to crystals with normal zoning derived from the host magma, crystals and liquid may be physically added to, or removed from, the system. Crystals removed from one melt and incorporated into another are described as cognate xenocrysts, and any crystals incorporated from the wall rock are described as accidental xenocrysts. The final bulk composition of the liquid plus phenocrysts may be drastically different from the composition of the initial magma. It is likely, therefore, that many magmas involved in open-system magmatic processes carry their own zoned phenocrysts, as well as cognate and accidental xenocrysts. It is important to establish whether phenocrysts are of xenocrystic origin as such crystals *may* cause the whole-rock geochemical trends to deviate from the true crystal-liquid fractionation paths.

# 4.5.2) Objectives

The porphyritic samples gathered during this research have intratelluric plagioclase phenocrysts that show different textures (Figures 4.2, 4.3 & 4.5), and the olivine phenocrysts found in association with these feldspars are also distinct (Section 4.3). The intratelluric plagioclase phenocrysts define two porphyritic petrological types of basalt that are essentially mutually exclusive in the field with respect to lava shields and fissure eruptions (Figure 4.9). These observations suggest that the two porphyritic petrological types are the result of variable intratelluric environments of crystallisation. In addition, variation in crystallisation mechanisms and/or crystallisation histories may have played important roles.

To investigate the magmatic processes that have influenced these basalts prior to eruption, the compositional relationship, in terms of the degree of equilibrium, is examined amongst the intratelluric phenocrysts, and between the phenocrysts and the bulk composition of the host rock. Certain predictions can be made for the various relationships expected for either open- or closed-system magmatism, and for the occurrence of foreign crystals. Phenocryst compositions would be more variable for open-system magmatism, and xenocrysts might be expected to be more strongly out of equilibrium than true phenocrysts. The methodology behind these investigations is outlined below.

# 1) Diagrammatic representation of equilibrium/disequilibrium

If equilibrium exists there should be no variation in the content of Fo or An between crystals in a single sample and there should be a correlation between Fo (and An) and the differentiation index, Mg # (<sup>1</sup>) (Roeder & Emslie, 1970). It is generally accepted that in natural magmatic systems pure equilibrium crystallisation does not occur. Therefore, the degree of disequilibrium may be estimated from the extent to which the above criteria for equilibrium are not met. For this reason, the Fo and An % of the olivine and feldspar crystals respectively are plotted against their respective whole-rock Mg # (Figure 4.12).

# 2) Computer modelling of the nature of the magmatic system

A computer program that models crystallisation processes is used (Nielsen, 1988). The objective is to establish a specific set of conditions for crystallisation and

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Mg # =

100	((MgO/40)
	[(MgO/40)+0.85(FeO*/72)]

(FeO<sup>\*</sup>=total iron in the sample, regardless of its oxidation state. The  $Fe_2O_3$ /FeO ratio = 0.15).

to see if the modelled products of this type of crystallisation match the observed products in the sample. If they do not, then the sample is not the product of crystallisation under those specified conditions.

#### 4.5.3) Diagrammatic representation of disequilibrium

Figure 4.12 shows the compositional variation in Fo and An % of the olivine and feldspar crystals respectively relative to the Mg # of the whole-rock compositions. Generally, multiple data points for a given Mg # value represent multiple crystals in one sample. All the samples show some variation in both Fo and An % of their crystals which confirms that pure equilibrium crystallisation has not occurred. The crystals are differentiated into intratelluric and quench varieties. In general, the Fo and An % of the quench crystals decreases with falling Mg #, whereas the intratelluric crystals tend to have higher values of these parameters at a given Mg #. This suggests that the quench crystals may be in equilibrium with their host liquid, whereas the intratelluric crystals are not. The compositional difference between intratelluric and quench crystals is greater for feldspar than it is for olivine.

In Figure 4.12a, the olivine compositions remain fairly constant at about Fo 80 until the Mg # decreases to about 40-45, below which there is a small number of considerably more fayalitic olivine crystals (Fo <60). In general, there is only a small compositional difference between the quench and intratelluric olivine crystals (up to 15.4 absolute molar Fo %), with the quench ones being dominantly more fayalitic than the intratelluric crystals within the same sample.

The olivine crystals in sample A512 (Hæðir lava shield) are positioned separately from the other data points with comparable whole-rock Mg # and have Fo compositions that are similar to those of olivine crystals in samples with much lower Mg # (Figure 4.12a). This basalt sample has large olivine phenocrysts with compositions of Fo 80, but the microphenocrysts and smaller crystals have a much lower Fo content. The seriate texture in the lavas from the Hæðir lava shield suggests that crystallisation was a steady and slow process that took place mostly *after* eruption in the closed system of the lava itself. The fayalitic composition of the smallest crystals can be explained by extensive fractional crystallisation in this closed system. The other samples from which olivine crystals were analysed in Figure 4.12a have either finer grained or very glassy groundmass. In these samples, the quenching has halted fractional crystallisation. This process is schematically illustrated in Figure 4.13.



Figure 4.12: Variation diagram of a) Fo in olivine, and b) An in feldspar crystals, both against Mg #. The mineral data are differentiated according to the environment of crystallisation. Crystals of uncertain crystallisation environment are omitted (<3% total crystals analysed) (except sample A512, see text).

As mentioned earlier, the feldspar compositions are more variable than the olivine compositions (Figure 4.12b). The An content of the quench feldspar crystals clearly falls steadily and sharply with decreasing Mg #, whereas the intratelluric feldspar crystals follow a similar trend with a more gentle slope and a much wider range of compositions for a given sample (up to 25.6 absolute molar An %). A line can be drawn through the field of quench feldspar crystals that may represent an equilibrium crystallisation trend. However, the observations made above for the olivine crystals of sample A512 may be taken as an indication that this may be an incorrect assumption. The last crystals to be formed will be in equilibrium with the Mg # of the last remaining liquid of the magma, and not the whole-rock Mg #. Where analyses on both whole-rock and glass are available from a single eruptive unit, there is usually a difference in the order of 5 absolute Mg # units, with the glass having the higher values (e.g. sample A28; Appendix 4, p. 310 and Appendix 5, p. 351). Nevertheless, there is a positive correlation between the whole-rock Mg # and An content of the quench feldspar, and any intratelluric feldspar compositions that fall on or near this trend are dominantly analyses acquired from the rims of the crystals.



Figure 4.13: Binary phase diagrams showing the effects of quenching a magma on the composition of olivine crystals under fractional crystallisation conditions: a) the expected composition of olivine under quench conditions; b) the expected compositions when cooling is slower and in a closed system. "start" = the Fo % in the starting liquid; and "End" = the Fo % in the final solid mineral.

In Figure 4.14a, the intratelluric feldspar crystals are differentiated into textural varieties (see Table 4.2), and are plotted together with the quench feldspar crystals. The feldspar megacrysts have the higher An %, and the euhedral crystals fall in, or just above the field defined by the quench crystals. The tabular crystals from radiating glomerophyric clusters are much more varied in composition and fill the

compositional gap between the megacrysts and the quench crystals with some overlap. Both the megacrysts and the tabular feldspar varieties are out of equilibrium with the samples in which they occur, but the degree of disequilibrium is greater for the megacrysts.

Figure 4.14b shows the compositions of the feldspar megacryst cores only, relative to the composition of the quench feldspar. The cores of the megacrysts are unzoned and show rounded and often resorbed margins separating them from their mantles. These cores are thought to represent the initial megacryst crystal before the mantle and rim were added. It is evident that the megacryst core compositions are virtually constant, with a slight negative trend, in the region of An 80-90%, regardless of the Mg # of the whole-rock or the composition of the quench feldspar of the basalt in which they occur. This suggests that they could have been the result of a single regular magmatic crystallisation process in an environment where the magma had a very uniform composition. This consistency in composition further suggests that the feldspar megacrysts are foreign to the rock in which they now occur.

There is one occurrence of feldspar megacrysts where the rims have similar compositions to the cores (S1 unit, Sample A416, Appendix 2A, p. 269). This suggests that the composition of the whole-rock is in equilibrium with the megacrysts, and this composition may represent the type of magma that crystallised these very commonly occurring crystals. However, this unit is compositionally unique within the Hveragerði area. The compositional relationship observed in sample A416, between the cores and rims of the feldspar megacrysts, has also been reported from the Krafla region, N Iceland (Nicholson, 1990).

In contrast to the megacrysts, the oscillatory zoned tabular feldspar crystals have much more varied An content (Figure 4.14a), which suggests that they have crystalised in an environment with very irregular composition. The oscillatory zoning indicates that the parental melts did not simply evolve towards precipitating less anorthitic compositions. The radiating texture of the tabular microphenocrysts of feldspar suggests that the crystallisation mechanism was consistent. Such a mechanism is not well defined or obvious, but the radiating clusters are similar to the variolitic habit adopted during quench crystallisation (Figure 4.8). It is proposed here that the radiating glomerophyric clusters of oscillatory zoned tabular feldspars may represent crystallisation en route to the surface in dykes extending to the surface from considerable depths.





Figure 4.14: Feldspar compositions (An %) plotted against the Mg # of the whole-rock samples in which they occur: a) Crystals differentiated into textural varieties of intratelluric feldspar phenocrysts (Section 4.2.1.2; Table 4.2) and quench feldspar; b) megacryst crystal core compositions plotted together with the quench crystals.

In summary, the diagrams presented above (Figures 4.13 & 4.15) may be used to infer the degree of disequilibrium between crystals and whole-rock in which they occur. The megacrysts and tabular varieties of intratelluric feldspar both represent disequilibrium phenocrysts. The main difference between them is the degree of variability in their composition. This difference is thought to reflect differences in their mechanism and environment of crystallisation.

## 4.5.4) Computer modelling of the nature of the magmatic system

Selected sample compositions were modelled using the computer program TRACE3.FOR (Nielsen, 1988). This program is designed to model low-pressure differentiation processes in natural mafic and intermediate magmatic systems. Given a magma composition, the program calculates the mineral compositions, proportions and equilibrium temperatures of crystallisation for each phase. After each calculation, a prescribed increment of the liquidus phase(s) is removed from the liquid and the program loops back to calculate the equilibrium temperature and composition for the residual liquid. The version of the program used here is written in FORTRAN and is run in DOS from a diskette.

The program provides an option for either closed- or open-system magmatism. The program options selected for this modelling are for closed-system fractional crystallisation. Whole-rock analyses have been used as starting compositions, and the predicted equilibrium phases and their compositions are compared with those observed in the eruptive unit concerned. If the sample is the product of closed-system fractional crystallisation, the program is expected to predict the observed assemblage. However, if xenocrysts of exotic compositions are present in small amounts, the modelling will be unable to predict them, and if they are very abundant the whole-rock composition may be so different from the original magma that none of the observed phases will be predicted. Samples from four eruptive units were modelled here that represent the two porphyritic petrological types, 2 and 3 (Table 4.3). These samples and the results of this modelling are summarised below and the details are provided in Table 4.4.

# 1) Petrological type 2 basalt: Sample A495

This is a highly porphyritic hyaloclastite basalt fragments from unit S12 with phenocrysts of feldspar (megacrysts, euhedral and tabular varieties), olivine and clinopyroxene. The feldspar megacrysts occur in agglutinating glomerophyric clusters with the olivines and clinopyroxenes. The groundmass is largely glassy, with local


occurrences of comb-pyroxene, and abundant quench microphenocrysts of feldspar and olivine. Disequilibrium of the phenocrysts is suggested in Figure 4.12. The phases predicted by the modelling are not observed in the sample as phenocrysts or as crystals in the groundmass. This suggests that the sample is not the product of closedsystem fractional crystallisation.

Sample	Eruptive unit	Pet type	I %	T°C	Calculated (Nielsen, 1988)	Observed
A495	S12	2	1	1197	An 66	Phen: Fo 86-75; An 85-59; augite
			9	1195	An 65	Ground: Fo 77; An 78
<b>A</b> 474	DB7b	3	<1	1286	spinel	Phen: Fo 86-83 + spinel Ground: Fo 83-77; An 78; augite
			1	1239	Fo 82 + spinel	
			4	1236	Fo 82	
A554	Ingolfsfjall	3	1	1225	Fo 81 + spinel	Phen: Fo 80-82; An 83-82
			8	1164	Fo 78 + An 63	Ground: An 71; olivine; augite
A550	GB4	3	<1	1275	spinel	Phen: Fo 85-89; An 81-85 Ground: felds., olivine, pyroxene
			1	1220	Fo 88 + spinel	
				1206	Fo 87, An 78 + sp.	

Table 4.4: Results of crystallisation modelling. I % = increment of fractional crystallisation expressed as a percentage; Pet type = petrological type (as defined in Table 4.3)

# 2) Petrological type 3 basalt: Sample A474

This is a lava from the DB7b eruptive unit (buried lava shield). The basalt has intratelluric olivine phenocrysts, the larger of which have chrome-spinel inclusions (Appendix 2D). The groundmass is fine-grained and granular with quench microphenocrysts of feldspar and olivine. The predicted phases for this sample closely match the observed assemblage, which suggests that the sample could have crystalised at low pressures in a closed system by fractional crystallisation processes, and that no significant addition of foreign phenocryst of drastically different composition has occurred.

# 3) Petrological type 3 basalt: Sample A554

This is a hyaloclastite basalt fragment from the Ingolfsfjall lava shield. Mineral analyses have been acquired from the glassy A9 sample of the same lava shield, which contains intratelluric micro and macrophenocrysts of olivine (plus chrome-spinel inclusions) and radiating feldspar microphenocrysts. The modelling predicts the compositions of the olivines, but fails to predict the high An content of the intratelluric feldspar compositions. There is a difference of 20 absolute molar An % between the predicted and the observed feldspar compositions.

## 4) Petrological type 3 basalt: Sample A550

This is a hyaloclastite basalt fragment from the GB4 eruptive unit. It too has mineral analyses acquired from a glassy sample from the same unit (sample A44). The whole-rock sample (A550) is less porphyritic than A554, from the Ingolfsfjall lava shield, but has the same constituent phenocryst phases in the same proportions. The modelling reproduces the observed olivine compositions, but not the feldspars, which are predicted to be much lower in An content than those observed; however, the difference in An % is only 7 absolute molar An %.

In summary, the results of the modelling suggest that petrological type 2 basalt (sample A495) has not experienced shallow-level, closed-system fractional crystallisation, and open-system magmatism is strongly suggested by the failure of the program to reproduce *any* of the observed phases. The DB7b lava shield may have been produced by closed-system fractional crystallisation, whereas Ingolfsfjall and unit GB4, of the same petrological types, could not have been. In all three examples of petrological type 3 basalt, the program matches well the olivine compositions, but not the feldspar, where it occurs. This modelling, coupled with petrographic observations and inferences made from Figures 4.13 & 4.15, suggests that the disequilibrium feldspar phenocrysts can be divided into two categories: the megacrysts and the radiating tabular feldspars, neither of which are the product of closed-system magmatism An explanation has to be sought to explain why one variety is consistent in composition (i.e. megacrysts), while the other produces variable compositions (i.e. tabular).

# 4.6) Summary

The rocks studied in this thesis are similar to other Icelandic basalts and MORB (e.g. Donaldson & Brown, 1977; Nicholson, 1990). Their crystals can be divided on textural grounds into intratelluric and quench varieties. Amongst the intratelluric feldspar phenocrysts there are two mutually exclusive varieties: the feldspar megacrysts that tend to form agglutinating glomerophyric clusters; and the tabular microphenocrysts that form radiating glomerophyric clusters. Feldspar crystals observed in cumulate xenoliths are of the megacrystic variety.

The samples are divided into petrological groups on this textural basis. Each group is further subdivided on the basis of observed mineral assemblages. By examining the regional and stratigraphic distribution of these basalt types, it is concluded that all the lava shields are composed exclusively of petrological type 3 basalt. The Hengill mountain is also composed of this type of basalt. In the Grensdalur volcano and the Hengill and Hrómundartindur volcanic systems the feldspar megacrysts become progressively more abundant with time. The mutually exclusive occurrence of these basalt types between lava shields and fissure type eruptive units suggests mutually exclusive environments of crystallisation, where the crystallisation mechanisms are different.

The degree of disequilibrium between the phenocrysts and their hosts is investigated, and the samples are modelled in terms of closed-system magmatism. It is observed that the megacrysts are of nearly constant composition, regardless of the degree of differentiation of the sample in which they occur. The radiating feldspars are also out of equilibrium, but by lesser and more varied amounts. The observed mineralogy of the two porphyritic suites can not be totally reproduced by modelling closed-system fractional crystallisation. For petrological type 2 basalt none of the observed phases are reproduced, whereas for petrological type 3 basalt only the radiating felspars can not be reproduced, which are consistently more anorthitic than the predicted compositions.

# Geochemistry

# 5.1) Introduction

In this chapter, the geochemistry of the rocks from the study area is described and used to elucidate their genesis. A brief explanation is given on the types of geochemical diagram used and the petrogenetic information that can be derived from them. The variation in the data is then examined in terms of the petrological types, and the regional and stratigraphic distribution of the samples, as described in Chapters 4 and 3 respectively. Finally, petrogenetic modelling is employed to constrain the origin of the rocks. Previous geochemical investigations in the study area have been restricted to various eruptive units in the Hengill and Hrómundartindur regions (Hardardóttir, 1983; 1986; Trønnes, 1990; Hansteen, 1991) and none has concentrated on the Hveragerði Volcanic System.

About 450 samples were analysed from the study area. Chemical analyses were performed on both whole-rock and volcanic glasses, and the techniques employed are detailed in Appendix 3. The whole-rock data are tabulated in Appendix 4, and the volcanic glass data are in Appendix 5. Major element oxides and some trace elements were routinely analysed by X-Ray Fluorescence (XRF) spectrometry, whereas other trace elements, including the rare earth elements (REE), were determined by Inductively Coupled Plasma-Atomic Emission Spectrometry (ICP-AES) and Inductively Coupled Plasma-Mass Spectrometry (ICP-MS). A summary of methods, elements and samples analysed is given in Figure 5.1. In cases where an element was analysed using more than one method, only the concentration obtained by the most accurate method is shown and tabulated in Appendix 4 (after Potts, 1987). Precision error bars  $(2\sigma)$  for the XRF analyses are assessed by the deviation of replicate analyses on International Reference Materials (IRM) and internal monitors (tabulated in Appendix 4, p.308).

Volcanic glass samples were analysed by electron microprobe for major element oxides (24 sample). These include those oxides listed in Figure 5.1 for the whole-rock fused powders, plus chromium. About 4-5 spots were analysed per sample and their compositions averaged. For both whole-rock and volcanic glass analyses, iron is expressed as FeO\*, which indicates total iron content detected in the sample, regardless of its oxidation state.



Figure 5.1: Flow chart illustrating the various analytical techniques used on the whole-rock samples; the number of samples analysed by each technique; and the element oxides and trace elements obtained from each.

Because of the presence of a high-temperature geothermal field in the study area, hydrothermally altered rock is locally abundant, particularly in the core of the Grensdalur Central Volcano. There have been numerous accounts of element mobility

\* SiO2, Al2O2 and MgO were also determined from powdered pellets, and are tabulated in Appendix 4; however, they are not included 106 in the arguments presented in this Chapter because these data are of too low quality. in basalts resulting from hydrothermal alteration (e.g. Mottl & Holland, 1978; Humphris *et al.*, 1980; Morrison & Thompson, 1983; Berndt *et al.*, 1988) and many of the samples demonstrate obvious alteration in thin section. For this reason it was necessary to evaluate the degree of element mobility in the present suite of samples. Elements found to be mobile include CaO, Na<sub>2</sub>O, K<sub>2</sub>O, Sr, Ba and Rb. Other elements that display random scatter, but to a lesser degree, include FeO\*, Co and V. These elements are present in secondary opaque minerals, such as pyrite, which is observed in some of the intensely altered areas. Altered samples are indicated accordingly in Appendix 4 and omitted from diagrams involving these elements (unless stated otherwise).

# 5.2) Geochemical systematics and data presentation

## 5.2.1) Systematics

There may be much compositional diversity within a suite of igneous rocks. This diversity, however, is normally not random and there is a geochemical "coherence" (Cox *et al.*, 1979). Often chemical variations are expressed as a continuum which is usually interpreted in terms of an evolutionary sequence. For example, unevolved, basic igneous rocks are often considered to change progressively to more-evolved, acid igneous rocks. This coherence stems from the differing chemical behaviour of the various elements that constitute the minerals which crystallise from magmas, and the consequent effects of the accumulation of such minerals within, or their separation from, an initial liquid.

In magmatic systems, differential partitioning of elements between crystals and melt is controlled by pressure, temperature and melt composition. Elements that preferentially partition into the solid phases are referred to as **compatible** elements and those which preferentially partition into the liquid are referred to as **incompatible** elements. The degree of compatibility of an element with a particular mineral phase is expressed by its distribution coefficient ( $K_D$ ); the concentration of the element in the solid divided by its concentration in the liquid (Arth, 1976).

In a MOR environment, the mineral phases involved in mantle melting processes are olivine, orthopyroxene, clinopyroxene, and spinel (i.e. a *dry* system). Garnet could also be involved when melting starts at greater depths (approximately 75-85 km; Nickel, 1986). The main mineral phases associated with basalt crystallisation processes are usually olivine, plagioclase, clinopyroxene and

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orthopyroxene but accessory minerals, such as chrome-spinel, magnetite, ilmenite, zircon and apatite, may also occur.

Certain major element oxides are compatible with specific mineral phases. MgO is compatible with olivine and pyroxene;  $Al_2O_3$  with feldspar and spinel; CaO with feldspar and pyroxene; FeO\* and TiO<sub>2</sub> with pyroxene, magnetite and ilmenite;  $Cr_2O_3$  with chromite and pyroxene; and  $P_2O_5$  with apatite. SiO<sub>2</sub> is present in most of the major rock-forming minerals but not in the accessory minerals that occur in a basaltic system.

The trace elements also preferentially partition into specific mineral phases: Cr is compatible with chrome-spinel; Co and Ni with olivine, orthopyroxene, clinopyroxene, garnet and spinel; Sc with orthopyroxene, clinopyroxene, magnetite and garnet; V with magnetite and pyroxene; Sr (and Eu) with plagioclase; the REE with apatite; Zr and the heavy REE (HREE) with zircon; Y and the HREE with garnet. The list of elements below are those that have been analysed during this research, and they are arranged in order of decreasing incompatibility in a dry basaltic system: K, Rb, Ba, Nb, La, Ce, Pr, Sr, Nd, P, Zr, Sm, Ti, Eu, Gd, Dy, Ho, Y, Er, Yb, Lu, Ga, V, Sc, Zn, Co, Ni, Cr (i.e. K is the most incompatible and Cr is the least incompatible element). However, this order varies slightly from author to author and is subject to new information on the values of  $K_D$  from experimental studies. All the trace elements, when referred to in this thesis, will be listed in their order of incompatibility, as above (from Kostopoulos, 1988).

## 5.2.2) Data presentation

The geochemical diagram is used to display correlations between different elements or combinations of elements. There are many different types of geochemical diagram available. Below is a brief description of the diagrams used here and how petrogenetic information may be extracted from them.

1) Variation diagrams: These are simple X-Y plots in which a single parameter (i.e. element or element combination) may be fixed on the X-axis whereas the Y-axis parameter is allowed to vary so that correlations between X and different Y elements may be observed. When examining a basaltic system, it is common to use on the X-axis a parameter compatible with mafic phases (e.g. Mg or Ni). This is because the mafic phases crystallise first relative to the salic ones, and thus provide a starting

point from which to examine an evolutionary sequence. This parameter on the X-axis is referred to as a differentiation index.

An ideal differentiation index would be a linear function of falling temperature in the magmatic system (Wager, 1956; Nielsen, 1988), and behave in a similar manner to the trend displayed in Figure 5.2. Variation diagrams of any element against a differentiation index for a cogenetic suite of rocks, therefore, depict the evolutionary path(s) of the parental melt(s) as fractional crystallisation proceeds. This path is often called the liquid line-of-descent (Cox *et al.*, 1979). It is shown best by analyses of samples that represent true liquid composition (i.e. aphyric whole-rock or volcanic glass samples).



Figure 5.2: A variation diagram of Mg # and temperature (Temps. °C) of the crystallising liquid. The primary melt composition is from Nui & Batiza (1991), and the liquid has been fractionated to 50 % using the crystallisation program of Nielsen (1988) (See Section 4.5.4). Increments of closed-system fractional crystallisation are indicated on the trajectory.

The differentiation indices used here are Ni ppm and magnesium number (Mg # - as defined in Section 4.5.3, p. 95). For the whole-rock samples, the differentiation index used is Ni because MgO was not determined for the majority of those samples (Figure 5.1). Reference to Figure 5.3, however shows that the choice of Ni as an alternative index is justifiable. For the volcanic glass samples the Mg # is used in the absence of Ni. The Mg # is preferred over MgO because it is unaffected by plagioclase crystallisation (Cox *et al.*, 1979).

2) Normalised multi-element diagrams: These allow comparison of the variation of a large number of element between samples. In this type of diagram the symbols for the elements appear on the X-axis, whereas their concentrations appear on the Y-axis on a logarithmic scale. The elements are usually arranged in order of increasing incompatibility in a basaltic system (right to left on the X-axis). Two variants of multi-element diagram are used here: normalised REE diagrams and normalised multi-trace element diagrams (also known as extended REE diagrams or spidergrams). Normalisation filters off intrasample variations (otherwise seen as troughs or peaks) caused by differences in the absolute abundances of elements. By using a logarithmic scale on the Y-axis, the *relative* enrichment or depletion in a particular element between different samples is easily observed.



Figure 5.3: Mg # plotted against Ni ppm on a logarithmic scale for samples with fusion major elements available.

Several compositions have been proposed in the past as normalising factors for multi-element diagrams. The most widely accepted are the average chondrite, the average Normal Mid-Ocean Ridge Basalt (N-MORB) and the average primitive upper mantle (PUM). The normalising values used here are those of the average chondrite (Boynton, 1984; Thompson *et al.*, 1984; Sun & McDonough, 1989) except for the elements K, P and Ti which are taken from Thompson *et al.* (1984). These values are listed in Appendix 3 (p.305).

The profiles on multi-element diagrams can be used to constrain partial melting and crystallisation processes, and also to detect mobility of elements due to

alteration within a suite of samples. During partial melting, incompatible elements will be much more efficiently extracted from the source than the compatible elements. At small degrees of melting, the melt will be enriched in these elements relative to the mantle source region. For N-MORB, this results in multi-element profiles for the liquids that slope downwards from left to right. With increasing degree of partial melting, the profiles are expected to rotate towards the source. This is due to the fact that larger extents of melting eventually bring about greater dilution of the concentrations of the most-incompatible elements as melt segregates from the source and accumulates (Kostopoulos & James, 1992). The actual behaviour of the profiles during melting processes will be strongly dependant on the style of melting and the initial composition of the source.

The multi-element diagrams are often used to describe the degree of enrichment or depletion in basalts. If parameter X is used in normalised form it is symbolised as  $X_n$ . For example, the normalised enrichment factor of the Light REE (LREE) over the HREE is commonly expressed as  $[La/Yb]_n$ . When this ratio is greater than one, the samples are described as LREE-enriched, and when it is less than one, they are described as LREE-depleted, relative to the normalising composition.

Following the same rationale as for partial melting, fractional crystallisation of a parental magma will also effect the profiles on these diagrams. However, fractional crystallisation generally increases the abundance of all the incompatible elements simultaneously, producing parallel profiles. It is only when fractional crystallisation becomes extreme that the ratios between the incompatible elements in these diagrams are affected (i.e. the reverse of partial melting).

There are some cases where elements are strongly partitioned into major crystallising phases (e.g. Sr and Eu into plagioclase), or into accessory minerals (e.g. Ti into magnetite). If such a mineral phase fractionates from a liquid, troughs, or negative anomalies, are observed in the multi-element profiles for the elements concerned. Conversely, if such a mineral phase accumulates in the liquid, peaks, or positive anomalies, are observed. In this way, these diagrams may be used to detect the extraction or accumulation of some mineral phases. Alteration processes cause enrichment or depletion (i.e. peaks or troughs) of mobile elements in the normalised multi-trace element profiles. There may be great variation in these elements for a suite of altered samples which is neither consistently enriched nor depleted (e.g. Ludden *et al.*, 1982; Bartley, 1986).

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3) Ternary normative diagrams: Ternary projections using CIPW-norm and similar calculations have long been used in igneous petrology to classify rocks (Muir & Tilley, 1964) and to depict melting paths or liquid lines-of-descent (e.g. O'Hara, 1968; Thompson, 1987). Projection from a particular phase implies saturation of the liquid in that phase, and it is common to project basalts from plagioclase onto various normative ternary diagrams comprising olivine (ol), nepheline (ne), hypersthene (hy), diopside (di) and quartz (qz). Rocks projecting into the di-hy-ol field are called olivine tholeiites, whereas those projecting into the di-hy-qz field are called quartz tholeiites. Rocks projecting onto the di-ol-ne field are generally silica-undersaturated basalts, and are called alkali olivine basalts and basanites. In these ternary projections, the experimental defined 1-atmosphere cotectic (Walker *et al.*, 1979) for the assemblage of olivine, plagioclase, clinopyroxene plus liquid is usually indicated. Samples that fall on or near this cotectic are generally regarded as products of shallow-level crystallisation (<2 kbars).

# 5.3) Overall compositional variation

## 5.3.1) Whole-rock compositions

In Figures 5.4 and 5.5, major element oxides (wt.%) and selected trace elements (ppm) are plotted against Ni (ppm) in variation diagrams with a semilogarithmic scale. There is a wide range in Ni (13-381 ppm) but most samples lie between 50-100 ppm, forming a dense cluster. This is emphasised by the larger number of samples analysed for certain elements (e.g. FeO\* in Figure 5.4c). There is considerable scatter for all element oxides and elements at a given degree of differentiation (i.e. Ni value), but it is more pronounced for  $Al_2O_3$ , FeO\*, CaO,  $Na_2O$ , TiO<sub>2</sub>, MnO, Sr, V and Sc. The scatter cannot be explained by alteration, as the affected samples are excluded from these plots, nor by analytical error which, in most cases, is less than the size of the data point in these diagrams.

Despite the scatter in Figures 5.4 & 5.5, the abundances of compatible element oxides and trace elements generally decrease with decreasing Ni (e.g. MgO and Cr), while the incompatible element oxides and elements increase (e.g. FeO\*, Na<sub>2</sub>O, K<sub>2</sub>O, TiO<sub>2</sub>, MnO, P<sub>2</sub>O<sub>5</sub>, Nb, Sr, Zr and V). In Figure 5.4a, SiO<sub>2</sub> increases only slightly until Ni drops below 20 ppm, at which point SiO<sub>2</sub> rises sharply. The trends displayed by FeO\*, TiO<sub>2</sub> and V change between 20-30 ppm Ni (Figures 5.4c & 5.5d). The trend for Al<sub>2</sub>O<sub>3</sub> is poorly defined, but generally decreases slightly with decreasing Ni (Figure 5.4b). The abundance of CaO increases first before decreasing as Ni falls below 100 ppm and, although there is considerable scatter for Sc, a similar trend is seen there (Figures 5.4e & 5.5e).



Figure 5.4: Major element oxides (wt. %) plotted against Ni (ppm) for whole-rock samples. Precision error bars (20), as assessed by the deviation of replicate analyses on International Reference Materials and internal monitors, are indicated in the bottom left-hand corner of the diagram, when greater than the size of the data point. Error (20) for Ni is 12 ppm (all values are tabulated in Appendix 3, p. 308).



Figure 5.5: Selected trace elements (ppm) plotted against Ni (ppm) for whole-rock samples. Precision error bars (20), as assessed by the deviation of replicate analyses on International Reference Materials and internal monitors, are indicated in the bottom left-hand corner of the diagram when greater than the size of the data point. Error (20) for Ni is 12 ppm (all values are tabulated in Appendix 3, p. 308).

The trends in these diagrams can be used to indicate which mineral phases have fractionated from the liquids represented by these samples (Section 5.2). The decrease in the abundance of the compatible element Cr suggests fractionation of chrome-spinel, and the decrease in MgO and Ni indicates that olivine has fractionated from the liquids. Both these minerals are observed as phenocryst phases with or without feldspar (Chapter 4). Samples with lower MgO and Ni have fractionated more olivine than samples with higher MgO and Ni. The change in the trend of CaO, coupled with Sc, at 100 ppm Ni is likely to indicate the addition of clinopyroxene to the crystallising assemblage. The change in the trends of FeO\*, TiO<sub>2</sub>, MnO and V at

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approximately 20-30 ppm Ni indicates a further change in the fractionating assemblage. All these oxides (and V) partition into magnetite, and the removal of this non-silicate phase also causes the sharp increase in the abundance of  $SiO_2$ . The oxides and trace elements, Na<sub>2</sub>O, K<sub>2</sub>O, P<sub>2</sub>O<sub>5</sub>, Nb and Zr, remain incompatible within this suite of rocks. Collectively, these observations are in accordance with the expected liquid lines-of-descent for a suite of basaltic to andesitic igneous rocks (Section 5.2).



Figure 5.6: Chondrite-normalised multi-trace element diagrams for 20 samples with the full chemical data set. All elements are normalised to chondrite, except K, P and Ti, which are from Thompson et al. (1984). Samples A550, discussed in the text, is labelled accordingly and plotted in bold. [Some altered samples are included here].

In Figure 5.6, some samples are plotted on a chondrite-normalised multi-trace element diagram. Certain elements display considerable variation relative to adjacent elements (e.g. Rb, K, Sr, P, Eu and Ti); some of these are known to be mobile during hydrothermal alteration (e.g. Rb, K and Sr) (Section 5.2). There is some correlation between the degree of variation in these mobile elements and the amount of interstitial glass observed in thin section. The differences in the abundances of Ti and Eu may be explained by the precipitation or accumulation of accessory minerals phases for Ti, and feldspar for Eu. However, the variable negative P anomalies are harder to explain. The trend in Figure 5.4j suggests that precipitation of apatite has not been involved in the fractionation of these liquids and that P is behaving as an incompatible element. This indicates that there may be slight mobility of the P in these samples, which is supported by the degree of scatter in Figure 5.4f which is equal to that in Figure 5.4g for K<sub>2</sub>O. One extremely glassy and altered sample has extreme enrichment in P<sub>2</sub>O<sub>5</sub> (samples A497, Appendix 4, p. 328). Such P mobility

has recently been reported for boninites from the western Pacific (Murton *et al.*, 1992). The behaviour of Sr in these profiles in Figure 5.6 could be due to either element mobility, or feldspar precipitation and/or accumulation.



Figure 5.7: Chondrite-normalised REE diagrams for 20 samples [Some altered samples are included here]. They are divided into four groups depending on their degree of enrichment, expressed as [La/Yb]n. All elements are normalised to chondrite (Boynton, 1984). Sample A550, discussed in the text, is labelled accordingly. The dashed line for this sample indicates the estimated profile for the LREE had it not experienced alteration (see text).

Figure 5.7 illustrates the variation in the chondrite-normalised abundances of REE. For clarity, the samples are separated into four groups with similarities in their REE patterns. The samples range from very LREE-enriched to slightly LREE-depleted ([La/Yb]<sub>n</sub> ranges from 5 to <1), but the LREE-enriched samples are the most abundant (Figure 5.7a & b). The most LREE-depleted sample (A550) of REE-group 4 appears to have a concave-upwards pattern between Nd and La (Figure 5.7d). This sample appears fresh in handspecimen and, in thin section, the constituent groundmass minerals (olivine and plagioclase) are fresh. However, the interstitial glass in this sample is altered to a combination of smectite and chlorite. Within the outcrop from which this sample was collected there is locally intense zeolite alteration and replacement of the hyaloclastite matrix. These observations suggest that this LREE enrichment in this sample is the effect of element mobility of La and Ce from

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within the interstitial glass. This sample also shows extreme relative enrichments in K, P and Sr in the chondrite-normalised multi-trace element diagram in Figure 5.6. The dashed lines in Figure 5.7d mark the predicted pattern of this sample had it not experienced LREE enrichment by alteration. Mobility of LREE has also been observed and described by Thompson (1973), Frey *et al.* (1974), Wood *et al.* (1976) and Bartley (1986).



Figure 5.8: Major element oxides (wt. %) plotted against Mg # for the volcanic glass samples.

#### 5.3.2) Volcanic glass compositions

In Figure 5.8, the major element oxides are plotted against Mg # for the volcanic glass samples, which ranges from 67-23 and clusters around 40-55. In this cluster there is also the maximum variation in the abundance of other elements. The compatible element oxides decrease with decreasing Mg # (e.g.  $Al_2O_3$  and MgO), whereas the incompatible element oxides increase (e.g. FeO\*, Na<sub>2</sub>O, TiO<sub>2</sub> and K<sub>2</sub>O). The element oxide CaO behaves as a compatible element in Figure 5.8e, which is in contrast to Figure 5.4e. This suggests that lower CaO, high Mg # basalts, that define the negative portion of the trend in Figure 5.4e for Ni-CaO, have not been sampled for volcanic glass.



Figure 5.9: Ternary normative diagram projected from plagioclase onto the plane ol-ne-hy-di-qz. Whole-rock and volcanic glass samples are plotted as squares and stars, respectively. Only unaltered samples are included in this plot. The 1-atmosphere cotectic is indicated (thick line with arrow pointing towards increasing extent of crystallisation) from Walker et al. (1979), and the thinner lines indicate the scatter of the original data published by these authors and Biggar (1983), referred to here as the cotectic-envelope.

## 5.3.3) CIPW normative compositions

In Figure 5.9, the samples are plotted on a ternary normative diagram projected from plagioclase onto the ne-ol-di-hy-qz plane. The 1-atmosphere cotectic is indicated (Section 5.2). The thinner lines parallel to the cotectic mark the boundaries of scatter within the original experimental data set of Walker *et al.* (1979) and Biggar (1983), and is described here as the *cotectic-envelope*. All samples are *hy*-normative (except one), which classifies them as members of the tholeiitic series. The

majority are *ol*-normative, falling within the di-hy-ol field (olivine tholeiites) and only 14% are *qz*-normative, lying in the di-hy-qz field (quartz tholeiites). Most of the glass data occur within the cotectic-envelope, but the whole-rock compositions show a large degree of scatter.

# 5.4) Compositional variation amongst petrological types

## 5.4.1) Petrological types and whole-rock samples

There are two main reasons for examining the compositional variation in terms of the petrographic character of the sample. Firstly, the whole-rock analysis provides a composition for the liquid magma *plus* any phenocrysts present at the time of eruption. If phenocrysts have been added to the magma, they may displace its composition from the true liquid line-of-descent. Field observations and petrographic evidence (Chapter 4) suggests that such addition may have occurred in the form of cognate (or accidental) xenocrysts and the chemical effects of this need to be established. Secondly, amongst the porphyritic samples there are two distinct and mutually exclusive suites, referred to as petrological types 2 and 3 (Table 4.3, p. 87). Computer modelling (Section 4.5.4) in conjunction with the exclusive nature of the two petrological types implies that they are not the product of a single evolutionary path. If this is the case, it may be reflected in the composition. In this following section, a concise summary is provided of key observations made from Figures 5.10-5.12 for each petrological type of basalt in the form of whole-rock samples.

Most of the **petrological-type 1a** samples are aphyric with groundmass olivine. They have a fairly uniform composition for both the major oxides and trace elements and they form a cluster, as seen in Figure 5.10c & 5.11a. However, some more fractionated examples also exist (Ni <50 ppm). The chondrite-normalised multitrace element diagrams in Figure 5.12a show that the samples have persistent negative Sr anomalies. This figure also illustrates that the samples have variable degrees of enrichment, relative to chondrite, in the more incompatible elements, and the degree of LREE enrichment, expressed as  $[La/Yb]_n$ , which varies between 1.6 and 4.8. The **petrological-type 1b** samples are aphyric without abundant groundmass olivine. They are generally highly fractionated as a group (Ni <80 ppm) and include the most fractionated samples of the suite (Ni 10-20 ppm) (Figures 5.10 & 5.11b). They overlap in element abundances with the more fractionated samples of petrologicaltype 1a.



Figure 5.10: Variation diagrams for selected oxides (wt. %) and trace elements (ppm) plotted against Ni (ppm) for the whole-rock samples differentiated into petrological types (Table 4.3, p. 87). A trajectory representing 50% addition of feldspar megacryst (crystal A416F3) to an aphyric sample (A247) using a mass balance equation is plotted on all of the diagrams (see text for details).



Figure 5.10: continued.



Figure 5.10: continued.

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**Petrological-type 2** samples are those that have megacrystic feldspar. They have a very large degree of scatter for most oxides and trace elements at a given degree of differentiation (Figure 5.10). The group generally has higher  $Al_2O_3$ , CaO and Sr, and lower FeO\* and V abundances than the other petrological types of basalt. They have a limited range of Ni that is similar to petrological-type 1a samples (Figure 5.11). The scatter observed in the variation diagrams (Figure 5.10) for certain elements suggests an influence of the highly calcic plagioclase megacrysts (An 80-90 mol %) on the whole-rock compositions of the samples. To test this further, the effect of progressive addition of feldspar megacrysts to an aphyric basalt (petrological type 1a) was modelled using a mass-balance equation (Table 5.1). The major element oxides in the feldspar composition are from a feldspar crystal analysed in this work (Appendix 2A, p. 269), and the trace element abundances are from published calcic feldspar compositions (Philpotts & Schnetzler, 1970).



**Figure 5.11:** Bar chart diagrams illustrating the distribution of Ni amongst the various petrological types (defined in Table 4.3, p. 87).

The equation adds the composition of a calcic plagioclase (An 86) to the composition of the aphyric sample in increments of 5% by weight. As the aphyric sample is progressively contaminated with calcic feldspar, the bulk composition changes. These effects are indicated in Figure 5.10 by the trajectories of element abundances for up to 50% feldspar addition. The petrological type 2 samples closely follow these modelled trajectories for certain elements (e.g.  $Al_2O_3$ , CaO, FeO\* and

V), and provides an explanation for some of the scatter in Figure 5.10. Samples with approximately 40% modal feldspar megacrysts occur at the 40% marker on the trajectory in Figure 5.10. The modelling indicates that the amount of megacrysts must exceed 5% by weight to have a significant effect on most elements.

	Whole-rock	Feldspar megacryst					
	sample A247	(A416F3) from sample					
	F • • • • • • •	A416					
SKO2	48.56	46.27					
AI203	13.29	32.87					
F62U3	13.59	0.48					
MgU Geo	5.76	0.62					
	12.00	1.21					
NE2U KOO	2.11	1.30					
N2U Tico	0.22	0.00					
1102	2.16	0.00					
HINU DOOD	0.22	0.05					
P205	0.25	0					
Mg#	47.1						
Nb	18						
Zr	129						
Y	31.3						
Sr	188	(381)					
Zn	107	(,					
NI	77						
v	394						
Cr	186						
Ga	19						
Sc	41						
Co	61						
Ros	2.5	(1.84)					
Ba	64	(23)					
La	11.2						
Ce	27.1	(3.52)					
Pr	3.7						
Nd	17.9	(2.04)					
Sm	4.6	(0.467)					
Eu	1.6	(0.401)					
Gđ	5.6	(0.627)					
Dy	6.1	(0.573)					
Ho	1.1						
Er	3.3	(0.42)					
Yo	3.1	(0.489)					
	0.5	(0.085)					
Mass balance equation used:[((1-n)*CiWR)+(n*CiFM)]							
where: n=percentage of feldspar megacryst added							
CiWR=concentration of element (wt. % or ppm) in whole-rock							
CiFM=concentration of element (wt. % or ppm) in feldspar megacryst							

**Table 5.1:** Compositions used in the modelling of feldspar megacryst addition. The whole-rock composition is from sample is A247 (Appendix 4, p. 315) and the major element oxide data for the feldspar megacryst is from sample A416 (Appendix 2A, p. 269 - A416F3). The trace element data for the feldspar megacryst are from published analyses of feldspar with a similar major element composition to that of the crystals in sample A416 (Philpotts & Schnetzler, 1970). These elements are bracketed. The modelling was done using a mass-balance equation defined in this table.

In chondrite-normalised multi-element diagrams, petrological type 2 samples might be expected to show positive Sr and Eu anomalies, because these elements partition into plagioclase. In Figure 5.12b, such subtle positive anomalies are observed for Eu in most of the samples, but they are not always accompanied by positive Sr anomalies. Sr is the only element for which the modelled trajectory does not match the composition of the petrological type 2 samples in Figure 5.10. However, the abundance of this element in plagioclase was estimated from published data, which may have been lower than the actual amount of Sr present in crystal A416F3. Nevertheless, the Sr anomalies observed for the petrological type 2 samples in Figure 5.12b will be dependent on the original form of the Sr anomaly of the starting magma composition (i.e. positive or negative), as well as the percentage of feldspar megacryst added to the magma prior to eruption. Figure 5.12b also illustrates that there are variable degrees of enrichment in the incompatible element ratios, relative to chondrite, for this petrological type of basalt; [La/Yb]<sub>n</sub> ranges between 0.92 and 4.48.

Overall, petrological type 2 basalt has a very large degree of compositional variation, both in the abundance of major element oxides, trace elements and the degree of enrichment in the incompatible element ratios, relative to chondrite. The modelling and geochemical observations suggest that most of the petrological type 2 samples represent essentially aphyric magmas that have had variable amounts of feldspar megacrysts added to them; a conclusion that is supported by petrographic and field evidence (Chapter 4).

Petrological-type 3 samples, which lack feldspar megacrysts and intratelluric clinopyroxene phenocrysts, display the largest range in the degree of differentiation (Ni 50-381 ppm) and are more evenly distributed within this range than the other petrological groups (Figures 5.10 & 5.11). In Figure 5.10, the majority of the samples have Ni>100 ppm and show considerable scatter for all elements at a given degree of differentiation. This is emphasised by the larger number of samples in Figure 5.10c, for example. The group shows an increase in the abundance of CaO with decreasing Ni, until Ni falls below 100 ppm. This complies with the absence of clinopyroxene as an observed phenocryst phase in these basalts (Section 4.3). In the chondrite-normalised multi-element diagrams (Figure 5.12c) this petrological type of basalt has variable anomalies in Sr. There is a large range in the degree of enrichment in the incompatible element ratios, relative to chondrite; variation in ([La/Yb]<sub>n</sub>) is between 0.8 and 3.25.



Figure 5.12: Chondrite-normalised multi-trace element diagrams for samples differentiated according to petrological type (Table 4.3, p. 87). [Some altered samples included].

Petrological type 3 basalt also has a large degree of compositional variation, proportional to that of petrological type 2 basalts, yet this group does not contain phenocrysts that are exceptionally abundant or consistently out of equilibrium with their host by such significant amounts as the megacrysts of petrological type 2 basalt

(Section 4.5). The compositional diversity of petrological type 3 whole-rock samples cannot, therefore, be accounted for by phenocryst accumulation alone and another reason is required to explain the compositional scatter amongst these samples.

## 5.4.2) Petrological types and volcanic glass samples

The volcanic glass analyses provide an opportunity to examine the composition of the magma without the interference of phenocrysts, and may provide insight into the origin and/or subsequent differentiation history of the magmas constituting the two porphyritic petrological types, 2 and 3. In Figure 5.13, the glass samples are differentiated according to the petrological types, and plotted for selected major element oxides against Mg #. As in the whole-rock compositions, most of the petrological type 1a and 2 samples fall within a narrow range of differentiation (Mg # 40-55), type 1b samples are generally more fractionated (Mg # <45) and type 3 samples have the largest range in Mg # (42-67). Within the cluster of samples between Mg # 40 and 55, petrological types 2 and 3 are separate for SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub> and FeO\* (Figure 5.13a, b & c) and also MgO, Na<sub>2</sub>O, K<sub>2</sub>O and TiO<sub>2</sub> (not shown), at a given Mg #. In this respect, within this cluster, they can be said to follow separate liquid lines-of-descent. The petrological type 1 samples overlap with both petrological types 2 and 3. The division amongst samples from different petrological types is not, however, observed for CaO (Figure 5.13d).

The overall trend for  $SiO_2$  displayed by petrological type 3 samples is not easily explained by crystal fractionation processes within a group of cogenetic samples, and this suggests that samples of petrological type 3 basalt are not all related to each other by fractional crystallisation. Similarly, a sample of petrological type 1b basalt at Mg # 42 is on the liquid line-of-descent of petrological type 2 basalts for SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub>, but this sample is in line with the petrological type 3 samples for FeO\*. This suggests variable ratios between the major elements within these glass samples of the various petrological types. In summary, the volcanic glass data suggest that the composition of the liquids that form the porphyritic petrological types (2 & 3) *are* distinct at a given degree of differentiation and that they are not related to each other by fractional crystallisation. There are also indications that the samples of petrological type 3 are not cogenetic which is supported by the scatter of these samples in Figure 5.10. This implies that they require the presence of multiple parental melts.



Figure 5.13: Selected major element oxides (wt. %) for the volcanic glass samples plotted against Mg #. The samples are differentiated according to the petrological type of basalt (Table 4.3, p. 87). Dashed lines indicate possible parallel liquid lines-of-descent. Key is the same as Figure 5.10.

## 5.4.3) Petrological types and their CIPW normative compositions

In Figure 5.14, the whole-rock and volcanic glass normative compositions are differentiated according to the petrological type of the samples. The whole-rock samples of petrological type 1a fall just below the cotectic-envelope (Figure 5.14a), whereas the limited number of glass compositions for this petrological type fall within the envelope (Figure 5.14b). The petrological type 1b samples are dominantly qz-normative in composition and lie on the 1-atmosphere cotectic for both the whole-rock and glass samples (Figure 5.14a & b). The petrological type 2 whole-rock samples mostly lie above the cotectic-envelope in the clinopyroxene phase field, displaced towards the di apex, and with only a few samples within the envelope or below it (Figure 5.14a). However, the volcanic glass analyses for this petrological type 3 whole-rock samples show the most variation in normative composition (Figure 5.14d)

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and are relatively close to the ol apex. In the case of the sample closest to this apex (sample A34) there is petrographic support for olivine accumulation. The more fractionated samples lie on or just below the 1-atmosphere cotectic. The glass samples of petrological type 3 have a relatively narrow compositional range and mostly fall within the cotectic-envelope.



Figure 5.14: Ternary normative diagrams projected from plagioclase onto the plane ol-ne-hy-di-qz. The samples are differentiated according to their petrological type (defined in Table 4.3, p. 87): a) whole-rock samples. b) volcanic glass samples. The 1-atmosphere cotectic is indicated (thick line with arrows) from Walker et al. (1979), and the thinner lines indicate the scatter of the original data from these authors and Biggar (1983) (referred to here as the cotectic-envelope).

The difference between the whole-rock and glass samples may reflect the mineralogical control on the compositions of these samples. Abundant olivine in type 3 basalt displaces the composition towards the ol apex, and the feldspar megacrysts in type 2 basalt displaces the composition towards the di apex in this particular projection. However, the aphyric petrological type 1 samples also have different normative compositions, depending on whether the sample is of whole-rock or glass type. This suggests that all the whole-rock samples may contain accumulated crystals, and indicates that they are unsuitable to illustrate the liquid trends of crystallisation in these ternary normative diagrams. Amongst the glass samples, petrological type 3 basalt falls closer to the di-ol join than those of petrological types 1 and 2 (Figure 5.14b). This indicates that petrological types 1 and 2 could be the results of fractional crystallisation from petrological type 3 parents at shallow levels.

# 5.5) Regional and stratigraphic compositional variation

## 5.5.1) Introduction

In the previous section (Section 5.4) it was established that the two porphyritic suites (petrological types 2 & 3) are not directly petrogenetically related to each other by fractional crystallisation. In the field, basalt of petrological types 1 and 2 occurs in the fissure-type eruptive units, with the petrological type 2 basalt generally restricted to those in the vicinity of the central volcanoes, and petrological type 3 basalt forms the lava shields (Section 4.4). This spatial distribution suggests that the two suites may be the result of two different differentiation paths through the crust. It was also established that the samples of petrological type 3 basalt may not be petrogenetically related to each<sub>A</sub> (Figures 5.10, 5.12 & 5.13). To investigate this further, the compositional variation within the stratigraphy is described for the Hveragerði to Ingolfsfjall region, and within and between the lava shields. These observations may constrain the differentiation paths followed by the various petrological types of basalt through the crust.

## 5.5.2) Hveragerði Volcanic System

In this section, the compositions of the eruptive units within the Hveragerði Volcanic System, other than the morphologically distinct lava shield of Ingolfsfjall, are examined. The units are dominantly composed of petrological type 1 and 2 basalt from fissure-type eruption events, with only minor occurrences of petrological type 3 basalt in the form of buried lava shields (e.g. unit DA1) (Chapters 3 & 4). It is also possible that there may be other examples of buried lava shields that have not been

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recognised within the volcanic pile (e.g. unit GB4) which are also of petrological type 3 basalt (Figure 4.10, p. 91). Fieldwork has provided constraints on the spatial and temporal relationships of these units (Chapter 3; Figure 3.2, p. 43, and Maps A1 & A2), and so the compositional variation can be examined in these terms.



Figure 5.15: Nb plotted against Ni for eruptive units that are stratigraphically well-constrained. For the majority of the units, each data point represents an average of multiple samples. A full stratigraphic log is provided in Figure 3.2 (p. 43).

In Figures 5.15-5.17, the incompatible element Nb is plotted with the compatible element Ni for the whole-rock compositions of the eruptive units. These elements are used to represent the degree of systematic variation in composition and monitor the degree of differentiation. Magmatic processes that operate in a crustal magma reservoir, and the shape and form of that reservoir, affect the degree of differentiation of the magmas. It follows, therefore, that the extent of differentiation and the overall variation in this parameter within the stratigraphy may be used to study the processes and form of the reservoirs from which the basalts were erupted, and how they have changed with time. Figures 5.16 and 5.17 illustrate the compositional variation for Ni and Nb within several stratigraphic profiles which are each spatially separate but overlap in time. However, most of the units in Figure 5.16 are generally older than those in Figure 5.17, which represents the Grensdalur Central Volcano (see Section 3.3).



Figure 5.16: Plot of Ni and Nb for samples arranged in stratigraphic order (Figure 3.2, p. 43) for the region to the west of Reykir, as far east as Ingolfsfjall. Also indicated are the petrological type, constrained age horizons, some stratigraphic marker horizons (arrow heads) and the occurrences of buried lava shields (BLS) (see Chapters 3 & 4). Poorly constrained eruptive units within the stratigraphy have been excluded.



Figure 5.17: Plot of Ni and Nb for samples arranged in stratigraphic order (Figure 3.2, p. 43) for the region to the east of Reykir, including the Grensdalur Central Volcano. Also indicated are the petrological type, constrained age horizons, some stratigraphic marker horizons (arrow heads) and the occurrences of buried lava shields (BLS) (see Chapters 3 & 4). Poorly constrained eruptive units within the stratigraphy have been excluded.

Most of the stratigraphic sub-groups presented in Figure 5.15 have very uniform abundances of Ni and Nb and form tight clusters (e.g. sub-groups GA), and some sub-groups overlap completely with others (e.g. GA and many of DB units in Figure 5.15). The tendency for units within a group to cluster diminishes with time, and the younger units from group S are the most compositionally varied (Figures 5.15 & 5.17). This group includes the more fractionated units in the region, but the degree of differentiation is highly variable.

In the stratigraphic profiles of Figures 5.16 and 5.17, the periods of uniform composition (e.g. sub-groups GA, TA and most of DB), and the periods of diverse compositions (e.g. sub-groups TB, DA and group S) are more apparent than in Figure 5.15. Some units are noticeably different in both Ni and Nb to those directly above and below (e.g. units DB7b and S1; Figure 5.17). These could be described as compositional breaks, or interruptions, within the stratigraphy that consist of one or more eruptive units that are each unique in composition. In some cases, the composition before this break is resumed by subsequent eruptions (e.g. after unit DB7b in Figure 5.17), but in other cases, the composition after the break marks the start of a new sequence of units with a different uniform composition. For example, the units TB and DA (Figure 5.17) mark a period of diverse composition, after which, the DB units are very uniform in Ni and Nb, and different in Nb to the older TA sub-group.

The episodes in the stratigraphy that exhibit the most compositional variation are dominated by either petrological type 2 or 3 basalt, while periods of uniformity are dominantly aphyric petrological type 1 basalt (e.g. TA). A pattern emerges through the stratigraphy where the composition changes from being monotonous to a sequence of eruptive units that are varied, and these periods are dominated by different petrological types of basalt. The degree of, and variation in, differentiation becomes greater in the younger units of the Hveragerði Volcanic System, in the vicinity of the extinct Grensdalur Central Volcano.

## 5.5.3) Lava shields

Each lava shield is a separate eruptive event that often occurs away from the main locus of magmatic activity (Jakobsson *et al.*, 1978). The shields are composed of petrological type 3 basalt (Section 4.4); approximately 90% of the petrological type 3 basalt can be assigned to them and the remaining 10% occur in units that are too poorly-exposed to determine conclusively the type of volcanic event from which they were extruded (e.g. unit GB4). The texture displayed by feldspar is very similar

for these basalts and yet their compositions are extremely varied. In this section, the compositional variation within and amongst the lava shields is described in order to explain the compositional diversity of petrological type 3 basalt; and to constrain the differentiation history of each lava shield. In Figure 5.18, the compositional variation of the lava shields is displayed for selected major element oxides and trace elements. The morphologically distinct lava shields are separated from the buried examples for clarity. Samples from the Hengill eruptive unit are also included because this eruptive unit has morphological and petrographical characteristics that typify the lava shields (Sections 3.4 & 4.4.3).

The Ingolfsfjall lava shield (ILS) has high abundances of FeO\* and incompatible trace elements (e.g. Nb & Sr), and low CaO compared to the other lava shields. There is much internal compositional variation within this shield. The Hæðir lava shield is distinct with the lowest abundances of FeO\*, Nb and Sr, and higher CaO for a given Ni. The Húsmúli shield is also distinct because it is consistently more fractionated than the others (Ni <66 ppm). It is also unusual because the lavas generally lack olivine as a phenocryst phase and have rare intratelluric clinopyroxene phenocrysts (Section 4.4.3). It is the only occurrence of petrological type 3c basalt in the region (Table 4.3, p. 87). Amongst the buried lava shields, the DB7b shield overlaps with Ingolfsfjall for most major element abundances (e.g. FeO\*), but less so for the trace elements (e.g. Nb and Sr). The **BB** shields are each represented by a sequence that changes from olivine-rich compound pahoehoe lavas at the base (units BB4 and BB6) to simple flows of appyric lavas at the top (units BB5 and BB7). It is not conclusive from the field or petrographic evidence, however, as to wether these evolved basalts belong to the BB4 and BB6 lava shields (Section 3.3.2.2). The latter are considerably evolved in composition (BB5, Ni <60 ppm; BB7, Ni <26). The DA1 and **BA3** lava shields are similar in composition to Hæðir with low abundances of incompatible trace elements (e.g. Nb and Sr) and high CaO.

Within the Hengill Volcanic System, the main Hengill mountain is reported to be the Hengill Central Volcano (Björnsson & Hersir, 1981; Foulger & Toomey, 1989; Trønnes, 1990). It contrasts with other Icelandic central volcanoes, which are described as repeated eruptions at one site (Sæmundsson, 1979), a characteristic that is clearly evident in the extinct Grensdalur Central Volcano (Section 5.5.2). The Hengill mountain, however, represents just one eruptive event (Árnason *et al.*, 1987) built up of approximately 8 km<sup>3</sup> of petrological type 3 basalt. The samples analysed here are similar in composition to the large-volume shield of Ingolfsfjall (Figure 5.18). This provides further evidence to suggest that the Hengill mountain represents a subglacial lava shield-type eruption and not the "Hengill Central Volcano".



Figure 5.18: Selected major element oxides (wt. %) and trace elements (ppm) plotted against Ni (ppm) for various lava shields in the region. The Hengill mountain is also included. The morphologically distinct shields are plotted in ai)-di) and the buried lava shields are plotted in aii)-dii) for clarity.



Figure 5.19: Vertical profiles through the lava shields showing the systematic variation in selected elements: a) BB4-5 shield included in profile IL (Appendix 1A, p.253); b) BB6-7 shield included in profile IX (Appendix 1A, p. 255); c) IS profile from the Ingolfsfjall lava shield (Appendix 1A, p. 256); d) Mógil gorge profile of Húsmúli; e) Purárhnúkur cliff profile from the DB7b lava shield where there are three sub-units (Appendix 1A, p. 238); f) Hæðir profile from a gorge near Illagil. The error bars are indicated where they exceed the size of the data point.
Collectively, the lava shields in Figure 5.18 display much variation in both the differentiation index and other element abundances. However, each individual lava shield has a distinctive composition with respect to abundance of any element at a given Ni value. Amongst the nine shields presented here (including the Hengill eruption), however, there are some that overlap completely in Ni and other elements (e.g. the two BB lava shields). There appears to be a relationship between the volume of the lava shields (where determinable) and their composition at a given Ni value (e.g. the large-volume shield of Ingolfsfjall has high FeO\*, and the smaller-volume shield, Hæðir, has lower FeO\*). Such a relationship for other element pairs has been reported between Postglacial lava shields from the Reykjanes Peninsula (Jakobsson *et al.*, 1978).

Each lava shield has some degree of compositional scatter, as observed in Figure 5.18, which suggests considerable intrashield variation. Many of the shields have been sampled from vertical lava profiles (Appendix 1A), and the nature of the compositional variation within these profiles provides insight into the variation with respect to time. Following the same logic as in Section 5.5.2, such variation may shed light on the type of magmatic reservoir and the processes operating there. In Figure 5.19a-f, the samples from each shield are examined in their position within the vertical profile of the sampled stratigraphic sections for FeO\*, TiO<sub>2</sub>, Ni and Nb. Some vertical profiles are uniform with variation that is within analytical error (2 $\sigma$ ) (e.g. BB6 compound lavas, Figures 17b). Others show oscillations with compositional breaks (i.e. Húsmúli, DB7b, Hæðir; Figure 5.19d, e & f). The trend across these breaks is, in some cases, towards more primitive compositions, and lower incompatible element abundances (i.e. Húsmúli and DB7b; Figure 5.19d & e). However, in others, the composition after the break is the same as it was before the break (i.e. Hæðir; Figure 5.19f).

The compositional breaks in some profiles in Figure 5.19 can be explained by the petrographic character of the sample. In the compound lava succession of BB6 (Figure 5.19b) sample A34 has elevated Ni, but not FeO<sup>\*</sup>, TiO<sub>2</sub> and Nb, relative to lavas directly above and below, and in handspecimen there are significantly more olivine crystals (i.e. crystal accumulation). Similarly, within the Ingolfsfjall profile (Figure 5.19c) FeO<sup>\*</sup> and TiO<sub>2</sub> are lower for sample A21 than the flow units sampled above and below, while Nb and Ni are essentially the same. This sample has noticeably fewer groundmass opaque minerals in thin section relative to the rest of the profile. Within the Hæðir lava shield (Figure 5.19b) there is a thick flow unit (no 5) that is compositionally different from the overlying and underlying flow units with respect to *all* elements. This represents a batch of magma supplied to the eruption that is different. This break, and those in Húsmúli and DB7b, implies a more open style of magmatic system.

A test was carried out to see if the composition of the lavas within one shield could be produced by closed-system fractional crystallisation using the same computer program (TRACE3.FOR; Nielsen, 1988) with the same parameters as those in Section 4.5.4. Fractional crystallisation of sample A474 from the DB7b lava shield (also used in Section 4.5.4; Figure 5.19e, sub-unit 3) was modelled in an attempt to produce the composition of the overlying lavas. In Figure 5.20, the results of this modelling are displayed as a trajectory for 0-40% fractional crystallisation for FeO\*, Nb and the  $[La/Yb]_n$  ratio against Ni. The other samples from the DB7b lava shield are also plotted and labelled according to the three sub-units observed in the field (Section 3.3.2.4; Figure 3.10. p. 60). The dotted-lines join successive samples from within the profile.

It can be seen from Figure 5.20 that the 0-40% fractional crystallisation trajectory does not coincide with the samples positioned above sample A474 in the vertical profile, and this indicates that the compositions of the overlying lavas were not produced by closed-system fractional crystallisation from this starting composition (Figure 5.20a & b). Additionally, following the same logic, sample A474 cannot be the produced from the underlying lavas in this way. The modelling illustrates that simple closed-system fractional crystallisation is not capable of changing the  $[La/Yb]_n$  ratio, and yet there is considerable variation in this parameter between the three sub-units (Figure 5.20c).

Collectively, the observations indicate that the eruptive products of the DB7b lava shield are not result of closed-system magmatism. However, the mineralogy observed within a single flow from the DB7b shield (olivine + chrome spinel phenocrysts) can be reproduced by modelling fractional crystallisation in a closed magmatic system (Section 4.5.4). This suggests that magma coming into the reservoir cannot be of drastically different major element composition in terms of the degree of differentiation. This indicates that the system is not a classic open system magma reservoir that involves extensive mixing between magmas of diverse compositions (after O'Hara, 1977; and others). However, modelling of fractional crystallisation carried out on other lava shields (e.g. Ingolfsfjall) fails to reproduce the observed feldspar compositions, giving much lower An contents (Section 4.5.5). The differences in the incompatible trace element ratios, such as [La/Yb]<sub>n</sub>, between

successive lavas from the DB7b lava shield requires a continuous magma supply to the reservoir that is not mixing before feeding the eruption.



Figure 5.20: Results of modelling closed-system fractional crystallisation (see text) as indicated by the 0-40% trajectory for a) FeO\*, b) Nb and c) [La/Yb]n. Tick marks indicate increments of 2% fractional crystallisation. The samples from the DB7b lava shield are also plotted on this diagram. Successive samples in the vertical profile are joined by dotted lines and arrows indicate the lower and upper samples in the profile. T=Top; B=Base. Units 1-3 in c) refer to those identified in the field (Section 3.3.2.4: Figure 3.10, p. 60).

The compositional variation described above for the lava shields suggests that the magmas are coming directly from the mantle source region and are not ponding in a crustal reservoir. This interpretation allows each lava shield to remain compositionally distinct. The similarity in the petrographic character of the petrological type 3 basalt suggests that nearly all the lava shields sampled here share a similar style of crystallisation process which implies a similar style of magmatic plumbing system. Húsmúli, with its fractionated basalts that lack abundant olivine, is the only exception in the area and requires a unique and slightly different magmatic plumbing system compared to the rest of the lava shields.

# 5.5.4) Summary of regional and stratigraphic compositional variation

The eruption events that constitute the Hveragerõi Volcanic System are compositionally varied. Within the spectrum of compositions, temporally and spatially related units tend to have very uniform compositions. However, the degree of uniformity decreases amongst the younger eruptive units of the Grensdalur Central Volcano, which are each quite unique and some are more fractionated. The lava shields are generally less fractionated, but each is unique in composition at a given degree of differentiation. Modelling and examples of vertical profiles through the shields suggests that they essentially lack a crustal reservoir, with magma coming straight to the surface. It is likely that the main Hengill eruptive unit represents a subglacial lava shield-type of eruption, and not a central volcano.

## 5.5.5) Constraints for the magmatic plumbing system

Following the logic of previous studies, a uniform magma composition suggests a well-mixed magma reservoir (e.g. Natland *et al.*, 1983) and the periods of diverse compositions indicate poor, or no mixing (e.g. Langmuir *et al.*, 1977). Compositions of a more-fractionated nature *without* abundant olivine crystals require a reservoir in which the olivine crystals could have been removed by settling, while the less-fractionated varieties *with* abundant olivine do not require a reservoir. From the compositional and petrographic variation observed in this suite of samples, the following model begins to develop, and is illustrated in Figure 5.21: a) Periods of activity producing uniform petrological type 1 and poorly-phyric petrological type 2 basalt could indicate the existence of a well-mixed reservoir; b) diverse periods of more fractionated basalt, where activity alternates between petrological types 1 and variably-phyric petrological type 2 basalt, may indicate a poorly-mixed reservoir, and one that may have developed pockets of melt isolated by crystal mush (i.e. the megacrysts and associated pyroxene and olivine crystals; Chapter 4); c) compositionally unique eruptions of more primitive petrological type 3 basalt, that



Figure 5.21: Three models for the magmatic plumbing system for the Hengill region, based on geochemical and petrographic constraints.

usually erupt to form lava shields (90%) with abundant olivine, represents a magma that has not passed through a shallow crustal reservoir, and has risen straight to the surface from the base of the crust. The nature of the magmatic plumbing system cannot, however, explain the *causes* of the differences in the compositions of the petrological type 3 basalts at a given degree of differentiation.

Húsmúli is the only exception to the other lava shields in the area. Following the same logic employed in the construction of the models in Figure 5.21, this lava shield requires a magma reservoir in which to fractionate and to deposit its olivine content. However, within the lava profile sampled here, the shield displays compositional breaks which require the input of new melt to the eruption. These melts have more dilute concentrations of incompatible elements relative to those below in the profile (e.g. Nb; Figure 5.19), but are equally fractionated (Figure 5.18). The plumbing system beneath Húsmúli still has to be capable of maintaining the process that produces the radiating habit of the feldspar, common to petrological type 3 basalt (Section 4.3), and it has been suggested here that this process in crystallisation in a deep-seated dyke (Section 4.5.3). This also suggests that Húsmúli has a plumbing system similar to that in Figure 5.21c.

# **5.6)** Petrogenetic constraints on mantle melting processes

# 5.6.1) Observations

The geochemical data presented in Sections 5.3 to 5.5 were used to investigate how crustal differentiation processes, such as alteration, fractional crystallisation and open- and closed-system magmatism, have affected the basalts from the study region. It has been established that the two mutually exclusive, compositionally varied, porphyritic basalts and the aphyric basalts are the result of different routes through the magmatic plumbing system beneath the region (Section 5.5.5; Figure 5.21). In this section, emphasis is now placed on the petrogenesis of the basalts before they start their separate routes. Attention is drawn to lines of evidence that may be used to investigate mantle melting processes, and how such processes contribute to the composition of these basalts. Incompatible element ratios are unaffected by crustal differentiation processes for basaltic magmas (Sections 5.2 & 5.5; Figure 5.20) (Treuil & Joron, 1975) and such ratios will therefore be used to investigate the origin of the eruptive units in terms of melting dynamics and differences in the composition of the mantle source region. Chapter 5



Figure 5.22: Chondrite-normalised REE diagrams for samples from a) the Hverageroi Volcanic System and, b) & c) the lava shields.

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In Figure 5.22, the chondrite-normalised REE patterns are presented for samples from the Hveragerði Volcanic System, the lava shields, and a sample from the main Hengill eruption. In Figure 5.22a, the samples represent different stratigraphic groups and sub-groups (excluding any samples from lava shields), and in Figure 5.22b and c, the lava shields are plotted. The degree of LREE enrichment, relative to chondrite, expressed as [La/Yb]n, and other information about these samples is summarised in Table 5.2. In Figure 5.22a, most of the samples have very similar REE profiles and are LREE-enriched, relative to chondrite, but there are also less-enriched basalts with flat and slightly dipped, LREE-depleted patterns. Two samples from the GA sub-group, an example of a compositionally uniform group of eruptive units (Section 5.5.2), have REE patterns that are indistinguishable from each other. The S group, an example of a compositionally diverse episode in the stratigraphy, has REE patterns that are LREE-enriched (units S5, S15) to slightly LREE-depleted (unit S1). The other LREE-depleted sample is from unit GB4. Figure 5.22b & c shows that there is also variation in the degree of LREE enrichment for the lava shields but, as in Figure 5.22a, most of them are LREE-enriched, with a few LREE-depleted samples, relative to chondrite (e.g. the small volume Hæðir lava shield).

These observations indicate that there is no correlation between the degree of enrichment in the LREE and the regional distribution in the field. The only consistency in the REE signature is for the GA basalts (Figure 5.22a) from a sequence of eruptive units from a single region that also have similar abundances of other elements. This complies with the suggestion made above that this sub-group was erupted from a well-mixed magma reservoir (Section 5.5.5; Figure 5.21a).

Amongst the compositional variation displayed by the lava shields, there is a clear relationship between the volume of the lava shield and the abundances of FeO<sup>\*</sup> and CaO, as exemplified by the morphologically distinct lava shields of Ingolfsfjall and Hæðir (Section 5.5.3; Figure 5.18). In Figure 5.23, the relationship between the abundances of FeO<sup>\*</sup> and the degree of LREE enrichment ([La/Yb]<sub>n</sub>), relative to chondrite, is presented. There is a clear correlation between the abundance of FeO<sup>\*</sup>, and the [La/Yb]<sub>n</sub> ratio, with FeO<sup>\*</sup> increasing with increasing degree of LREE enrichment. These findings are in agreement with those of Elliot *et al.* (1991). The high-FeO<sup>\*</sup> lava shields include the larger volume Ingolfsfjall shield, while the low-FeO<sup>\*</sup> lava shields include the smaller volume Hæðir lava shield. There is, however, some scatter, and samples that fall away from the trend are either the more-

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Figure 5.23: FeO\* plotted against [La/Yb] normalised to chondrite. The samples are plotted according to location within the Hveragerői Volcanic System (petrological types 1 & 2), or in a lava shield (plus other petrological type 3 basalts). Samples with greater than 5% megacrysts are indicated accordingly. Two samples from the GA sub-group, units GB4, ILS and S1 are indicated, and are discussed in the text. The three samples from the DB7b lava shield are encircled by the dot-dash line. On this figure, the trajectories for 40% closed-system fractional crystallisation and 50% feldspar megacryst addition are indicated.

fractionated samples (Ni <90 ppm) (e.g. the GA units) or have significant megacrysts (>5% by weight). These samples can be back-projected onto the main trend along their relevant trajectories. The three samples from the DB7b lava shields are encircled; their relative positions in this diagram emphasise the large degree of variation in the degree of LREE enrichment within a single shield (Section 5.5.3), thus providing further evidence which suggests that this lava shield was not produced by closed-system magmatism (Section 5.5.5).

In summary, the REE signatures (i.e. enriched or depleted, relative to chondrite) are not consistent with the petrological type of basalt (Section 5.4.1) or the type of volcanic eruption event from which the magma was erupted (i.e. lava shield or fissure). Samples from the lava shields show distinct abundances of FeO\* at a given degree of enrichment, as do the other samples from the Hveragerði Volcanic

Hveragerði Volcanic System									
Groups	Sample - units		Petrological	REE-group	[La/Yb] <sub>n</sub>				
			type						
GA	A522 - GA5		1a	3	1.61				
	A529 - GA11		1a	3	1.59				
GB	A550 - GB4		3a	4	0.77				
ТА	A247 - TA4		1a	2	2.43				
D	A142 - DB7a		2b	2	2.73				
	A478 - DB8b		1a	2	2.14				
S	A416 - S1		2a	4	0.92				
	A413 - S5		2b	1	4.48				
	A346 - S15		2b	1	4.02				
Lava shields									
Shield name		Sample		REE-group	[La/Yb] <sub>n</sub>				
BB4-BB5		A546 [BB5]		3	1.83				
BB6-BB7		A36 [BB7]		2	2.56				
Ingolfsfjall		A554		2	2.64				
DA1		A357		2	2.10				
DB7b		A474		2	2.38				
		A4	79	2	2.92				
		A480		2	3.10				
Hæðir		A512		3	1.95				
Húsmúli		A499		2	2.95				
Hengill eruption		A374		2	3.05				

System once the effects of fractional crystallisation and megacryst addition have been allowed for.

**Table 5.2:** The degree of LREE enrichment, expressed as [La/Yb]n, for various eruptive events within the Hveragerői Volcanic System and the Lava shields. The sample numbers, eruptive unit names, petrological types (Section 4.3, p. 87), and REE-groups (Section 5.3, Figure 5.7) are also indicated. All the lava shields are of petrological type 3 basalt (except units BB5 & BB7, which belong to petrological type 1a).

# 5.6.2) Petrogenetic Modelling

## 5.6.2.1) Introduction

The observations made in this chapter and Chapter 4 have gradually focussed attention on some specific problems. These include; 1) the origin of the calcic plagioclase megacrysts prior to their accumulation in the crustal reservoir as cognate xenocrysts; 2) the diversity of compositions displayed by petrological type 3 basalts in the absence of crystal accumulation; 3) the causes of the variability in the REE patterns that do not correlate with field and petrographic observations, but have a strong correlation with abundances of certain major element oxides (e.g. FeO<sup>\*</sup>). In

this section, modelling will be employed to investigate these relationships and problems, in an attempt to further constrain the developing petrogenetic model (Section 5.5.5; Figure 5.21). These problems are addressed by modelling fractional crystallisation of primary melts produced under variable melting conditions, and by modelling the generation of primary melts themselves from variable starting compositions within the source region. The results of the modelling are compared to the observed compositional variation discussed above in Sections 4.5.3, and 5.3 to 5.5.

Melt generation studies involve a dynamic melting column (Langmuir et al., 1977) in which the mantle rises adiabatically and undergoes decompression melting, with a number of variable melting processes in operation simultaneously (McKenzie & Bickle, 1988; Elliot et al., 1991; Kostopoulos & James, 1992). This is schematically represented in Figure 5.24. In this model, the upwelling mantle begins to melt when it intersects its solidus, which is dependant on the potential temperature of the mantle (McKenzie, 1984). The mantle column continues to undergo polybaric melting (Falloon & Green, 1988) for the rest of its ascent while instantaneous melts are extracted and segregated at all levels and accumulated. It is envisaged that a steady-state system must exist, in which unmelted, "fertile" mantle enters the bottom of the column and melted, depleted mantle leaves the top of the column (Klein et al., 1991). The structure of the melting column dictates that the degree of melting should not be consistent in all parts of the column and that there should therefore be a variety of instantaneous melt compositions produced simultaneously. The centre of the column will produce a point-depth average composition representing the accumulated mixture of all the instantaneous melts (McKenzie & Bickle, 1988). Most recent studies, however, provide evidence for the segregation and extraction of instantaneous melts from the melting column (Elliot et al., 1991).

The composition of a basalt produced by partial melting is primarily governed by the interplay between five variables: 1) the type of melting process (i.e. accumulated or instantaneous); 2) the initial depth of melting (i.e. variable pressure); 3) the degree of partial melting (% PM); 4) the mantle P-T conditions; 5) the initial composition of the mantle source region. The mantle that constitutes the Icelandic plume is thought to be compositionally enriched in incompatible trace elements relative to N-MORB (e.g. Schilling, 1973; Hart *et al.*, 1973; Wood *et al.*, 1979; Meyer *et al.*, 1985) and is thought to have higher mantle potential temperature (PT°C) (McKenzie, 1984).

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Figure 5.24: Schematic representation of a dynamic melting column (after Elliot et al., 1991). Melting and melt segregation occur along the mantle flow-lines (shown as solid lines), but are illustrated only at specific points (shown as boxes), with arrows indicating the direction of focused segregation of the instantaneous melts. The degree of depletion within the melting column increases from the base upwards.

5.6.2.2) Major element oxide modelling

### Introduction

The effects of four of the variables (i.e. processes, depth, % PM and PT°C) that effect melt composition are investigated here in three models, schematically illustrated in Figure 5.25. These models use major element oxides from published primary melt compositions that have been derived in different ways (McKenzie & Bickle, 1988; Nui & Batiza, 1991). Models 1 and 2 use melts from Nui & Batiza (1991). These authors have calculated primary melt compositions produced from a depleted MORB mantle (DMM) source that undergoes partial melting from various initial depths. During melting, each increment of melt produced is accumulated and pooled. Melts produced from the deepest levels (i.e. tallest melting column) represents the largest degree of partial melting, whereas melt from shallower (i.e. shorter columns) represent lower degrees of partial melting. Model 1 consists of four melts that represent four columns of variable height. The melts represent variable depth and variable % PM. The four melts in model 2 also represent melts derived from variable starting depths, but the accumulation is stopped after 16% partial melting, so that these melts also represent variable initial depth, but constant % PM. Model 3 uses melts derived from a DMM source that has already undergone 23-24%

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Figure 5.25: Schematic illustration of three models of mantle melting. Briefly the models are as follows: 1) accumulated melts derived from DMM mantle at variable depths producing variable % PM; 2) accumulated melts derived from DMM by constant degree of partial melting (16%) at variable pressures; 3) instantaneous melts derived from DMM mantle (variable mantle PT°C) that has already had 23-24% partial melt extracted from it. Models 1 and 2 use melt compositions from Nui & Batiza (1991), and model 3 uses melts derived from the parameterisation of McKenzie & Bickle (1988). In each model, the stippled area indicates the portion of the melting column that undergoes melting. These melt compositions are tabulated in Appendix 6.

melting, using the parameterisation of McKenzie & Bickle (1988). An instantaneous melt is extracted from this considerably depleted mantle and erupted at the surface. This has been done for three different mantle PT°C (melts 1-3). All eleven melt compositions are tabulated in Appendix 6.

The effects of the type of melting process (i.e. accumulated or instantaneous) on the composition of the melts are revealed by comparing models 1 and 2 with 3; the effects of variable initial depth of melting are revealed by comparing models 1 and 2; the effects of the degree of partial melting is tested within model 1; and the effects of variable mantle PT°C is investigated within model 3. The basalts in the study area are fractionated by variable amounts, and so the compositions of all the melts in models 1-3 were input to TRACE3.FOR computer program that models crystallisation processes (Nielsen, 1988). The program was set to the same parameters as in Section 4.5.4 for simple closed-system fractional crystallisation. The results in the form of major element oxides and mineral compositions are compared to the observed compositions of feldspar and volcanic glass from the basalts from the Hengill to Ingolfsfjall region.

## Modelled feldspar compositions

In Figure 5.26, the An content of the calculated feldspar, for each model, is plotted against the Mg # of the melt with which they are in equilibrium (Nielsen, 1988). The first increment of fractional crystallisation at which feldspar joins the crystallising assemblage is indicated with a symbol, as is every tenth increment up to 70% thereafter. The increments are joined by dashed lines which are back-projected to the Mg # of the initial primary melt. The observed feldspar compositions from the study area are also plotted. They are differentiated into megacryst cores from petrological type 2 samples, tabular feldspar crystals from petrological type 3 basalts, and quench crystals from all basalt types.

The melts within model 1 produce the largest variation in feldspar compositions, whereas for models 2 and 3, the melts within each produce feldspar of a much more uniform composition. Within model 1, melt that represents the largest degree of melting from the greatest depths produces the most anorthitic feldspar (An 79), but these are slightly less anorthitic than the compositions of the least anorthitic feldspar megacryst cores (An >80) (Figure 5.26a). Model 2 melts produce low anorthitic plagioclase (An 75) that overlaps with some of the observed quench feldspar compositions at extreme degrees of fractional crystallisation (An 60) (Figure 5.26b). Such melts do not appear to have been extensively involved in the generation

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Figure 5.26: Results of modelling 70% fractional crystallisation from different initial primary melts from three different models (Figure 5.25, ). The modelled An content of the feldspar and the Mg # of the liquids are plotted in the form of trajectories. Each trajectory represents the path followed by 70% fractional crystallisation, and the symbols on each indicate every tenth increment, up to 70%. The dashed lines are extended back to the starting Mg # of the primary melt. The analysed feldspar compositions are also plotted on these diagrams, and are differentiated into megacrysts cores from petrological type 2, tabular feldspars from petrological type 3 and quench feldspars from all basalt types.

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of these basalts, according to the compositions of the feldspar phenocrysts. Model 3 produces feldspars that are extremely anorthitic in composition (An >80), and ones that overlap completely with the field defined by the feldspar megacryst cores analysed here. Lower mantle PT°C produce slightly higher An content in the feldspars, which is in agreement with the occurrence of more anorthitic megacrysts in normal segments of MOR (Table 4.1, p. 77) (e.g. Donaldson & Brown, 1977).

The observations from model 1 indicate that variably sized melting columns, where the melt is accumulated, are capable of producing varied feldspar compositions. Observations from model 2 indicates that the controlling parameter in model 1 is the percentage of partial melting and not the initial depth of melting. However, neither of models 1 and 2, using such accumulated melts, can produce the observed megacryst compositions. Only instantaneous melts from a very depleted mantle, as in model 3, are capable of producing An >80 and the observed megacryst compositions upon subsequent fractional crystallisation. However, the melts from such a depleted mantle source are incapable of precipitating feldspar with An <78 in a closed magmatic system.

The tabular feldspar from petrological type 3 basalt is variable in composition between samples, and there is also considerable variation within an individual sample, such as Húsmúli (Mg # 45 and An content between 72 and 85). The modelling in Section 4.5.4 illustrates that many of the tabular crystals in petrological type 3 basalt are out of equilibrium with their host, but by lesser and more varied amounts than the megacrysts. Húsmúli and other petrological type 3 samples require a history of crystallisation in which they are repeatedly exposed to liquids of diverse compositions, in terms of their potential An content; they must also have exposure to liquids that are capable of producing compositions of An 85. This may be achieved if instantaneous melts from a highly depleted DMM source and accumulated melts from "fertile" DMM (as depicted by the melt 1 in model 1 and the melts in model 3) are poorly mixed within a single lava shield. This is supported by the variation in the [La/Yb]<sub>n</sub> in a single lava shield (Section 5.6.1; Figures 5.22 & 5.23) as well as the observations from the modelling of phenocryst equilibrium (Section 4.5.4).

## Modelling liquid compositions

In Figure 5.27, FeO\* and Mg # are plotted for the residual fractionated liquids from models 1 to 3. The symbols on each trajectory indicate increments of 10% fractional crystallisation, as in Figure 5.26. Ten basalt glass compositions from petrological type 3 units are also plotted on this figure and they are from units GB4



Figure 5.27: Results of modelling 70% fractional crystallisation from different initial primary melts (models 1 to 3; Figure 5.25). The FeO\* and Mg # of the calculated liquids are plotted in the form of trajectories. Each trajectory represents the path followed by 70% fractional crystallisation, and the symbols on each indicate every tenth increment, up to 70%. Analysed volcanic glass compositions are also plotted on these diagrams. These are from the units with petrological type 3 basalt and include glass from GB1, GB4, and the lava shields BA3 and Ingolfsfjall.

and GB1, and the lava shields of Ingolfsfjall and BA3. Model 1 indicates that deeper melting to larger extents produces more iron-rich melts, relative to shallower, smaller degree melts. This signature in the FeO\* abundances remains in the liquid through closed-system fractional crystallisation. Model 2 implies that it is depth that controls the observed relationship in model 1 because the relationship observed in model 1 persists in model 2. For model 3, the mantle PT°C has only a subtle effect on the abundance of FeO\* in the compositions of the primary melts. However, this relationship becomes enhanced in the liquids with increased differentiation.

In model 1, the glass samples with high Mg # (i.e unit GB4 samples) lie on the shallower and smaller degree melt trend (10 kbar, 16% PM), whereas the lower Mg # glasses (i.e. Ingolfsfjall) lie on the deeper and larger-degree melt trend (16 kbars, 26% PM). Samples with intermediate Mg # and FeO\* contents occur between the two aforementioned trajectories (i.e. units GB1, BA3) indicating the continuity of melt compositions represented by the glass samples. In model 2, the relationship between these trends and the glass samples is essentially the same, but lesser degrees of partial melting at 16 kbars raises the FeO\* content of the melt slightly relative to model 1, so that the low Mg #, high-FeO\* glasses of Ingolfsfjall could be produced at slightly shallower depths by smaller % PM than in model 1. In model 3, the high Mg #, low-FeO\* glasses of unit GB4 lie on the calculated 1380°C trend, whereas the other units lie above these instantaneous melt trends.

## Conclusion from modelling major element oxide compositions

This modelling indicates that, to produce highly calcic feldspar compositions (An >80) the liquids have to be either the product of very large degrees of partial melting (>28%), or the source region has to be very depleted and melt instantaneously. Variations in the depth of melting alone cannot affect the overall composition of the feldspars. The more compositionally varied tabular feldspars are probably the result of poor mixing of liquids produced by variable melting processes.

For the major element compositions of the residual liquids, the modelling suggests that all the samples with lower FeO\*, at a given degree of differentiation, are the products of accumulated melts from shorter, shallower melting columns, and/or relatively small % PM from a "fertile" DMM source; the higher FeO\* melts are from accumulated melts of a taller, deeper melting column, and/or by larger % PM from a "fertile" DMM source. This is in accordance with finding from experimental studies (e.g. Falloon & Green, 1988) and other geochemical investigations (Elliot *et al.*, 1991). The modelling also indicates that the low FeO\* melts can be produced from

instantaneous melting of a depleted source at the top of the melting column (Figure 5.25c). This will be caused by the lower pressures, coupled with the effect from depletion of  $FeO^*$  in the source region from previous melting episodes.

## 5.6.2.3) Trace element modelling

The melt compositions modelled above do not contain trace element abundances. In this section, trace element compositions of variable mantle source regions are used from Kostopoulos & James (1992). These authors provide abundances of incompatible trace elements for Bulk Silicate Earth (BSE) and DMM sources, and parameterisation to produce melt compositions from adiabatic decompression melting at variable mantle solidus T°C. Each successive increment was pooled in the melting column (i.e. accumulated melts). To represent the hotter, compositionally-enriched Iceland plume, a mantle with a composition of BSE and a solidus temperature of 1400°C was investigated. The mantle of the normal MOR systems is represented by a DMM source at 1300°C. These two mantle types represent the end members of the modelling, and for comparative purposes, a DMM source was melted at 1400°C.

In Figure 5.28, the incompatible trace element ratios Zr/Nb and Nb/Ti are plotted for the three calculated melts in the form of trajectories. The composition of the first increment of melt and subsequent 1% increments are indicated on these trajectories with symbols labelled accordingly. The petrological type 3 basaltic units from the study area and unit S1 (sample A416) of petrological type 2 basalt are also plotted. In this figure, the modelled partial melting trajectories all run parallel to each other. The melts from the hotter mantles (1400°C) have very similar incompatible element ratios, while the melt from the cooler mantle (1300°C) has higher Nb/Ti ratios for a given Zr/Nb ratio. However, when the degrees of melting along each trajectory are examined, the DMM melts have equal Zr/Nb ratios at a given % PM, while the BSE melts are the products of larger % PM (e.g when Zr/Nb=11). These observations indicate that melt can be produced from source regions with different trace element compositions and still have the same incompatible element ratios, providing the solidus temperature is the same, but these melts will not be the products of the same % PM.

The petrological type 3 basalts and unit S1 sample plotted in Figure 5.28 coincide with the BSE and DMM trajectories of 1400°C, which is in accordance with their location on the Iceland plateau where the mantle is hotter. The spread of samples

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along these trajectories indicates that they may not be the result of identical melting conditions, and supports the major element modelling (Section 5.6.2.2). They could either be the result of a single source composition and variable degrees of melting, or visa versa. The modelling implies that most samples are the result of approximately 1.5-3.5% PM of BSE, or even smaller melting of DMM at 1400°C. This is contradictory to the conclusions drawn from the major element modelling above where most of the glass samples follow trajectories representing large % PM (Section 5.6.2.2; Figure 5.27).



Figure 5.28: Results of modelling partial melting of BSE and DMM at variable solidus temperatures (parameterisation from Kostopoulos & James, 1992). The incompatible trace element ratios of Nb/Ti and Zr/Nb are used to illustrate the degree of enrichment of these trace elements in the melts. Every 1% increment of % PM is indicated and labelled accordingly. Samples of petrological type 3 basalt, and unit S1, of petrological type 2, are plotted on this diagram. Unit GB4, discussed in the text is labelled accordingly.

# 5.6.2.4) Synthesis of modelling results

In this section, examples of eruptive units are examined in terms of petrogenesis in light of the information derived from the modelling above. The units considered include the Ingolfsfjall lava shield (ILS), and the eruptive units of GB4 and S1. Firstly, Ingolfsfjall is compared with unit GB4, and subsequently unit GB4 is compared with unit S1.

	Unit	GB4	Ingolfsfjall lava shield	
	Observations	Possible origins	Observations	Possible origins
Feldspar	An 81-85	Instantaneous melting	An 71-83	1) Instantaneous
		from a highly depleted	-	melting from a highly
		DMM source		depleted DMM source
				2) Accumulated melts
				from <i>very</i> large % PM
				(>28%)
Liquids	Mg # 65	1) Accumulated melts	Mg # 50	1) Accumulated melts
	FeO* 8.7 wt. %	from 10 kbars, 16%	FeO* 14 wt. %	from 16 kbars, 26%
		РМ		РМ
		2) Instantaneous		2) Accumulated meits
		melting from a highly		from 12 kbars, 16%
		depleted DMM source		РМ
		at 1380°C (PT°C)		
Trace element	High Zr/Nb (i.e.	1) BSE 6% PM	Low Zr/Nb (i.e.	1) BSE 2% PM
ratios	depleted liquids)	2) DMM 3% PM	enriched liquids)	2) DMM 1% PM

**Table 5.3:** Summary of the observations for unit GB4 and the Ingolfsfjall lava shield (feldspar, liquid and trace element ratios), and the possible causes of these, as inferred by the modelling in this chapter.

# Ingolfsfjall and unit GB4

In Table 5.3, the observed feldspar compositions, FeO\* and Mg # of the glasses and the incompatible trace element ratios for these eruptive units are summarised. The possible mechanisms that result in the observed compositions, as defined by the modelling above, are also listed. The composition of feldspar analysed from within these units are similar and high in An content (intratelluric crystals have An >81). The plagioclase crystals in the Ingolfsfjall lava shield are usually zoned microphenocryst that are not in equilibrium with their host (Section 4.5), whereas those in unit GB4 are quench crystals in a glass sample that *are* in equilibrium. The

composition of the feldspars in these units suggests an influence of instantaneous melts from a highly depleted source.

For the liquid compositions, the modelling suggests that the low-FeO\* signature of unit GB4 is either the result of shallow and moderate degrees of melting, or instantaneous melting of a depleted mantle (1380 PT°C). In view of the quench feldspar composition (Table 5.3), the unit is probably results from the latter process. Ingolfsfjall has a high-FeO\* signature that could have been produced by either melting to just <26% from 16 kbars (Figure 5.27a), or by shallower melting by lesser degrees (Figure 5.27b). This contradicts the origin inferred from the intratelluric microphenocrysts of feldspars (An >81). Mixing of depleted instantaneous melts with accumulated melt prior to ascent of the lavas is a mechanism that would be able to produce the observed relationship in the Ingolfsfjall lava shield. If this is the case, the signature of the depleted melts is only preserved in the feldspar compositions, and any effect on the bulk-rock composition is diluted.

The interpretation given above for the origin of these two eruptive units is supported by the observed relationship between FeO<sup>\*</sup> and the  $[La/Yb]_n$  ratio in Figure 5.23. Unit GB4 has lower FeO<sup>\*</sup> abundances and a low  $[La/Yb]_n$  ratio relative to the Ingolfsfjall lava shield (Section 5.6.1). It is concluded therefore, that unit GB4 is the result of instantaneous melting of a very depleted DMM source, and that Ingolfsfjall is the result of accumulated melting from a tall column that was probably poorly-mixed with instantaneous melts from the top of the column just prior to eruption; the signature of which is preserved only in the composition of the feldspars.

The observations in Figure 5.28 suggest that both these units are the result of melting from a mantle with a solidus temperature of 1400°C. However, the relative difference in the % PM between them inferred from the trace element modelling is contradictory to the conclusions drawn above. In Figure 5.28, Ingolfsfjall appears to be the products of smaller % PM than unit GB4. This can be explained, however, if the initial source of GB4 is more depleted; as implied by its feldspar and major element oxide compositions. The absolute abundances of the Nb/Ti and Zr/Nb ratios predicted for the Icelandic samples suggests that the source for the Icelandic basalts must have *extremely* enriched incompatible element ratios, and more than those in the BSE mantle model presented here (Kostopoulos & James, 1992).

## Units GB4 and S1

It was concluded above that the GB4 unit was the product of instantaneous melting from a depleted DMM source. In Figures 5.22a, 5.23 and 5.28, both units

GB4 and S1 are similar in bulk composition. However, they belong to different petrological types. Unit S1 is of petrological type 2 basalt with only plagioclase megacrysts, and unit GB4 is of petrological type 3 basalt with olivine and plagioclase microphenocrysts (Table 4.3, p. 87). Unit S1 is the only petrological type 2 sample in which the feldspar megacrysts are *not* drastically out of equilibrium with the host; as assessed by the composition of the crystal rims (Section 4.2.1.1, p. 75). Unit GB4 also contains equilibrium calcic feldspar in the form of quench crystals (Figure 5.26b). Both these eruptive units have REE patterns that are more LREE-depleted, relative to chondrite, than the rest of the suite analysed from the region. These REE profiles and the relationship observed in Figure 5.23 for FeO\* and [La/Yb]<sub>n</sub> strongly suggest that the eruptive units are the result of similar melting processes (i.e. instantaneous melts from a depleted DMM source).

The question rises as to why unit GB4 basalt is devoid of megacrystic feldspar and saturated in olivine, while unit S1 basalt crystallised megacrysts in abundance and lacks olivine. The occurrence of the feldspar megacrysts in the vicinity of the Grensdalur Central Volcano suggests that this reservoir is in some way the cause of the generation of these crystals. It has been suggested that megacrystic plagioclase is produced from supercooling of melts when they encounter melts of drastically different composition and temperature (Kuo & Kirkpatrick, 1982). This interpretation is supported by the evidence presented in this thesis. The abundance of these crystals in the porphyritic type 2 basalts, and in gabbroic xenoliths, implies that magmas similar to units GB4 and S1 should have been a commonly produced magma type in the region, and yet only a small fraction of the eruptive units in the region have such compositions (including the Mælifell eruption studied by Hardardóttir (1986), Trønnes (1990) and Hansteen (1991)). This suggests that most of these magmas are being interrupted en route the surface by a crustal reservoir and being supercooled, mixed and diluted prior to eruption; however, occasionally, mixing does not occur and the magma reaches the surface (e.g units such as Mælifell, S1 and GB4). When this occurs, it is suggested that either the magma reservoir has become clogged with crystal mush of this sort, and so the magma is essentially isolated (for units S1 and Mælifell), or that the magma passes straight to the surface (for unit GB4).

# 5.7) Summary

There is scatter amongst the whole-rock compositions, but coherent trends exist for the volcanic glass data. The chondrite-normalised multi-element profiles illustrate that the majority of the samples have enriched incompatible element ratios,

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relative to chondrite, but that there is significant variation in the degree of enrichment in the suite of samples. All the samples are tholeiitic, and the majority are classified as olivine tholeiites, despite their petrographic character.

The aphyric petrological types 1a & 1b samples (whole-rock and glass) have a uniform composition, whereas petrological type 2 (whole-rock) samples, with significant feldspar megacrysts, are scattered and fall along a modelled trajectory representing feldspar megacrysts addition, suggesting they represent aphyric samples that have had xenocrystic calcic plagioclase added to them. There is as much scatter observed for the petrological types 3 whole-rock samples that cannot be explained by crystal accumulation. The glass compositions of the two porphyritic suites, petrological types 2 and 3, define parallel liquid lines-of-descent, suggesting that they are *not* petrogenetically related to each other by fractional crystallisation. However, this is not directly supported by the normative compositions. Amongst the various petrological types of basalt there is no consistent relationship between the degree of enrichment in the incompatible element ratios and the petrological type of basalt.

When the samples are examined in terms of regional and stratigraphic distribution, a pattern emerges where the composition changes from being monotonous to varied, and these periods are dominated by different petrological types of basalt. The extent of, and variation in, differentiation becomes more enhanced in the younger units of the Hverageroi Volcanic System, which occur in the vicinity of the extinct Grensdalur Central Volcano. The lava shields are generally less fractionated, but each unique in elemental abundances at a given degree of differentiation. Modelling and examples of vertical profiles through the shields suggests that they are not the products of closed-system magmatism. However, the constraints provided by the mineralogy (Section 4.5) indicate that open-system magmatism, involving mixing of magmas of diversely fractionated compositions, is also not a viable mechanism. The lava profiles of the lava shields indicate that the mixing must be between lavas with subtle differences in the degree of differentiation, as well as the incompatible element abundances and ratios. The main Hengill eruptive unit is identical to Ingolfsfjall in composition, and therefore is believed to represent a subglacial lava shield type of eruption, and not the Hengill Central Volcano.

The compositional variation within the stratigraphy of the region provides some constraint on the nature of the magmatic plumbing system there. The occurrence of successive eruptions of uniform basalt may indicate thorough mixing in the reservoir, while the progressive increase in contamination by megacrysts in a given region, coupled with the more abundant occurrence of extensively fractionated eruptions, indicates that the magma reservoir is "clogging up" with crystal mush and exists as isolated pockets of magma. Alternatively, the magma of a single eruption event (lava shield) arrives at the surface having undergone no detectable crystal addition or mixing with extensively fractionated magmas. The compositional variation within the lava shield eruptions does not fit the classic models of either open- or closed-system magmatism, and it is suggested that the lavas are erupted straight to the surface without encountering a crustal reservoir of any sort. The magmas may pond and accumulate at the base of the crust before eruption begins, but it is envisaged that the eruption is then fed with melts that come straight off the mantle melting column. Húsmúli is the only exception to this type of model, and may require a magma reservoir.

The REE signatures of the basalts are not consistent with the type of volcanic eruption event from which the magma was erupted (i.e. lava shield or fissure). However, the basalts from the lava shields have a distinct relationship between FeO\* and the degree of enrichment in the incompatible element ratios; as do the other samples from the Hveragerði Volcanic System, once the effects of fractional crystallisation and megacryst addition have been accommodated for. Samples with high FeO\* are more enriched in their incompatible element ratios, such as [La/Yb]n.

Modelling indicates that the feldspar compositions are primarily controlled by the % PM and the level of depletion in the source prior to melting. The highly calcic plagioclases (megacrysts and some tabular feldspars) originated from instantaneous melting of a depleted DMM source. The more compositionally varied tabular feldspars are probably the result of poor mixing of liquids produced in variable ways. For the major element oxide compositions, modelling suggests that all the samples with lower FeO\* are products of shallower melting. Basalts with identical trace element ratios can be produced in more than one way by varying either the %PM, source composition or the mantle solidus T°C. The samples from the study area coincide with the 1400°C mantle solidus T°C trajectory, but could also be produced by from a variably depleted source.

The genesis of basalts from the research area are constrained by the modelling, and are likely to be the products of highly variable melting processes from a mantle that must be extremely enriched in incompatible element ratios. A model similar to that of Elliot *et al.* (1991) is applicable, with instantaneous melts from a depleted mantle being effectively extracted and erupted. Instantaneous melt form

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depleted mantle may either erupt as picritic lava shields, or enter the crustal reservoir where they are supercooled, producing megacrysts (Kuo & Kirkpatrick, 1982).

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# Chapter 6

# Oceanography

# **6.1) Introduction**

During the period of this research the author participated in two research cruises to the Reykjanes Ridge. The first cruise was on the R/V Maurice Ewing in October 1990 (cruise EW9008) and the second was on the Bjarni Sæmundsson in November 1990 (cruise B16-90) (Figure 6.1). Data acquired during these cruises include bathymetric charts and profiles, sidescan sonar images and petrographic observations and geochemical data from dredged rocks. In this chapter, the data from each of these cruises are described in turn and each is compared with corresponding data from Iceland.

The Reykjanes Ridge is an obliquely spreading MOR in the North Atlantic Ocean. Bathymetric mapping of the ridge prior to the work described here had yielded a 1:2,400,000 map of the Northeast Atlantic (Laughton *et al.*, 1982). A more detailed map of the area north of 63°N, as far north as Iceland, exists (Johnson & Jakobsson, 1985; Ólafsson *et al.*, 1991) and there are many single line profiles across the ridge (Vogt, 1971; Shih *et al.*, 1978; Jacoby, 1980; Johansen *et al.*, 1984; Johnson & Jakobsson, 1985). Collectively the existing bathymetric data indicate that the ridge shallows towards Iceland. Additionally, the ridge profile perpendicular to the axis gradually changes from that typical of a slow spreading ridge in the south with a well developed median valley, to one that mimics a fast-spreading ridge with no median valley in the north towards Iceland.

Existing sidescan sonar images from GLORIA II (Searle & Laughton, 1981) reveal two main tectonic lineaments. One is parallel to the trend of the ridge axis (036°), and the other oblique to it and perpendicular to the spreading direction (099°). The ridge-parallel lineaments occur outside the axial zone and persist for variable distances off-axis. They are interpreted as faults (Searle & Laughton, 1981). The oblique lineaments orientated at approximately 014° are restricted to the median

valley and are interpreted as en echelon volcanic ridges (Searle & Laughton, 1981). The orientation of these features north of 63°N is west of 020° (Johnson & Jakobsson, 1985).



Figure 6.1: Bathymetric map of the Reykjanes Ridge showing the locations of the two research cruises. Areas A-C refer to the three survey areas of the EW9008 cruise and B16-90 refers to the location of that cruise. The en echelon volcanic systems on the Reykjanes Peninsula and Hengill are also indicated (Jakobsson et al., 1978). The location of the DSDP Leg 49, Hole 409 is also indicated with filled circles (Luyendyk, Cann et al., 1978); and the locations of the earthquake swarms of May 1989 (Nishimura et al., 1989), September 1990 (RIDGE Events, 1990) and the November 1990 (Ólafson et al., 1991) are also indicated.

# 6.2) EW9008 cruise - R/V Maurice Ewing

# 6.2.1) Cruise objectives

The cruise took place during October 1990 departing from Bergen (Norway) and arriving in New York (U.S.A.) one month later. The two main objectives of the cruise were to study the tectonics of an obliquely spreading MOR, and to examine the tectonics at varying distances from the Icelandic plume. Three survey areas were chosen at various localities between Iceland and the Bight Fracture Zone to the south. These are referred to as areas A, B and C, from north to south, and are equally spaced relative to each other (Figure 6.1).

The three survey areas coincided with segments of the ridge that had recently been investigated by other workers who found features of specific interest. In area A  $(61^{\circ}30'N-62^{\circ}03'N)$  submersible dives had yielded detailed accounts of the morphology (Holcomb *et al.*, unpublished manuscript, 1988). Area B ( $60^{\circ}35'N-59^{\circ}40'N$ ) includes the site of the May 1989 earthquake swarm (Nishimura *et al.*, 1989; Vogt, *et al.*, 1990) and, prior to the EW9008 cruise, the *RV Maurice Ewing* surveyed this region using both multibeam echo-sounders and the 'SeaMarc II' sonar device (cruise title EW9007) (Appelgate & Shor, 1991; Bell & Buck, 1992). In area C ( $57^{\circ}42'N-58^{\circ}28'N$ ) active hydrothermal sites had been reported from Soviet submersible dives (Kuznetsov *et al.*, 1985).

## **6.2.2)** Instrumentation

The R/V Maurice Ewing is an American scientific research vessel from the Lamont-Doherty Geological Observatory, Columbia University. It is equipped with geophysical instrumentation which includes a swath-bathymetry multibeam echosounder (the Krupp-Atlas "Hydrosweep"), single beam echo-sounders (12 and 3.5 kHz), GPS navigation, a gravimeter and a magnetometer. The Towed Ocean Bottom Instrument (TOBI) was brought aboard from the Institute of Oceanographic Sciences Deacon Laboratory (IOSDL). This is equipped with a double-sided 30 kHz sidescan sonar device, sub-bottom profiler (7.5 kHz), tri-axial magnetometer, 'transmissometer and a thermistor. The echo-sounders, gravimeter and magnetometer were in operation from port to port, whereas TOBI was deployed at selected locations (within areas B and C; Figure 6.1). This was the first time TOBI had been deployed over a MOR.

The sonar devices were the primary tools used during this survey. The Hydrosweep multibeam echo-sounder was used to produce a bathymetric map. It has

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Figure 6.2: a) Example of Hydrosweep bathymetry data. The contour intervals are 50 m (absolute range = 750-1300 m below sea-level). This example is from area B. It includes the location of the May 1989 earthquake swarm (Nishimura et al., 1989), half of profile B from Figure 6.4, and is of the same area as Figure 6.2b. X and Y are seamounts and are indicated in Figure 6.2a & b and can be used for a references of the relative location within the swaths from these two types of data.

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b)



Figure 6.2: b) Example of the TOBI deeptow, high resolution, sidescan sonar data. The TOBI track is in the centre of the swath and is indicated in Figure 6.2a. Areas of higher reflectivity are lighter in shade (labelled accordingly). This example is the same area as Figure 6.2a. The swath width is approximately 6 km.

a swath width equivalent to twice the water depth and an example of the data is given in Figure 6.2a. The contour intervals are 50 m in this example with ticks pointing towards deeper contours. The entire data set from each of the three survey areas was gridded and perspective views generated, examples of which are displayed in Figure 6.3.

The deeptow, high resolution, sidescan sonar device, TOBI was towed approximately 400 m above the sea-floor. It produces an image with a swath width of approximately 6 km. An example of the data is given in Figure 6.2b. The black band down the middle of the swath marks the TOBI track. The sidescan sonar emits a signal and measures the acoustic backscatter from the sea-floor. Generally, the degree of backscatter is proportional to the roughness of the surface. Young fresh lavas tend to be rough and have a high reflectivity, and as the surface of the lava is progressively covered with sediment, it is smoothed and reflectivity is diminished. This variation in reflectivity can be seen in Figure 6.2b with the more reflective material appearing as a lighter shade of grey. Shadows are created by the undulating topography on either side of the TOBI track where the signal does not penetrate. The region in Figure 6.2a & b is the same, and is from area B (Figures 6.1 & 6.3). The height of the features observed on the TOBI image (Figure 6.2b) can be measured using the Hydrosweep data (Figures 6.2a). For example, seamounts X and Y are 150 m and 200 m in relief, respectively.

# 6.2.3) Observations from the EW9008 cruise

An overview of the observations from the cruise are described and discussed in Murton *et al.* (1990) and Parson *et al.* (submitted to EPSL) and only a summary of the results is presented here. Emphasis is placed on results relevant to the Icelandic research discussed in Chapters 1-5 of this thesis. Results from the cruise confirm that there is variation in the bathymetric profile and shallowing of the ridge towards Iceland. Profiles across the ridge in areas A to C differ in absolute depth and the degree of elevation of the axial zone relative to the flanks (Figures 6.3 & 6.4). Area C in the south displays a valley that is approximately 10 km wide, 400 m deep and asymmetric in profile. This shallow median valley becomes progressively reduced in magnitude northwards until at 59°N, midway between areas C and B, the profile gradually changes and mimics that of a typical fast-spreading ridge (Figures 6.3 & 6.4). Oceanography





Figure 6.3: Hydrosweep data gridded and generated into perspective views. An example from each of the three survey areas is given in a, b and c. The location of the profiles in Figure 6.4 is displayed (dashed lines). The colour scale in each picture is in meters below sea-level.

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Figure 6.3: continued.

The ridge-parallel faulting evident in the GLORIA II data (Searle & Laughton, 1981) was imaged here in more detail by TOBI and can be quantitatively measured from Hydrosweep bathymetric charts (Spencer *et al.*, 1991). The throw on these faults is in the region of 500 m in area C where they form the median valley walls (Figure 6.4c) but this gradually diminishes to <100 m in area A where there is no median valley (Laughton *et al.*, 1982; Parson *et al.*, submitted to EPSL).

The profiles in Figure 6.4 each have an elevated zone in the vicinity of the axis, although in area C, it is less obvious due to the presence of the median valley. These elevated features are axial volcanic ridges (AVRs) and because of their en echelon nature on the Reykjanes Ridge are defined as en echelon volcanic ridges (EVRs) (Murton *et al.*, 1990; Parson *et al.*, submitted to EPSL). The elevated regions in Figure 6.4 have a maximum height of 400 m and are 8-9 km wide. In areas A and C, the elevated feature is split into two unequal halves. It can be seen form Figure 6.3a that this divide represents the gap between adjacent overlapping EVRs. The same is observed in area C (Figure 6.3c & 6.4), but profile B-B' crosses only a single EVR (Figures 6.2-6.4). The EVRs persist in all three survey areas (Figure 6.3) and average 20 km long and 5-6 km wide. The constructional nature of these ridges is revealed here in the detail of Hydrosweep and TOBI data (Figure 6.2).



Figure 6.4: Profiles A-A', B-B' & C-C', from areas A, B & C respectively. The locations of the profiles are indicated in Figure 6.3, and profile B is also indicated in Figure 6.2. The location of the axis, median valley and the EVRs are indicated.
A typical EVR is shown in Figure 6.5. They are each built up of a combination of the following features: i) curvi-linear scarps, interpreted as faults; ii) conical mountains (aspect ratios <2) described as seamounts; iii) hummocky terrain interpreted as pillow lava flows; and iv) flat lying sheet flows which are detected by their smooth texture, but only when there is a contrast in the acoustic reflectivity between the flow and the underlying material (Figure 6.5). The interpretation presented in Figure 6.5b is based on the collective observations from both Hydrosweep and TOBI sonar data, with the exception of the sheet flows which are detectable only by TOBI.

a)



Figure 6.5: An example of a typical EVR: a) TOBI image; b) line drawing and interpretation of a). This example is the southern-most EVR in area C. The main constructional features and the faulting are labelled accordingly.

EVRs appear to be made up of coalesced conical seamounts and hummocky terrain. The latter often has a linear fabric (Figure 6.5). The relative proportion of seamounts, pillow lavas and sheet flows varies within and between EVRs. The degree of faulting is also highly variable. There appears to be an interplay between the tectonic features (faulting) and the robustness of the various volcanic features (seamounts, hummocky terrain and sheet flows). Some EVRs exist as highly reflective regions with well preserved volcanic morphologies, whereas others have a much lower reflectivity and volcanic morphologies cut by faults. In view of this, the EVRs are said to be in different tectonic states. There is no consistent relationship between the tectonic state of the EVR and the distance from the Iceland plume. Adjacent EVRs often exist in different tectonic states, as shown in Figure 6.2b by EVR 1 and EVR 2.

## 6.2.4) Geomorphological comparison with Iceland

## 6.2.4.1) Icelandic geomorphology

There are many striking similarities between the sea-floor observations described above (Section 6.2.2) and the neovolcanic zone in Iceland. Attention has been drawn to some of these in the past, such as the en echelon segmentation (Thorarinsson, 1968; Jakobsson, 1974; Johnson & Jakobsson, 1985). However, with the increased resolution of the data acquired from the EW9008 cruise (and EW9007 cruise) this comparison can be taken further. Below, the bathymetric and sidescan sonar data from the Reykjanes Ridge is compared with topographic maps and aerial photography from Iceland.

The Reykjanes Peninsula is an onland continuation of the oblique spreading Reykjanes Ridge, and in this sense it is the most suitable part of the Icelandic neovolcanic zone for a comparison. Additionally, the volcanic systems lack central volcanoes (Chapter 1) and are arranged in a similar en echelon fashion (Jakobsson *et al.*, 1978) (Figure 6.1). The volcanic products have been produced by both subglacial and subaerial volcanism and, in a comparison of Icelandic topography with sea-floor bathymetry, the products of subglacial eruptions are likely to provide the most suitable analogue because they are also subaqueous in origin (Thorarinsson, 1968; Searle, 1983).

On the Reykjanes Ridge, the EVRs are defined by the break in slope that surrounds each one (Figures 6.3 & 6.4). The subglacial mountains of the Reykjanes Peninsula are treated accordingly and divided into volcanic systems. On the entire Reykjanes Peninsula this produces six volcanic systems (including the Hengill Volcanic System) which differs slightly from other interpretations (Jakobsson *et al.*, 1978; Einarsson & Sæmundsson, 1985) (Figure 6.1). Three of these volcanic systems, as defined here, are displayed in detail in Figures 6.6 & 6.7.



Figure 6.6: A topographic map of part of the Reykjanes Peninsula, SW Iceland. The contour interval is 100 m. The 200 m contour is in bold, in order to approximately represents the extent of the three volcanic systems defined here which are separated by areas of flat ground. The location of the region displayed in Figure 6.7 is indicated.

Each subglacially produced volcanic system is composed of the following features: i) scarps, interpreted as faults; ii) conical mountains, described as table mountains; iii) undulating terrain, usually with a linear fabric (moberg ridges). These geomorphologies are observed on topographic maps, aerial photographs and satellite imagery (Figures 1.4, 1.5 & 6.6-6.8). The conical table mountains are mostly lava shields, and the undulating terrain with a linear fabric is usually the product of fissure-fed eruptions (Chapter 1; Figures 1.3 & 1.4).

Adjacent volcanic systems on the Reykjanes Peninsula have varying degrees of faulting and other lineaments (Figures 6.6-6.8). The westernmost Fagradalsfjall Volcanic System has a smoother outline and surface, whereas the Núpshlíðar and Krísuvík volcanic systems have more faulting and are fissure dominated (Jakobsson *et al.*, 1978). Using the same sea-floor terminology, the Icelandic volcanic systems can also be said to be in different tectonic states (Figure 6.9). The maximum relief of these subglacial volcanic systems in Figure 6.7 is 350 m (Fagradalsfjall).

6.2.4.2) Geomorphological similarities between Iceland and the Reykjanes Ridge

The similarities between the Reykjanes Ridge and the Reykjanes Peninsula are many. Both are oblique spreading zones which results in segmentation of the neovolcanic zone into an en echelon arrangement. These segments are described as EVRs on the Reykjanes Ridge (Murton *et al.*, 1990; Parson *et al.*, submitted to EPSL) and volcanic systems on the Reykjanes Peninsula in Iceland (e.g. Jakobsson *et al.*, 1978). In both areas they are approximately of equal dimensions ( $5\times30$  km). Both the EVRs and the volcanic systems show much variation in the relative proportions of the various volcanic features and the degree of faulting, and are observed to be in different tectonic states.

There are also differences between the two areas. These emerge when details of the Icelandic morphology and geology are examined. The differences are primarily a consequence of the variable eruptive environments in Iceland (Chapters 1 & 2), the subsequent weathering that modifies the original morphology, and the Icelandic plume tectonic environment. These differences are outlined briefly below. A comparison between subglacial and submarine morphology may be acceptable if these factors are bourn in mind.

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a)



Figure 6.7: a) (left) Aerial photograph taken over part of the area in Figure 6.6. The morphologically distinct features are labelled accordingly; b) (above) sketch of the photograph in a) indicating the extent of hyaloclastite in the region and the Postglacial volcanic vents (geological information from Sæmundsson & Einarsson, 1980).

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Figure 6.8: Photograph of the Reykjanes Peninsula looking to the NW. The three volcanic systems featured in Figures 6.6 & 6.7 are visible. They are observed to be in different tectonic states. The lake is situated in the right hand-side of the photograph is Kleifarvatn.

The difference between subglacial and submarine eruptive environments is poorly understood as neither have been directly observed. The veneer of hyaloclastite debris produced during a subglacial eruption (Section 2.2.1.1) is not observed offshore in the same abundance (e.g. Lonsdale & Batiza, 1980; Holcomb et al., unpublished manuscript, 1988). Debris of this kind would smooth over hummocky topography of pillow lavas, as it does in the hyaloclastite edifices on Iceland (Figure 2.2; p. 20). The presence of such hyaloclastite debris on Iceland could therefore explain the difference in the textures observed between TOBI data of the submarine EVRs (Figures 6.2b & 6.5) and aerial photographs of the subglacial hills (Figure 6.7a). In addition, many subglacial eruptions breach the ice water level to produce subaerial lavas (Section 2.2.1.3) which will modify the morphology. It may be the case that the steep-sided, flat-topped, subglacial table mountains in Iceland acquire their characteristic shape from the interaction of the eruptive products with the confining walls of the ice and the capping subaerial lavas, and the similarity in shape to submarine seamounts may only be a coincidence. Alternatively, the shapes may be the result of a crustal generation process that operated in Iceland and the Reykjanes Ridge. The shape of the subglacial table mountains (subglacial lava shields) in Iceland has been attributed to the result of lower effusion rates of the eruptive material, which in turn is thought to indicate that the lavas originate from deeper in the crust than the fissure eruptions (G. P. L. Walker, 1971; Sæmundsson, 1979) (Chapter 1; p. 5-8).

The eruptive environment within Iceland has alternated between subaerial and subglacial in the last 3.1 Ma, and it has been entirely subaerial for the last 10,000-12,000 years in most parts of the island (Chapter 1). This means that a comparison of the subglacially erupted products will not be representative of a continuous period of volcanic activity, and may not represent activity at the present locus of spreading. This is apparent when the distribution of the Postglacial volcanic vents sites is examined (Figure 6.7b). Additionally, the outlines of volcanic systems on the Reykjanes Peninsula (Figures 6.6 & 6.7) are probably greatly affected and reduced in size by the burial effect of the laterally extensive Postglacial subaerial lavas. Furthermore, weathering modifies the landscape considerably on Iceland, and subglacial edifices that have suffered extensive weathering are poor candidates for a morphological comparison.

The other main factor to consider is the effect of the Icelandic plume on the volcanism of the island. The volume of magma produced in Iceland is estimated to be considerably greater than that of a normal MOR (McKenzie, 1984). This effect is seen further inland towards the centre of Iceland where the neovolcanic zone becomes considerably wider and the volume of \_\_\_\_\_\_ individual eruptions becomes much larger. This is true for lava shield and fissure type of eruptions (Figure 1.3) (e.g. Skjaldbreiður and Laki respectively).

The main conclusion from this qualitative morphological comparison is that both EVRs and volcanic systems are composed of many individual eruption events, and in Iceland, these may be divided into two end member types: lava shields and fissure eruptions. It is suggested here that the conical mountains observed in TOBI and Hydrosweep data may be equivalent to the lava shields, and the hummocky terrain, interpreted as pillow lavas, originates from fissure eruptions. Chapter 5 describes how the two types of volcanic eruption event in the Icelandic study area (lava shield and fissure) erupt different petrological types of basalt, the differences of which are thought to be the result of different routes through the crustal magmatic plumbing system to the surface (Figure 5.21, p. 142). The same petrographic types have been reported from the MAR in the past (e.g. Aumento, 1968; Langmuir *et al.*, 1977; Donaldson & Brown, 1977; Wood *et al.*, 1978). It remains to be seen whether a correlation exists on the submarine spreading centres between petrographic basalt types and submarine volcanic morphology; i.e. seamounts and elongate outcrops of hummocky pillow lavas.

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## 6.3) B16-90 cruise - Bjarni Sæmundsson

## 6.3.1) Introduction and cruise objectives

On 30<sup>th</sup> October 1990 an earthquake swarm was reported near 63°N on the Reykjanes Ridge (Pall Einarsson of the Science Institute, Reykjavik). A rapid response cruise was organized (Professor J. R. Cann, University of Leeds) with the cooperation of Icelandic scientists, the RIDGE community and UK institutions (University of Durham and IOSDL). The *Bjarni Sæmundsson* research vessel was made available and the cruise was led by Jon Ólafsson from the Marine Research Institute (Reykjavik). The vessel was equipped with several single beam echosounders of various frequencies, dredging equipment, a rosette water sampler with CTD (conductivity temperature depth) and GPS navigation. Additional equipment supplied for the investigation included 12 sonobuoys (Keflavik US airbase), receivers and recorders (University of Leeds) and the WASP underwater camera (IOSDL).

The primary objective of the cruise was to investigate the site of the earthquake swarm, and see if it was associated with a volcanic eruption event. If an eruption site was located, the objectives were to sample fresh volcanic rock and gases, and to observe the hydrothermal activity associated with an eruption. Attempts to detect and pin-point the eruption, in the absence of accurate earthquake locations, were made by sampling the water column to detect instabilities and chemical anomalies there. A zigzag grid of CTD stations crossing the ridge axis was designed to locate the eruption site and dredges were taken from various sites along this section (Figure 6.9). The process of event detection is described below, followed by an account of initial petrographic and geochemical observations made on the dredged samples. The petrology of the dredged basalts from this segment of the Reykjanes Ridge is then compared to that of the Hengill-Ingolfsfjall region on Iceland, in light of the petrogenetic model developed in Chapters 4 & 5.

## 6.3.2) Observations from the B16-90 cruise

## 6.3.2.1) Bathymetry

The bathymetric maps available at the time of the B16-90 cruise were limited. A combination of fishery charts, published diagrams (Johnson & Jakobsson, 1985) and the Hydrosweep coverage from the R/V Maurice Ewing cruises (EW9008 and EW9007) were used as guide-lines to the underlying bathymetry. Subsequently, in

April 1991, the area was mapped in more detail, and this map is used in Figure 6.9 (Ólafsson *et al.*, 1991). In addition, several bathymetric profiles were recorded during the cruise and these are presented in Figure 6.10 with their locations marked on Figure 6.9. They coincide with some of the dredge sites.



Figure 6.9: Bathymetric map of the Reykjanes Ridge (Ólafsson et al., 1991). Contour interval=30 meters. The location of the B16-90 CTD stations and dredge sites are indicated. Five EVRs present in the region were investigated, and are labelled 1-5 accordingly. EVR2 has been named Steinaholl (Ólafsson et al., 1991). The location of the single beam echo-sounder line profiles of Figures 6.10 are also indicated.

From the more detailed bathymetry map obtained subsequent to the cruise, several EVRs can be defined by the topographic lows between them. These are referred to as EVR 1 to 5 from south to north, respectively (Figure 6.9). They are approximately 10 km in length which is considerably shorter than those surveyed further south by the EW9008 cruise. Along the length of the ridge, there are lateral offset. These occur between EVRs 3 and 4, and similarly between EVR 1 and the area to the south, and between EVR 5 and the area to the north. The bathymetric low separating EVRs 4 and 5 is poorly defined. The offsets occur at 20 km intervals on the ridge. This scale of segmentation is similar to the EVR-segmentation further south, and the lengths of the volcanic systems on the Reykjanes Peninsula (Jakobsson *et al.*, 1978). The bathymetry of this region may therefore represent five EVRs or just two larger ones with en echelon structures within them, or the presence of a secondary, larger-scale segmentation in this region (Johnson & Jakobsson, 1985).

The profiles in Figure 6.10 are taken at various orientations to the axes of EVRs 2 and 3. Profiles 1 and 2 are parallel, and are perpendicular to the elongate axis of the Steinahóll EVR (EVR2). Both profiles are similar showing undulating rounded topography (120 m relief) and a few steeper scarps. These are probably seamounts and faults, respectively. Profiles 3 and 4 cross EVR 3 and are at right angles to each other. The main feature on this profile is the large seamount (150 m relief) that has a complex symmetrical distribution of troughs and peaks on its otherwise flat-topped surface. The estimated plan view of this is also indicated (Figure 6.10c). Profile 5 is across a deep area to the north of EVR 3. The topography is jagged and may represent faulted terrain. From the bathymetric information in Figures 6.9 & 6.10 it can be seen that EVR 2 is much more robust than EVR 3, the outline of which is much less well defined. It is essentially composed of only a few large seamounts. EVR 2 is in a different tectonic state to EVR 3.

Together the profiles in Figure 6.10 indicate that the structure of the EVRs at 63°N are very similar to those in areas A, B and C (Figure 6.4), with seamounts, faults and undulating terrain, but that there is much less relief in the EVRs in this more northerly area (150 m compared to 400 m in areas A, B and C). These EVRs are also less robust than the partly buried Icelandic volcanic systems on the Reykjanes Peninsula (350 m maximum relief). This may suggest either significant burial of the EVRs at 63°N by sediment derived from the Iceland land mass, or a low magmatic flux over the past 1 Ma relative to other areas.



Figure 6.10: Bathymetric profiles from the single beam echo-sounders. Their locations are marked on Figure 6.9. Profile 2 includes the image of the hydrothermal plume.

## 6.3.2.2) Water column

The water column profiles were examined at 23 CTD stations (Figures 6.9). Each profile measures the variation in the potential temperature, potential density, salinity and light transmissivity of the water with increasing depth. An eruption event on the sea-floor would alter the temperature, density and transmissivity of the background profile by emitting heat, rock particles and gases into the water column. The problem found during the B19-90 cruise was the absence of information on the background profile for this area, other than those we obtained to the north and south of the suspected eruption site. However, the degree of instability at each station was assessed in terms of how well mixed the water columns were accompanied by transmissivity anomaly (Figure 6.11b) which indicates a high particle and/or gas content.



Figure 6.11: Water column profiles from CTD stations: a) an example of a stable water column; b) an example of an unstable water column (i.e. well-mixed water column) from Ólafsson et al. (1991).

Using this approach, an area of the ridge was identified that had consistently unstable profiles (Figure 6.12). This included the Steinahóll EVR (EVR 2) and EVR 3. The anomaly was present on the shallower tops of the EVRs (200-245 m) but it was also present over a deep basin to the N and NW of EVR 2, extending down to >300 m depth. Silica and manganese anomalies (high concentrations) are also reported from this deep basin (Figure 6.12) (Ólafsson *et al.*, 1991).



Figure 6.12: CTD stations differentiated according to the degree of instability in the water column. The chemical SiO2 concentrations (in contour intervals of 2, 4, 8, 16, 32, 64 nmol  $1^{-1}$ ) are also indicated (from Ólafsson et al., 1991). The area shown is the same as Figure 6.9.

During a traverse across the peak of Steinahóll (EVR2) the echo-sounders revealed a turbid zone in the water column. This is interpreted as a hydrothermal plume (Figure 6.13). A subsequent traverse along the length of the ridge revealed several of these features often extending to 10 m below sea-level. This indicates that the Steinahóll EVR is presently hydrothermally active with vents situated along the crest of the EVR. Fishermen report dredging "coloured rocks" from this region for some time (as we did) which indicates that it is probably not a new hydrothermal vent. The location of anomalous water at depths of 300 m (100 m below hydrothermal vent seen at Steinahóll) suggests that there could be a second hydrothermal source in this deep basin. This, coupled with the water chemical anomalies, suggests that this may have been the site of the eruption. However, the evidence is only suggestive and is not conclusive.



Figure 6.13: Photograph of the single beam echo-sounder after the ship had just passed over the Steinahóll crest from NW to SE. A plume of material (particles and/or gases) is seen extending upwards from the crest to very near the surface. This is interpreted as a hydrothermal plume (see also Figure 6.10b).

## 6.3.2.3) Rock dredges

## Sample locations

Dredging took place at various intervals during the CTD survey and the sample locations are indicated in Figure 6.15. In such shallow water (<400 m), accurate positioning of the dredge bag is possible. The details and contents of each haul are summarised in Table 6.1. Most of the dredge hauls recovered abundant fresh

basalt with minimum fauna, but one haul was completely void of basalt fragments (dredge station 2). The bulk of the material recovered is being processed by Dr Sveinn Jakobsson (Natural History Museum, Reykjavík) but one sample from each haul has been investigated here. The petrological summary provided below for the dredged samples represents an initial investigation, and more detailed work and comparative research will be required to fully constrain the petrogenesis of these samples.

Dredge station	Ship-station code	Samples #	Latitude	Longitude	Depth	EVR #	Dredge haul contents	Beselt type
D1	890	D1cw	63'05.6'N	24°32.3W	250	2	Basalt & corals	B
D2	904		63°10.5'N	24°26.5W	330	3	Sediments	
D3	905	D3cw	63'08.6'N	24 <b>*2</b> 9.4W	235	3	Basalt	в
D4	909	D4cw	63°15.5'N	24°12.8W	93	5	Basatt	B
D5	910	D5cw	63°13.6'N	24°20.8W	215	4	Basalt & shells	в
D6	911	D6cw	63'11'N	24°24W	360	3	Basatt	A
D7	913	D7acw	63*05.3'N	24 <b>°3</b> 2.7W	160	2	Basalt & hyaloclastite	в
D7	913	D7bcw	63'05.3'N	24 <b>*3</b> 2.7W	160	2	Basalt & hyaloclastite	в
D8	914	D8cw	63'05.4'N	24°32.6W	150	2	Basatt	в
D9	915	D9cw	63*05.4'N	24,32.5W	230	2	Basalt & coloured, altered basalt fragments	в
D10	917	D10cw	63°09.2'N	24°25.7W	220	3	Basah	A
D11	918	D11cw	63*09.7*N	24°25.2W	245	3	Basatt	A
D12	919	D12cw	63'08.5'N	24°32.3W	290	2	Mud & shells with a few basalt fragments	A

 Table 6.1: Petrography of the 12 dredge hauls.

## Petrography of the basalts

There is very little petrographic diversity amongst the basalts from the four EVRs sampled here (EVRs 2-5) which occupy 45 km of the ridge. All the basalt samples are either very fine grained or glassy, highly vesicular, and essentially aphyric with few intratelluric microphenocrysts of feldspar and sometimes olivine. There are abundant quench microphenocrysts of feldspar, with or without olivine and clinopyroxene. The quench microphenocrysts display a radiating habit, similar to that displayed by the intratelluric microphenocrysts of feldspar by petrological type 3 basalt (Section 4.3; Table 4.3, p. 87).

The only discriminating petrographic feature in the basalts is the type of quench phases present, and on this basis, the samples can be divided into two groups. In some samples quench olivine dominates over clinopyroxene (type A), and in others, the situation is reversed (type B). These types are exemplified in Figure 6.14, and are indicated for each sample in Table 6.1. The regional distribution of the two types is indicated in Figure 6.15. It can be seen that there is no consistent relationship



Figure 6.14: Examples of the two petrological varieties of basalt shown in XP: a) Type A - quench microphenocrysts of olivine dominate over clinopyroxene (sample D11cw); b) Type B - quench microphenocrysts of clinopyroxene dominate over olivine (sample D7bcw). Field of view in both a & b is 18 mm.

between the petrography of the sample and the EVR with which it is associated. In light of the present bathymetric data available it can also be said that there is no consistent relationship between the petrography of the sample and the location of the samples within a single EVR. The variation in the quench phases is thought to represent subtle differences in the degree of differentiation of the basalt.



Figure 6.15: Map illustrating the distribution of the two types of basalt on the ridge.

## Compositional variation amongst the dredged basalts

Chemical analyses were made for the basalt samples employing the same analytical methods used for the Icelandic samples (Chapter 5, Appendix 3). They include analyses of selected major element oxides (FeO\*, CaO, Na<sub>2</sub>O, K<sub>2</sub>O, TiO<sub>2</sub>, MnO and P<sub>2</sub>O<sub>5</sub>) and trace elements (Nb, Zr, Y, Sr, Ga, Ni, Cr, V, Sc, Rb, Ba, La, Ce and Nd) from pressed powdered pellets on the XRF. Fused discs were made for two samples (D3cw and D7acw) and were analysed for all the major element oxides (SiO<sub>2</sub>, Al<sub>2</sub>O<sub>2</sub>, FeO\*, MgO, CaO, Na<sub>2</sub>O, K<sub>2</sub>O, TiO<sub>2</sub>, MnO and P<sub>2</sub>O<sub>5</sub>) on the XRF. The data are tabulated in Table 6.2.

Sample EVR	D1cw	D3cw	D4cw	D5cw	D6cw	D7acw 2	D7bcw	D8cw	<b>D9cw</b>	D10cw	D11cw	D12cw
Location	63'05.6'N 24'32.3'W	63°08.6'N 24°29.4'W	63°15.5'N 24°12.8'W	63'13.6'N 24'20.8'W	63'11'N 24'24'W	63'05.3'N 24'32.7'W	63°05.3'N 24°32.7'W	63°05.4'N 24°32.6'W	63°05.4'N 24°32.5'W	63°09.2'N 24°25.7'W	63*09.7N 24*25.2W	63'08.5'N 24'32.3'W
Depth	250	235	93	215	360	160	160	150	230	220	245	290
Group	A	A	В	в	в	A	A	A	A	A	A	A
SiO2		49.61				49.94						
A12O3		13.43				13.23						
Fe2O3	15.47	14.03	15.32	16.09	16.43	14.29	15.25	15.31	13.86	14.95	15.01	14.85
MgO		7.02				6.53						
CaO	11.77	11.28	11.13	11.26	11.39	11.05	11.74	11.76	11.68	11.91	11.91	12.90
Na2O	1.83	2.35	2.49	2.23	2.02	2.24	1.95	1.88	2.07	1.89	1.70	1.82
K2O	0.10	0.24	0.15	0.16	0.16	0.15	0.11	0.11	0.13	0.10	0.10	0.12
TiO2	1.43	1.40	1.81	1.90	1.76	1.39	1.44	1.47	1.46	1.40	1.41	1.45
MnO	0.23	0.22	0.22	0.24	0.24	0.22	0.23	0.23	0.22	0.23	0.23	0.23
P2O5	0.10	0.16	0.11	0.13	0.13	0.15	0.11	0.12	0.11	0.12	0.10	0.11
Total	100.80	99.71	100.92	101.22	100.27	99.17	100.66	101.56	101.93	101.00	100.70	100.47
Mg #	46	51.2	42	40	40	48.9	45	45	50	46	47	47
Nb	7	8	10	10	9	8	7	8	7	6	7	8
Zr	77	77	101	103	94	78	77	79	79	74	75	72
Y	32	30	37	37	34	32	31	32	33	29	32	28
Sr	88	110	112	111	101	90	88	89	97	81	82	94
Ga	20	18	14	20	20	18	15	18	22	19	20	22
Ni	67	60	64	64	65	63	63	61	60	63	71	65
Cr	62	87	96	93	104	62	65	<b>59</b>	63	38	56	87
V	386	347	408	432	429	385	385	390	407	393	381	379
Sc	51	43	46	49	52	50	49	54	47	50	49	48
Rb	1	6	3	4	7	3	4	3	3	4	3	3
Ba	42	40	50	46	46	39	35	46	41	39	37	43
La	3	4	4	4	5	1	3	5	1	5	1	4
Ce	15	22	17	21	18	12	17	17	12	15	16	19
Nd	7	12	11	9	11	9	12	10	4	11	8	9
<b>n</b>		-				-						

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**Table 6.2:** Chemical compositions of 12 samples dredged from the Reykjanes Ridge (B16-90 cruise, November 1990). The data were acquired by the same analytical techniques as the Icelandic samples (Appendix 3).

Key to EVR (See Figure 6.9)

1=S of Steinahóls 2=Steinahóls 3=NE of Steinahóls 4=NE of 3 5=NE of 4

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The petrographic character of these samples is typical of that described as tholeiites (qz-normative) dredged to the north of this location (Brooks *et al.*, 1974; Jakobsson, 1974; Johnson & Jakobsson, 1985). However, these samples include both qz-normative and *ol*-normative samples (D3cw and D7acw, respectively). In Figure 6.16, the samples are plotted in variation diagrams of the sort used in Chapter 5 (Section 5.3) and are differentiated according to which EVR they were dredged from. The composition of the basalts is very uniform within the suite, and there is no consistent relationship between the petrographic type and the compositions of these basalts (Figures 6.15 & 6.17).



Figure 6.16: Composition of the dredged samples differentiated according to the EVR (2-5) from which they were collected (see key). The various elements are plotted against Ni ppm. Precision error bars ( $2\sigma$ ), as assessed by the deviation of replicate analyses are indicated in the bottom left-hand corner of the diagram, when greater than the size of the data point. Error ( $2\sigma$ ) for Ni is 12 ppm (all values are tabulated in Appendix A, p. 308).

In Figure 6.16, most of the variation observed is within the limits of analytical error. There are higher abundances of TiO<sub>2</sub>, Nb & Zr for three of the samples, but the distinction becomes less pronounced for elements with larger error. The samples from EVR 2 and most of those from EVR 3 are compositionally indistinguishable, whereas three samples from further north (samples D4cw, D5cw & D6cw) have higher abundances of incompatible elements at the same degree of differentiation. The samples can be said to form to compositional groups, A and B. These are indicated in Table 6.2 for each samples. Group B samples are spatially separated from the rest (Group A), and are ,from EVR 4, EVR 5 and from the basin to the north of EVR 3 (Figure 6.17). It is not clear whether the sample from dredge station 6 (D6cw) at 360 m depth is derived from EVR 3, as the location and the bathymetry suggest, or if it is derived from EVR 4. It is noted that the subtle compositional break along the ridge axis occurs after the small lateral offset of the axis between EVRs 3 and 4. This subtle compositional difference may represent a correlation between melt composition and ridge segmentation (Klein & Langmuir, 1987), or it may represent the chemical gradient reported to exist for the Reykjanes Ridge (Schilling, 1973).



Figure 6.17: Map illustrating the distribution of the two compositional types of basalt on the ridge.



Figure 6.18: The composition of the dredged basalts relative to the eruptive units of the Hveragerŏi Volcanic System. Precision error bars (2 $\sigma$ ), as assessed by the deviation of replicate analyses on International Reference Materials and internal monitors, are indicated in the bottom left-hand corner of the diagram, when greater than the size of the data point. Error (2 $\sigma$ ) for Ni is 12 ppm (all values are tabulated in Appendix 4, p. 308).

In Figure 6.18, the samples are examined relative to the compositions of the Icelandic basalt samples from the Hveragerði Volcanic System. The degree of compositional overlap is variable between elements, however, the two suites are completely separated for certain trace elements. The group has high abundances  $FeO^*$ , and also SiO<sub>2</sub>, which overlap with those of the compositionally uniform TA units from the Grensdalur region, but have lower abundances of most other major

element oxides (e.g.  $Na_2O$ , and also  $K_2O \& TiO_2$ ). Trace element abundances are generally lower than the Hveragerði samples (e.g. Nb, Sr, Zr, Cr, and also Ba, La, Ce & Nd) and this characteristic of the marine samples is particularly well pronounced for the more incompatible trace elements (e.g. Sr, and also Ba). The marine samples do, however, overlap completely for Y and V, and have higher abundances of Sc (not shown).

The Reykjanes Ridge basalts do not contain any megacrysts (petrological type 2 basalt; Section 4.3) that would otherwise imply input of instantaneous melts from a depleted DMM source (Section 5.6.2.4). Neither are they saturated with intratelluric olivine and radiating feldspar crystals (petrological type 3 basalt) which is thought to indicate passage through the crust without interaction with a shallow crustal reservoir. Instead, relative to the model developed in Section 5.5.5 (Figure 5.21, p. 142), these marine samples are equivalent to the petrological type 1 basalt, and the uniform composition suggests that they were erupted from a magma reservoir that was thoroughly well mixed. It is noted that there is no significant difference in composition or petrographic character the two basalts sampled from the well defined seamount on EVR3 (Figure 6.10), other than quench olivine is more abundant than clinopyroxene.

The variable amount of overlap for the incompatible trace elements in Figure 6.18 suggests that the marine samples have different trace element ratios compared to samples from the Hveragerði Volcanic system. This may be caused by either differences in the trace element composition of the mantle source region, differences in the mantle potential and solidus temperatures, or differences in the melting process involved in the generation of these basalts. In Chapter 5, the solving such petrogenetic problems was addressed by using the combined observations from mineral, volcanic glass and REE data. However, such data types do not exist for the marine samples, as yet, and so the petrogenetic investigation is limited. However, abundances of incompatible trace elements do exist. In Figure 6.19, the ratios of Nb/Ti and Zr/Nb are plotted for the marine samples with petrological type 3 basalt and the TA units from the Icelandic study area. Petrological type 3 samples are included here because their origin has been investigated and discussed in detail in Section 5.6. The TA units are included because they are fractionated by similar amounts to the marine samples, and have similar major element oxide abundances (Figure 6.18a). Modelled trajectories for partial melting are also plotted in Figure 6.19 and the derivation of these is described in detail in Section 5.6.2.3. The trajectories represent accumulated

melts derived from melting mantle with variable starting compositions (DMM and BSE), and variable mantle solidus T<sup>o</sup>C.



**Figure 6.19:** The composition of the dredged basalts relative to the eruptive units of the Hveragerői Volcanic System to illustrate the relative degrees of partial melting according to modelling which is fully described in Section 5.6.

In Figure 6.19, the marine samples have slightly higher Nb/Ti ratios at a given Zr/Nb ratio relative to the Icelandic samples and lie on the DMM-1400°C trajectory. In Section 5.6.2.4, however, it was demonstrated that trace element ratios may be the result of more than one petrogenetic process. Following the rationak outlined in Section 5.6.2.3, the slightly higher Nb/Ti ratios for a given Zr/Nb ratio of the marine samples may indicate that they are derived from either a DMM at 1400°C, or from a BSE mantle with a slightly lower mantle solidus T°C than the Icelandic samples. Either interpretation is appropriate for the relative locations of these two suites with respect to the centre of the Iceland plume.

Icelandic samples with similar Zr/Nb ratios are those interpreted to be the products of instantaneous melting from a very depleted DMM source with a mantle solidus T°C of 1400°C (e.g. unit GB4; Section 5.6.2.4). The major element oxide composition of unit GB4 (with high Mg # and low FeO\* and CaO abundances; Appendix 4, p. 310) is very different from those of the marine samples. This difference suggests that the trace element ratios for GB4 and the marine samples are not caused by the same melting processes, as such processes have a significant effect on the major element oxide abundances in the magmas (Section 5.6). Icelandic samples that do have similar major element oxide abundances are those of the TA units (for Mg # and FeO\*; Appendix 4, p. 314 - 317). In Figure 6.19, the TA and marine samples do not overlap in their trace element abundances. The TA units occur at 2% PM on the BSE-1400°C trajectory, whereas the marine samples occur around the 2% PM of the DMM-1400°C trajectory. This suggests that the differences in trace element compositions between the TA and the marine samples is caused by a difference in the source region composition, and not mantle P-T conditions or melting mechanisms. The Zr/Nb ratios of the marine samples are likely to represent an increased level of depletion in the source region relative to the enriched mantle under the Hengill-Ingolfsfjall area, but they may be derived from a mantle with a similar mantle solidus T°C. A more extensive data set is required to support this hypothesis.

## 6.4) Summary

During this research, sonar data and rock samples were acquired from the EW9008 and B16-90 cruises to the Reykjanes Ridge. The ridge profile on this obliquely spreading ridge varies from a typical slow spreading ridge in the south, to one that mimics a fast spreading ridge in the north. The neovolcanic zone is segmented into en echelon volcanic ridges ( $5\times30$  km). These are constructional features built up of seamounts and other eruptive products and are cut by variable degrees of faulting. The interplay of faulting and volcanism produces EVRs that appear to be in different tectonic states.

The Reykjanes Peninsula displays a number of en echelon volcanic systems. The hyaloclastite hills are qualitatively compared to the submarine EVRs. In this comparison, such phenomena as weathering, differential eruptive environments within Iceland, and the differences between the subglacial and submarine eruptive environments have to be born in mind. The Iceland plume seems to result in larger volcanic features further inland. From this comparison, it is suggested that subglacial

lava shields may be analogous to seamounts, while the more linear structures on the seafloor could be produced from fissure type eruptions.

The second cruise was intended to investigate a seismic event 150 km south of Iceland (63°10'N). The 45 km section of ridge under investigation consisted of five EVRs (or two larger ones), and the locus of activity was centred between EVR2 (Steinahóll) and EVR3, in a 300 m deep basin. The location was marked by instabilities in the water column, coupled with water-chemical anomalies. Twelve dredge hauls were collected from this section of ridge. The basalt is very similar in petrographic character and chemical composition. This may represent magmas erupted from a well mixed magma chamber. Samples from the more northerly EVRs have slightly higher incompatible trace element ratios at a given degree of differentiation that the southerly ones. This may indicate the presence of two separate reservoirs that correspond with the ridge segmentation (Klein & Langmuir, 1987), or it may be sampling part of the geochemical gradient reported to exist for the Reykjanes Ridge (Schilling, 1973). By comparing the marine samples to the petrogenetic modelling in Chapter 5, it is inferred that they could be the results of melting of a DMM source at 1400°C (solidus temperatures). This means that these basalts are from a source region with hotter mantle solidus T°C than normal MORs (1400°C), and similar mantle solidus T°C to Iceland. The source is less-enriched in trace elements than the Iceland study area.

Chapter 6

# **Chapter 7**

## **Synthesis and Conclusions**

## 7.1) Synthesis

## 7.1.1) Introduction

This chapter provides a summary of the main observations documented in this thesis, from this and other research, that are directly relevant to the main conclusions that will be arrived at in this chapter. A magmatic plumbing system model for the Hengill to Ingolfsfjall region is presented in Figure 7.1. It is constrained by both geophysical and petrological observations. This model and its relevance to studies of mid-ocean ridges will be discussed in this chapter.

## 7.1.2) Summary of observations

## 7.1.2.1) Background research

The Iceland plateau is the result of a hot mantle plume that has an enriched trace element geochemical signature, relative to chondrite and MORB (e.g. Schilling, 1973). Adiabatic decompression mantle-melting models suggest that hotter mantle should produce thicker crust by undergoing larger degrees of partial melting (McKenzie, 1984). In the study area the crust, as defined by geophysists, is thought to be approximately 7.5-10 km thick at the axis (Pálmason, 1971; Hersir, 1980; RRIPS, 1985; Hersir *et al.*, 1990; Flóvenz & Gunnarsson, 1991) compared to an average of 6 km for other slow-spreading MORs (e.g White, 1989). The area encompasses three volcanic systems (Hengill, Hrómundartindur, and Hveragerði) and contains the active Hengill and the extinct Grensdalur central volcanoes (Sæmundsson, 1967; Björnsson *et al.*, 1974; Foulger & Toomey, 1989). The crust was produced over a 900,000 year period by both lava shield and fissure-type eruption events.

Geophysical surveys reveal higher densities beneath the Grensdalur Central Volcano and the lava shields (Hæðir, Húsmúli and Skálafell) and lower densities to the north and south of the Hengill mountain (Thorbergsson *et al.*, 1984; Foulger & Toomey, 1989; Hersir *et al.*, 1990). Seismic tomography has constrained the position and dimensions of the crustal structures causing the density anomalies. The high-density volumes under the lava shields may be in the form of conduits extending to at least 5 km depth (Figure 1.7, p. 14), whereas the high-density volume under Grensdalur (40 km<sup>3</sup>) extends from the surface down to only 3 km. The low-density volume beneath the Nesjavellir valley, to the north of the Hengill mountain, is small (5 km<sup>3</sup>) and situated between 2-4 km depth (Foulger & Toomey, 1989).

7.1.2.2) Observations from this research

## Petrography

Amongst the rocks studied in this thesis, the intratelluric feldspar phenocrysts occur as two essentially mutually exclusive varieties: the feldspar megacrysts that tend to form agglutinating glomerophyric clusters; and the tabular microphenocrysts that form radiating glomerophyric clusters. Feldspar crystals observed in gabbroic xenoliths are of the megacrystic variety. On this textural basis, the samples are divided into three petrological types, 1-3 (Table 4.3, p. 87). Type 1 basalt is aphyric; type 2 is megacrystic; and type 3 has radiating clusters of feldspar microphenocrysts. Examination of the regional distribution of these basalt types reveals that they are generally confined to different volcanic eruption types (lava shields and fissures). Petrological type 1 and 2 basalt occurs in the fissure-type eruptive units, with type 2 basalt (within the Hveragerði Volcanic System) generally restricted to fissures in the vicinity of the Grensdalur Central Volcano. Ninety percent of the occurrences of petrological type 3 basalt are found in lava shields (and the Hengill mountain), and the remaining ten percent are in eruptive units that are too poorly exposed to attribute to a particular eruption type. Temporally, the petrological type 2 basalt becomes progressively more abundant with time in the vicinity of the Grensdalur Central Volcano.

The cores of the plagioclase megacrysts are nearly constant in composition (An 80-90) and out of equilibrium with the host sample. The radiating tabular feldspar microphenocrysts are also out of equilibrium with their host, but by lesser and more varied amounts. Computer modelling of crystallisation processes suggests that the assemblages within petrological types 2 and 3 are *not* the products of closed-system magmatism. However, the variety of phenocryst types and compositions,

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coupled with the modelling results suggest that petrological type 2 basalt may be the product of classic open-system magmatism (O'Hara, 1977), whereas petrological type 3 basalt is not.

## Geochemistry

The petrological type 1 samples (whole-rock and glass) have a uniform composition, whereas petrological type 2 samples (whole-rock) are scattered and fall along modelled trajectories that represent feldspar megacryst addition to aphyric magmas. There is as much scatter observed for the petrological type 3 whole-rock samples as there is for petrological type 2, yet in this case, the variation cannot be explained by phenocryst accumulation. The glass compositions of the two porphyritic suites, petrological types 2 and 3, define parallel liquid lines-of-descent, suggesting that they are not petrogenetically related to each other by fractional crystallisation. However, this is not directly supported by normative projections, which suggest that the same samples of petrological type 1 and 2 liquids could be derived from petrological type 3 liquids by fractional crystallisation.

When the samples are examined in terms of stratigraphic distribution, a pattern emerges whereby the composition alternates from being monotonous to varied, and the degree of variability correlates with the petrological types of basalt found there. The extent of and variation in differentiation become more enhanced in the younger units of the Hveragerði Volcanic System, which occur in the vicinity of the extinct Grensdalur Central Volcano. Basalts in the lava shields and the main Hengill eruption are generally less fractionated, but each shield is distinct in elemental abundances at a given degree of differentiation. Modelling and examples of vertical profiles within individual shields suggests that the magmas being progressively supplied to the eruptions have *subtle* differences in the degree of differentiation, as well as the incompatible element abundances and ratios.

From these petrological observations three different styles of crustal magmatic plumbing system are envisaged (Figure 5.21, p. 142). According to this model, the petrographic character and degree of differentiation of the basalt is entirely dependant on which of the three paths through the crust is followed by the magmas. This is supported by the REE signatures of the basalts (that are unaffected by crustal differentiation processes) which show no correlation with the petrological type or the type of volcanic eruption event (i.e. lava shield or fissure) of the samples. There is a well-defined trend between FeO\* and  $[La/Yb]_n$ . Samples with high FeO\* are also accompanied by enriched incompatible element ratios, such as  $[La/Yb]_n$ , compared to low FeO\* samples. Petrogenetic modelling supports the following conclusions: 1) highly calcic plagioclase (An >80) can only crystallise from instantaneous melts from a depleted DMM source (i.e. top and crest of the melting column); 2) shallower melting produces a low-FeO\* signature in the basalts (polybaric melting); 3) a given incompatible element ratio may be produced in more than one way by varying the source composition, mantle P-T conditions and the degree of melting.

When the Icelandic samples are examined in light of this information, their genesis can be suggested. For example, Ingolfsfjall is an eruptive unit that could have been produced from accumulated melting from depth that mixes incompletely with depleted melts from the top of the melting column just prior to eruption. Unit GB4 represents instantaneous melt produced at shallow depths from the depleted crest of the melting column that is segregated and erupted without mixing with accumulated melts. Two eruptive units within the area, S1 and GB4, are almost identical in composition in terms of FeO\* and incompatible trace element signature, and they both possess equilibrium calcic plagioclase (An >80), but only one has megacrysts (i.e. S1; petrological type 2 basalt). This megacrystic sample occurs in the vicinity of the Grensdalur Central Volcano. Most basalt from the study area may have been produced from melting a hot (solidus temperatures of 1400°C) and enriched mantle source (more enriched than BSE).

## Oceanography

On the Reykjanes Ridge, the ridge profile changes from a typical slowspreading ridge in the south, to one that mimics a fast-spreading ridge in the north. The neovolcanic zone is segmented into en echelon volcanic ridges (EVRs) ( $5\times30$  km) that are comparable in dimensions to the volcanic systems on the Reykjanes Peninsula. Constructional features on the EVRs include seamounts, pillow lavas (often with a linear fabric) and other eruptive products, all of which are cut by variable degrees of faulting. The interplay of volcanism and faulting results in EVRs that appear to be in different tectonic states. In a qualitative comparison of the EVRs with the subglacial volcanic systems on the Reykjanes Peninsula, factors such as surface erosion, variable eruptive environments and the plume effect on the volume of magma produced have to be borne in mind. Such a comparison suggests that lava shields may be analogous to seamounts, while the more linear structures on the seafloor could be produced from fissure type eruptions. Twelve dredge hauls were collected from 63°10'N, the site of the possible November 1990 eruption event. The basalts are all aphyric and uniform in chemical composition, but samples from the more northerly EVRs have slightly higher incompatible trace element ratios at a given degree of differentiation. From this initial investigation, there does not appear to be any correlation between the volcanic morphology and the basalt petrography or composition. If the dredged basalts are compared with the Iceland samples from the study area, they are analogous to petrological type 1 basalt and, following the petrogenetic model presented here, they require a convecting magma reservoir in which to fractionate prior to eruption (Figure 5.21, p. 142). Compositional differences exist between the Iceland and marine samples in their incompatible element ratios and abundances. According to petrogenetic modelling, the marine samples could have been produced by melting a DMM source with a solidus temperature of 1400°C.

#### 7.1.3) Synthesis and discussion

## 7.1.3.1) Melting dynamics

Recent ideas on mantle melting at ocean spreading centres involve a dynamic melting column (e.g. Langmuir *et al.*, 1977; McKenzie, 1984; Klein & Langmuir, 1987; McKenzie & Bickle, 1988; Kostopoulos & James, 1992). If the melts produced are accumulated to a point-depth average (McKenzie & Bickle, 1988) larger degrees of melting bring the composition of the liquids produced closer to that of the source. Therefore, large degrees of melting of the enriched Icelandic source should produce magnesium-rich, trace element *enriched* basalts, as oppose to magnesium-rich, trace element degree of melting of a DMM source.

A recent study by Elliot *et al.* (1991) on Icelandic basalts from the NVZ and the WVZ (Mg # 58-82) provided some constraints for the dynamic melting processes that operate beneath the Icelandic neovolcanic zone. These authors found that, in contrast to the relationship predicted by existing melting models (e.g. high MgO and trace element enriched; McKenzie & Bickle, 1988), the highest MgO basalts sampled (15-18 wt. % MgO) had the most depleted incompatible element ratios. These samples also had unexpectedly low FeO\* abundances. The MgO-rich, FeO\*-poor melts represented by such samples are interpreted as small accumulations of instantaneous melt from the top of the column where the source region has already experienced previous melting at deeper levels. The depleted nature of this mantle results in high MgO, high CaO/Al<sub>2</sub>O<sub>3</sub> ratios and depleted incompatible element

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ratios, and the shallowness of the melting event causes the low FeO\* signature. Melts that accumulated from the entire length of the column will also have high MgO, high  $CaO/Al_2O_3$  ratios but will be enriched in their incompatible element ratios, and the depth of the melting will produce a high FeO\* signature in the basalts. The effect of this melting process on the CaO/Al\_2O\_3 ratios adequately explains the "phantom clinopyroxene" in MORB (Elliot *et al.*, 1991; Nicholson, 1990). The relationship between these compositional parameters (MgO, FeO\*, CaO/Al\_2O\_3 ratios and the degree of trace element enrichment) is also observed in the basalts from this research, and the interpretation provided by Elliot *et al.* (1991) is therefore supported here.

The resulting liquids produced by this modified dynamic melting column model (Elliot et al., 1991) should be either MgO-rich, FeO\*-rich, and enriched in the highly incompatible elements (very dense), or MgO-rich, FeO\*-poor, trace element depleted liquids (less dense). However, the FeO\*-rich melts have lower than expected MgO, according to Elliot et al. (1991). To explain this they suggest that the dense melts (high FeO\*) are trapped under the less dense melts (low FeO\*) until they have fractionated sufficient olivine (i.e. reducing their MgO content and density) which allows them to mix and erupt. This is not supported by the primary melt compositions used in the modelling here. The accumulated melts of Nui & Batiza (1991) have primary Mg # values of approximately 75, whereas the instantaneous melts from McKenzie & Bickle (1988) have Mg # of approximately 80. Ingolfsfjall basalt (accumulated origin) has Mg # 59, whereas unit GB4 (depleted & instantaneous origin) has Mg # of 70. The basals in both of these units have similar abundances of olivine. This suggests that more extensive fractionation of the accumulated melts prior to eruption is unlikely. The systematic variations in the incompatible trace element ratios in the vertical lava profiles within individual shields suggests that magmas are supplied to the eruption in batches straight from the mantle. This also suggests that extensive mixing and fractionation of the accumulated melts does not occur. However, mixing of the accumulated melts with the depleted instantaneous melts (i.e. part of the accumulation process) could be the mechanism that causes the variation in the An content of the tabular feldspar found in petrological type 3 basalt. Long periods of ponding of the accumulated melts at the base of the crust is not supported here.

## 7.1.3.2) Magmatic plumbing system

## Introduction

In addition to providing supporting evidence to this recently enhanced model of mantle melting (Elliot et al., 1991), the basalts from the research area studied here

provide constraints on the history of the magmas once they have segregated from the mantle, and the character of the magmatic plumbing system in this part of the constructive plate boundary. The data set presented in this thesis has the support of detailed field study (Chapters 2 & 3; and Árnason *et al.*, 1987) and numerous geophysical investigations into the seismic and density structure of the crust in the region (see Table 1.1; e.g. Thorbergsson *et al.*, 1984; Foulger & Toomey, 1989; Hersir *et al.*, 1990). The collective evidence suggests that there are two alternative routes that a magma can follow once it has segregated from the mantle: straight to the surface (lava shields); or its ascent may be interrupted by a crustal reservoir (fissures and central volcanoes) (Figures 5.21 & 7.1). In this section, this model for the magmatic plumbing system for the Hengill to Ingolfsfjall region is discussed, and its relevance to off-shore magmatism considered.



Figure 7.1: Model for the magmatic plumbing system beneath the neovolcanic zone of S. W. Iceland, based on observations from the Hengill-Ingolfsfjall region. The model is constrained by both geophysical and petrological observations reported here: A) lower magma supply; B) higher magma supply. The plumbing system is envisaged as highly ephemeral, and the position of the crustal reservoir may be situated deeper than 2-4 km (as above - based on the present day Hengill Volcanic System).

## Lava shields

It is suggested here that the lava shields are fed by magmas coming straight from the base of the crust/lithosphere (Figure 7.1A). Evidence for this comes from the density structure of the crust (geophysical results), the shield morphology and the petrology of the basalts. Where data exist all the lava shields are underlain by high density volumes of the crust, and for one shield (Húsmúli) the density anomaly has the shape of an elongate conduit extending to considerable depths ( $\geq 5$  km) (Foulger & Toomey, 1989). The magma effusion rate in the lava shields is low compared to the fissure eruptions, and this is thought to be because the shields originate from much deeper reservoirs (G. P. L. Walker, 1971; Sæmundsson, 1979). This suggestion follows from the assumption that magma effusion rate is controlled by the variable lithostatic head (i.e. the thickness of the overlying material relative to the buoyancy of the magma). However, magma composition will also have an effect on the effusion rate. The shape of the domes for basalt of similar composition could therefore reflect the depth of the source region for individual shields with flatter ones coming from shallower reservoirs (Rutten, 1964; G. P. L. Walker, 1971).

The radiating habit of the tabular feldspar microphenocrysts mimics that displayed by the quench crystals, yet the former exist with true quench crystals in the same rock so that they could not have been formed *at* the surface. It is proposed here that such a radiating habit develops during crystallisation en route to the surface in a conduit, such as that imaged for Húsmúli. The fact that this texture is a characteristic of the lava shields suggests that the other lava shields may each have elongate cylindrical conduits that extend at least to the lower crust. Further targeted geophysical studies to investigate this would be a valuable test of this hypothesis. The well developed oscillatory zoning of these feldspars may have been produced by turbulent flow in a conduit of considerable length (Sparks *et al.*, 1980) again implying a deep reservoir for the lava shields.

It is suggested that each conduit is isolated from any others and from shallow crustal reservoirs. Evidence for this comes from the spatial distribution of the shields and their distinctive basalt compositions. Also, the Húsmúli conduit extends past the low-velocity volume to the north of the Hengill mountain and down into the lower crust. Additionally, the absence of feldspar megacrysts and the abundance of olivine phenocrysts in the basalts suggests that the two porphyritic suites do not interact with each other. It may be possible that a reservoir for the lava shields exists in the lower crust. However, modelling of the mineralogy and whole-rock compositions indicates that the products of this style of magmatism do not fit the classic open- or closedsystem magmatism, both of which assume a crustal reservoir. The variation within a vertical profile of lavas indicates that the magma supplied to the eruption varies in trace element signature and the degree of differentiation which is in accordance with magmas coming in batches straight from the mantle. Magmatic processes operating in crustal reservoirs would homogenise, mix and contaminate the magmas with foreign phenocrysts, for all of which there is no evidence. Basalt from the lava shields, and probably other petrological type 3 basalt, is the most likely to provide information on the melting processes and compositional state of the source region.

Húsmúli is different from the other lava shields in its extent of differentiation. If the same logic used to infer the style of plumbing system, as described above for the other lava shields, is applied to the interpretation of the origin of Húsmúli, contradictory evidence is noted. The deep nature of the plumbing system feeding the eruption is supported by the imaged conduit, the feldspar texture and the shield morphology. The systematic compositional breaks observed in the lava profile suggests that the magmas are supplied to eruption from a source of magma that is periodically diluted by new magmas of more primitive composition. The fractionated nature of the lavas, the absence of abundant olivine and the rare occurrence of pyroxene all suggest that the Húsmúli magmas have resided in a chamber. If this is the case, how can the compositional breaks within the lava stratigraphy be formed? Additionally, it is not feasible for all the feldspar to have crystallised en route to the surface subsequent to extensive olivine and pyroxene fractionation. The feldspars have a wide range in An content (85-72) which, according to the modelling presented here, requires an input of depleted instantaneous melts, and also that these small quantities of melt are not extensively mixed with the accumulated melts. These observations for Húsmúli indicate that the lava shield magmatic plumbing system model described above is, as yet, incomplete. However, the available evidence is insufficient to solve this problem and further the model at present.

The Hengill mountain is not thought to represent the central volcano within the Hengill Volcanic System. Evidence for this comes from the morphology of the subglacially derived mountain, and is supported by the petrology of the basalts, which are comparable in texture and composition to those of the Ingolfsfjall shield. When the area was first mapped, the mountains of Hengill and Ingolfsfjall were both described as table mountains (Sæmundsson, 1967).

Lava shields have been mapped in the older Tertiary Icelandic lava piles (Rutten, 1964), and therefore it is unlikely that they are only the result of favourable

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isostatic conditions subsequent to the onset of deglaciation (Gudmundsson, 1986). If the model presented here is realistic, the only requirement for the eruption of a lava shield is magma availability and the absence of an interfering crustal reservoir. The presence of lava shields may therefore provide a mapping device to detect the present absence of the crustal reservoirs through the neovolcanic zone of Iceland. The effect of the Icelandic plume may manifest itself in the progressively larger size of these eruptive units towards the centre of Iceland (e.g. the Langjökull region). Theoretically, a thicker crust should enhance features such as the dome shape of the shields and the development of the radiating feldspar texture. Again, further work could test this hypothesis.

#### Fissures and central volcanoes

The plumbing system for a fissure eruption is thought to involve a shallow crustal reservoir (Figure 7.1). The presence of such a reservoir in the Hengill Volcanic System is supported by a small volume (~5 km<sup>3</sup>) of exceptionally low seismic velocity situated between 2-4 km depth within the crust. If low P-wave seismic velocities and gravity lows are interpreted as indicators of partial melt (Mavko, 1980) then the low-density volume might be a magma reservoir (Foulger & Toomey, 1989). The low densities in this area persist despite the expected increase in density caused by hydrothermal epidote mineralisation and the intensity of dyking (80-100%) that are observed between 1.6-2 km depth (Franzson et al., 1988), and also the high density of the Hengill mountain, thought to be a lava shield eruption event. The earthquakes identified from this low-velocity volume suggest that the magma presently exists as pockets of melt as oppose to an open "free" reservoir (i.e. Figure 7.1A as opposed to 7.1B). In the Grensdalur region, below the site of the extinct central volcano, the high densities observed are probably caused by the increase in the percentage of intrusives, in the form of dykes, and epidote mineralisation in the absence of lower density partial melts. This is supported by the localised surface alteration of the rocks to chlorite (Greenschist facies metamorphism) and the numerous dykes mapped at the surface in the Grensdalur region, as part of this research, relative to adjacent areas.

The shallow nature of the magma reservoir that feeds the fissure eruptions in the Hengill Volcanic System is also supported by the distance travelled by lavas erupted from fissures, which is related to the magma effusion rate, which, in turn, is related to reservoir depth (G. P. L. Walker, 1971). The existence of a reservoir is required by the petrography of the basalts (which lack abundant euhedral olivine crystals) and their compositions (fractionated and often uniform). Eruptive units that are thought to have encountered a convecting crustal reservoir (Figure 5.21a) may represent a time-averaged composition for varied melts, whereas eruptive units thought to have encountered a clogged crustal reservoir (Figure 5.21b) may retain their melting signature if they do not mix with existing pockets of melt, but fractionate independently.

Megacrystic plagioclase is common in fissure eruptive units from the study area. There is abundant evidence to suggest that these crystals (and probably associated olivine and pyroxene) represent flotation cumulates (e.g. field evidence, xenoliths and glomerophyric clusters, crystal-melt disequilibrium, etc.). However, this does not constrain their origin prior to accumulation. There have been numerous models proposed to explain the occurrence of such crystals in Iceland (e.g. Trønnes, 1990; Nicholson, 1990; Hansteen, 1991) and at other spreading centres (e.g Aumento, 1968; Bougault & Hekinian, 1974; Donaldson & Brown, 1977; Natland et al., 1983; Griffin et al., 1983). Most of these authors agree that the plagioclase megacrysts represent xenocrysts (i.e. foreign to the host magma in which they occur). Some authors favour accidental xenocrysts (i.e. wall-rock/ mantle crystals; e.g. Trønnes, 1990; Hansteen, 1991), whereas others favour a cognate origin (i.e. derived from a different magma; e.g. Rhodes et al., 1979; Flower, 1980; Natland et al., 1983; Nicholson, 1990). Both origins are possible and distinguishing between them cannot be achieved on compositional and textural grounds alone (Donaldson & Brown, 1977). Authors who support a cognate origin differ in opinion on where exactly the megacrysts crystallised and possibilities include: at depth in the mantle by polybaric crystallisation (Flower, 1980; Natland et al., 1983; Griffin et al., 1983); prior to arrival into the crustal reservoir, either at the base of the crust (Nicholson, 1990), or in some other deeper crustal reservoir (Rhodes et al., 1979); or in the shallow crustal reservoir itself (Aumento, 1968; Bougault & Hekinian, 1974; Kuo & Kirkpatrick, 1982; Larsen et al., 1989).

The realisation by Elliot *et al.* (1991) that highly depleted melts can segregate from the top of the melting column and supply eruptions has provided a crucial constraint on the origin of highly calcic megacrysts. Authors studying the origin of such crystals (also endiopside and highly forsteritic olivine) in highly porphyritic basalts have always been aware that they must have crystallised from refractory (depleted) melts. However, it has not been evident how such melts relate to the much disputed "primary" melt compositions for ocean-ridge basalts. The modelling here illustrates that the megacrystic compositions (An >80) can only be produced from instantaneous melts from a depleted mantle source which, according to the melting

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column concept exists only at the top and crest of the melting region. Such depleted melts could also crystallise highly forsteritic olivine and FeO\*-poor endiopside.

If such refractory melts are only produced at the top of the melting column, then the polybaric crystallisation mechanism proposed by Flower (1980), Natland *et al.* (1983) and Griffin *et al.* (1983), in which refractory melts crystallise megacrystic plagioclase and endiopside at pressures of 10 kbars (~33 km depth) is incorrect. Accidental xenoliths may occur, the most convincing line of evidence being strain extinction which is observed for some megacrysts of pyroxene and olivine here and in the Mælifell eruption (Trønnes, 1990; Hansteen, 1991). It is doubtful that the exact location of the origin of these accidental xenoliths can be determined, especially in the absence of knowledge of the constituents of the lower crust beneath Iceland and oceanic spreading centres. Recently, however, depth of inclusion-entrapment in xenocrysts has been determined by investigating molar volumes of undisturbed  $CO_2$  inclusions within in olivine crystals of the Mælifell basalt. These investigations estimate entrapment between 7-10 km depth and at temperatures of 1230°C (Hansteen, 1991).

The research presented here supports a cognate origin for the majority of these megacrysts. The fact that the megacrysts are found in basalts erupted from fissures, and not in lava shield eruptions, suggests that a shallow reservoir not only promotes their appearance by accumulating them, but that it may be directly involved in their generation. The fact that they are not observed in lava shields casts doubts on suggestions made by Nicholson (1990) and Hansteen (1991) that they could be formed in the melt-rich top of the asthenosphere identified as a highly conductive layer under most of the Iceland plateau (e.g. Hersir et al., 1984). If this were the case, plagioclase megacrysts would be expected to occur randomly in both fissure- and lava shield-type eruptions, which, at least in the study area, is not observed. Also such a layer is not reported from other slow-spreading MORs and yet megacrysts exist in the basalts erupted there. Two units within the Hveragerði Volcanic System are similar in composition in terms of mantle melting signature but one contains megacrysts and the other does not. The one with megacrysts was erupted in the vicinity of the Grensdalur Central Volcano whereas the other was not. From the spatial distribution of the plagioclase megacrysts it is concluded that they form in crustal magma reservoirs, and in this study, one that is associated with a central volcano.

The mechanism by which the megacrysts form must allow them to maintain their uniform chemical composition and texture. It has been inferred that the glass

inclusions within plagioclase megacrysts indicate quench crystallisation of these crystals (Dungan & Rhodes, 1978; Flower, 1980; Natland et al., 1983). Others believe that they form by slow crystallisation entrapment (Griffin et al., 1983). In the study area, glass inclusions are observed in euhedral plagioclase phenocrysts in habits that do suggest rapid and incomplete growth of the crystal, but they are also found as inclusions in poikilitic intercumulus crystals that must have crystallized slowly. This suggests that the presence of glass inclusions does not constrain the rate of the mechanism of crystallisation for the megacrysts. The limited compositional range within the cores of the plagioclase megacrysts, however, lends support to a mechanism of rapid quench crystallisation. It has been proposed that high Mg # magmas enter a fractionated magma reservoir and undergo quench crystallisation (Stolper & D. Walker, 1980; Sparks et al., 1980; Huppert & Sparks, 1980) and that megacrystic plagioclase crystallises in this way (Kuo & Kirkpatrick, 1982). This proposal is supported by the evidence gathered in this research. This mechanism of quenching does not seem applicable to eruptive units that contain megacrysts that are in equilibrium with their host rock (e.g. S1 and Mælifell), unless the megacrysts within them are derived from the crystal mush of the reservoir, and not the host magma itself.

## Temporal variation

In the study area, temporal variation exists in the type of volcanism (fissure or lava shield), and in the abundance of megacrysts within the fissure eruptive units. It follows that the nature of the magmatic plumbing system is also highly variable in time (Figure 7.1). Within the stratigraphy of the Hveragerði Volcanic System, there are periods where the volcanism has been dominated by off-axis lava shields and highly megacrystic eruptive units that occur on-axis (Figure 3.2, p. 43; Figures 5.16 & 5.17, p. 132-133). According to the model in Figure 7.1, this suggests that there are periods when open, "free" crustal reservoirs do not exist. The periods dominated by lava shields and highly megacrystic eruptive units may coincide with low magma supply rates (Figure 7.1A). The Grensdalur crustal reservoir appears to have become "clogged-up" with crystal mush towards its extinction and this may either be the result of dwindling magma supply or a relative increase in the proportions of depleted instantaneous melt entering the chamber. It is envisaged that the depth of the crustal reservoir is likely to vary considerably in space and time.

In the Hengill Volcanic System, the main Hengill eruption event indicates that lava shield eruption events are capable of occurring on-axis, although most of them occur off-axis. According to the model (Figure 7.1), there cannot have been a crustal reservoir beneath the Hengill mountain area at the time of its eruption because the basalts contain no megacrysts, or evidence of mixing between magmas of diverse composition. Such features would be expected if the Hengill magmas had passed through a fractionated crustal reservoir of the sort that must have produced the older and more fractionated units in that region. These observations indicate that, in the study area, the processes that control the style of the magmatic plumbing system are highly variable with time, on the scale of 1 Ma. They may also be highly variable spatially. The most obvious process is the magma supply rate to the eruption, but there may be other factors controlling the form of the plumbing system.

### Iceland in relation to MORs

The rocks collected from the Hengill-Ingolfsfjall region display mineral assemblages and textures that are characteristic of rocks from elsewhere in Iceland (e.g. Carmichael, 1964; Jakobsson et al., 1978; Nicholson, 1990), and of MORB from the North Atlantic and other slow-spreading ridges (e.g. Aumento, 1968; Luyendyk, Cann et al., 1978; Batiza, 1982; Schilling et al., 1983; Stakes et al., 1984). Megacrystic plagioclase has also been reported in basalts from slow-spreading ridges such as the North Atlantic, the Costa Rica Rift (Natland et al., 1983), the Indian Ocean (Natland, 1991), but also from the EPR (Griffin et al., 1983) where they are less abundant. The research presented in this thesis suggests that the petrographic character of basalts is a direct result of processes operating on the magmas as they pass through the crust. This conclusion and the plumbing system model presented in Figure 7.1 are constrained by detailed field work and geophysical investigations into the structure of the crust which is not currently possible in such detail on the seafloor. If the same basalt textures are observed in rocks from the ocean spreading centres, then perhaps those basalts also resulted from similar processes to those that affected the Icelandic magmas. This may indicate that although the crust is slightly thinner off-shore, the basalts may pass through a similar plumbing system to that proposed in Figure 7.1.

The observations made from the EW9008 cruise indicate that the ridge is composed of volcanically constructed EVRs similar to the volcanic systems on the Reykjanes Peninsula. It has been suggest here that the seamounts may be phenomena similar to the Icelandic lava shields, whereas the more linear volcanic features on the seafloor represent material erupted from fissures. Recent cruises to the MAR using TOBI have confirmed that there are numerous seamounts of volcanic origin and similar in texture to those on the Reykjanes Ridge (Smith & Cann, 1990; Lawson *et al.*, 1991; Cann, 1992). These studies reveal that the neovolcanic zone is frequently dotted with seamounts that occur both on-axis as well as slightly away from it. Linear features have also reported from the FAMOUS (Ballard & vanAndel, 1977) and MARK areas (Brown & Karson, 1988). It seems then that the axial volcanic ridges on the MAR and the Reykjanes Ridge possess two types of volcanic morphology, conical and elongate which may be analogous to the lava shields and fissure eruption types in Iceland.

The most recent models proposed for the nature of crustal magma chambers at MORs (Smith & Cann, 1992; Detrick & Sinton, 1992) abandon older ideas of large reservoirs (e.g. Cann, 1974; Pallister & Hopson, 1981). However, these new models, illustrated in Figure 7.2, are in complete contrast to each other. Smith & Cann (1992) propose, on theoretical grounds, that the lower crust at slow-spreading MORs is composed of numerous small plutons, and that each one, in its life-time, had fed a single seamount eruption (Figure 7.2a). In this model the height of the seamount is proportional to the depth of the reservoir and is controlled by effusion rate of the volcanic material; an interpretation similar to that of G. P. L. Walker (1971) for the depth of origin of the Icelandic volcanic morphologies. In contrast, Detrick & Sinton (1992) postulate, after reviewing relevant geophysical and petrological literature, that a single reservoir exists from the upper 2 km of the crust down to the Moho (Figure 7.2b & c). The reservoir exists in the form of crystal mush with transition zones separating it from the isotropic gabbros of the crust. At fast-spreading ridges the chamber has a thin melt lens at 2 km depth (Figure 7.2c), whereas at slow-spreading ridges it does not (Figure 7.2b).

Aspects from each of these models are supported here and correspond with some of the constituents of the model in Figure 7.1. The plumbing system for seamount-dominated volcanism (Smith & Cann, 1992) is similar to the plumbing system proposed here for the lava shield volcanism in Iceland (Figures 5.21c & 7.1A). However, none of the evidence presented in this research supports the presence of isolated reservoirs situated in the lower crust. However, Húsmúli may indicate that such reservoirs exist. Basalts similar to those found in the lava shields are observed off-shore (e.g. Batiza, 1982; Langmuir *et al.*, 1977). Langmuir *et al.* (1977) propose that the olivine porphyritic basalts they sampled from the MAR were erupted straight from the base of the crust; a proposal that is supported here.

The plumbing-system models proposed by Detrick and Sinton (1992) are comparable in style to that proposed here for the fissure eruptions. The slowspreading model (Figure 7.2b) is comparable to that of the "clogged-up" reservoirs

(Figure 5.21b), and the fast-spreading model (Figure 7.2c) is comparable to that of the "free", open reservoir (Figure 5.21a). The difference between the models of Detrick & Sinton (1992) and that in Figure 5.21a & b lies in the vertical extent of the chamber. The Mælifell eruptive unit, however, suggests that deeper varieties of crustal chambers may exist (Hansteen, 1991). From Figures 1.5a (p. 10) and 3.11 (p. 63), it is noted that the Mælifell, and other units of the Hrómundartindur Volcanic System, do not form well-defined linear topographic features. This is in accordance with a reservoir at deeper levels in the crust. There will be a critical reservoir depth at which the effusion rate will change from that which forms lava shields to that which forms fissure eruptions.





Figure 7.2: Recently proposed models for the crustal structure and magmatic plumbing system at constructive plate boundaries: a) from Smith & Cann (1992) for slow-spreading MORs; b) & c) are from Detrick & Sinton (1992) for slow- and fast-spreading MORs, respectively.

#### Synthesis and Conclusions

The effect of higher magma supply rates in Iceland may result in shallow crustal reservoirs like those proposed for the fast-spreading ridges (Figure 7.2c) except the melt lens at the top of the crystal mush may only be present periodically. The samples from the Reykjanes Ridge examined here also require a "free" style of magma chamber (Figures 5.21a & 7.2c).

There is no consistent relationship between the varying states of the EVRs and their distance from the Icelandic plume. This lack of variation suggests that the formation of the tectonic and volcanic features is independent of the processes operating in the mantle associated with the plume (higher mantle temperatures), bot depends on crustal processes. The control on the development of the volcanic morphologies and the tectonic state of the EVR is likely to be the magma flux. When the magma budget to one part of the ridge is low, extension is taken up by normal faulting perpendicular to the spreading direction; when budget is intermediate, seamounts volcanism dominates; and when the budget is high, the extension is taken up by dyking and associated fissure volcanic activity. Despite the predictions made above, the initial studies of twelve dredged basalts from the Reykjanes Ridge (Section 6.3.2.3) do not appear to possess any systematic relationship between the morphology of the volcanic edifice and the type of basalt erupted there. Other studies are underway investigating the petrology of basalts dredged from the various volcanic morphologies on the MAR (K. Lawson pers. comms., 1992).

There are several aspects of the magmatism on Iceland and at MORs that need to be evaluated in light of these recent models and the model derived from this research. The crystal mush in the model presented here is dominated by megacrysts that crystallised from the depleted top of the melting column. If such a "clogged" reservoir were to extend to the base of the crust, then a large proportion of the crust must be made up of depleted instantaneous melts. In view of current melting column models, it is unlikely that the more voluminous accumulated melt only forms the upper 2 km of the crust. However, the abundance of the megacrysts within the crust is supported by the compositions of feldspars analysed in the xenoliths from Mælifell (Trønnes, 1990; Hansteen, 1991) and from unit S12 of the present study. At oceanic spreading centres, Donaldson & Brown (1977) report that plagioclase phenocrysts are more abundant than plagioclase megacrysts within the porphyritic basalts. These observations are in accordance with larger degrees of partial melting of the mantle at hotspots (Dick *et al.*, 1984; McKenzie, 1984).

One of the main differences between the axial volcanic ridges off-shore and the volcanic systems in Iceland is the central volcanoes which have not been reported from off-shore. The effect of these on the petrology of the basalts could be assessed by comparing the basalts in the Hengill area with those on the Reykjanes Peninsula where no central volcanoes have been reported. Jakobsson *et al.* (1978) report large plagioclase phenocrysts (8 mm) with highly calcic compositions (up to An 88%), but they do not describe them as megacrysts. The central volcanoes on Iceland may represent an over-development of crustal reservoirs, like that in Figure 7.2b or c, thought to exist off-shore.

It is proposed here that the Hengill mountain represents a lava shields type eruption in the middle of the swarm and not the Hengill central volcano. However, the radial component of stress is centred on the mountain (Figure 1.5a, p. 10). The low-velocity volume that may represent a crustal reservoir is situated to the north of the Hengill mountain, and could be the site of the Hengill central volcano. However, the only occurrence of rhyolites in the system is to the south of the mountain. From these observations, it is concluded here that it is not feasible to encircle a well-defined central volcano within the Hengill Volcanic System.

## 7.2) Conclusions

 $\star$  Lava shield eruption events represent magmas that ascend from deep, probably at the crust/mantle boundary. Fissure eruptions result from magmas erupted from shallow crustal reservoirs.

The magmatic plumbing system is highly variable with time, even on the scale of 1 Ma. The higher magma supply rates in Iceland result in the periodic occurrence of melt-dominated shallow crustal reservoirs, similar to those at fast-spreading ridges.

Finite Analysis and Analysis a

Flagioclase megacrysts accumulate in shallow crustal reservoirs and form flotation cumulates. Associated olivine and pyroxene are also flotation cumulates. Plagioclase megacrysts crystallise from depleted instantaneous melts by quench crystallisation when such melts encounter more fractionated ones in a crustal reservoir.

Seamounts and elongate features on the submerged MORs may be equivalent to Icelandic lava shields and fissure eruptions, respectively. An initial investigation on twelve dredged basalts from  $63^{\circ}10$ 'N on the Reykjanes Ridge indicates that there is no relationship between petrology and bathymetry at this location.

★ The Hengill Mountain does not represent the Hengill Central Volcano, but rather an on-axis, subglacial lava shield.

Five stratigraphic groups are identified in the Hveragerði Volcanic System, separated by six marker horizons. The locus of activity can be traced over a 700,000 year period, during which time volcanism has randomly alternated between fissureand lava shield-type eruptive events. There has been a general migration of volcanism to the west.

# 7.3) Future Work

- The magmatic plumbing system model presented here could be tested by further targeted geophysical studies in Iceland (gravity and tomography). Such studies on the Reykjanes Peninsula, where central volcanoes do not exist, would provide a comparative study to the Hengill region in order investigate the density structure of the crust with and without central volcanoes.

- According to the model presented here, a thicker crust should enhance features such as the dome shape of the shields and the development of the radiating feldspar texture, and further work could test this hypothesis.

- Sampling of the neovolcanic zone (WVZ and NVZ) within individual lava shields (vertical profiles) would give a better indication of the overall petrological diversity. Such studies should be coupled with analyses of the lava shield shapes.

- Within the Hveragerði region, there exists a discrepancy between the age of the DB7b lava shield as defined by K/Ar methods and that derived by field mapping. Geochemical analyses on the lavas in this Núpafell-Kviar region may be able to solve this discrepancy and constrain the stratigraphy there.

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Chapter 7

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## Appendix 1 - A Stratigraphic logs

During fieldwork, a number of stratigraphic logs were constructed in order to establish the stratigraphy and to record sample locations accurately within lava profiles. Figure A1.1 is a map of the study region with the location of the logs detailed in this appendix. The logs have been grouped into regions, from west to east, and are presented in this order according to the list below:

Hengill region	- Hæðir					
	- Húsmúli					
	- Núpafell					
Grensdalur region	- Djúpagil					
	- Klóarfjall					
	- Selfjall					
	- Reykir Profile; - R1 to R8 (see Appendix 1 - B)					
Gljúfur region	- G1 to G4					
Ingolfsfjall region	- IF					
	- IL					
	- IU					
	- IX					
	- IS					

Within each log the following are recorded: lithology (as described in Chapter 2); stratigraphic markers, groups and eruptive units (as described in Chapter 3); sampled horizons (volcanic glass sample numbers are in brackets, whereas whole-rock samples are not); the petrological type of basalt (as defined in Chapter 4). The key below is for the lithological type, as illustrated in the logs. Abbreviations include ILS = Ingolfsfjall lava shield; LK = Leo Kristjánsson *et al.*, (1988). The vertical scale on each log is in meters

Key



HB (hyaloclastite breccia)

tuffs

lava

pillow lavas





Figure A1.1: Map of the study area showing the location of the logged sections detailed in this appendix.



Region: Hengill Log title: Húsmúli





Region: Hengill Log title: Núpafell

<sup>&</sup>lt;sup>1</sup>Unit DB7b - samples from sub-units 1 and 2 were collected from the cliff that faces eastwards (illustrated in Figure 3.10), whereas samples from sub-unit 3 were collected from a small gorge behind this cliff, that faces SW. The sample from unit DB8b was collected from the top of the profile.

	Lithology	Lithological description	Marker	Stratigraphic Group	Stratigraphic Unit	Sample	Petrological type	Other
250 -		lava		Hengill	Bitra		2	trom Hengill Volcanic System
,		i de tra			DB9	A292		
200-		sediments				A291 A290		possibly a different unit to DB8a
•		lava		DB	DB8a		2ь	
		HB				A289 A288		
		HB	1		DB4	<u>↓</u>	2a	_
150 -		sediments .				A287		these may belong to DB7b
	-	lava	DA1	DA	DA1		3a	pers. comm. 1992 )

A286

Region: Grensdalur Log title: Djúpagil Region: Grensdalur Log title: Klóarfjall Stratigraphic Group Stratigraphic Unit Petrological type Lithological description Lithology Marker Other tuffs S16 2c 400 HB S15 2b s ି pillow lava sediments 0 õ C S12 (\$12) (2C) 00 S1 S1 350 lava coarse grained basalt with ophitic texture DB10c 2a 300 DB DB10b HB & feeder sheets 250 coarse grained basalt with ophitic texture DB10a altered basalt outcrop TA TA2 1a 200 (HB)





Log title: Reykir Profile - R1 Stratigraphic Group Stratigraphic Unit Petrological type Lithological description Lithology Sample Marker Other HB S5 2b pillow lava Ó A413 A 415 A 414 lava A.416 S sediments Compositionally 250 A409 S1 S1 distinct lavas A410 lava A408 A151 2a sediments **A4**11 lava TB6 A150 sediments ΤВ A417 200 HB твз 2b A418 no exposure 150. A70 **A6**9 HB+ 100 таз feeder TA 1a HΒ A73 TA1

**Region: Grensdalur** 

Region: Grensdalur Log title: Reykir Profile - R2



Region: Grensdalur Log title: Reykir Profile - R3

	Lithology	Lithological description	Markers	Stratigraphic Group	Stratigraphic Unit	Petrological type	Other	
250 -		НВ		S	S5	2b		
-		tuffs			S4	1a		
230 -		HB		тв	TB3	2b		



# Region: Grensdalur Log title: Reykir Profile - R4

Appendix 1 - A

	Lithology	Lithological description	Marker	Stratigraphic Group	Stratigraphic Unit	Sample	Petrological type	Other
				S	<b>S</b> 3		1a	aphyric HB
-		HB tuffs			тв6		2c	
150 —		нв						
		tillite	т2	TB				moderately phyric
		HB <sub>lava</sub>			BA3 TB3		3A2B	
-		tuffs			_TB2		1	aphyric tuffs
		distal HB			TB1		2b	phyric HB
-		sediments	<u>T1</u>					megacrysts
		lava			TA10	A299 A121	1a	Sparsely phyric
- - 100		sediments		TA				aphyric sediments with smectite alteration

Region: Grensdalur Log title: Reykir Profile - R8

,
Appendix 1 - A



**Region:** Gljúfur Log title: G2



				200				300 -		Reg Log
									Lithology	ion: Gljúfur title: G3
	sediments	HB	tuffs	sediments	pillow lava	Ŧ	HB	H	Lithologica descriptior	ll N
	IM1?								Markers	
				88	}			BA	Stratigraph Group	nic ,
		BB3				BB4	BA2	BA3	Stratigrapł Unit	nic
			_	30			1a	323	Petrologica type	N.
									Other	
ľ										

,

Region: Gljúfur Log title: G4

			200 -		3	
· · · · · · · · · · · · · · · · · · ·	0 0 0 0 0 0	0000				Lithology
tillite	pillow lava	HB	B	НВ	HB	Lithological description
?INI						Markers
		<u>,</u>	B	BA	ILS	Stratigraphic Group
		BB3	BB4	BA3		Stratigraphic Unit
			36	န္မ		Petrological type
						Other

۰.





Appendix 1 - A



Region: Ingolfsfjall Log title: IL



Log title: IU 2 Stratigraphic Group Stratigraphic Unit Petrological type Lithological description Lithology Marker Sample Other HΒ ILS 3a A522 0 O 0 pillow lava 150 tillite IM2 **A5**0 LK = No 5A49 A48 LK = No 4 lava BB6 3ь BB LK = No 3A47 sediments lava BB5 LK = No 2**A4**6 1a sediments. 700 ka lava GB6a A45 **2**c LK = No 1bedded sediments IM1 100-GB A551 3a GB4 HB (A44)

<sup>2</sup> LK = Léo Kristjánsson *et al.*, (1988), and refers to lava numbers presented in the profiles therein. These numbers were read from the outcrop.

Region: Ingolfsfjall



 ${}^{3}LK = Léo Kristjánsson$ *et al.*, (1988), and refers to lava numbers presented in the profiles therein. These numbers were read from the outcrop.

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Appendix 1 - A

Region: Ingolfsfjall Log title: IS



## Appendix 1 - B Reykir Cliff Profile

The Reykir cliff occurs in the Hveragerði Volcanic System at the SE corner of the Grensdalur Central Volcano. It is 1.5 km long and varies in absolute height from 100 to 270 m, from south to north respectively. The eruptive units and sediments exposed there have been deposited over a period of at least 350 ka (four separate tillite horizons). The stratigraphy has been correlated from eight logged sections, the details of which are provided in Appendix 1A, and the interpretations of which are summarised below.



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Appendix 1 - B

Appendix 1 - B



Figure 1B.1: The Reykir Cliff Profile: a) view of the Reykir cliff face with the location of the sections and the interpretation of the geology indicated; b) eight logs (R1 to R8) of the stratigraphy of the cliff face.

Appendix 1 - B

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## Appendix 1 - C Compositional Evidence for the S1 Marker Horizon and Other Units

During interglacial periods, subaerial compound aa lavas were erupted in the Grensdalur region. They represent repeated eruptions of the fissure type. The lavas may be grouped or separated as eruptive units by their degree of continuity in the field, and distinguishing features, such as the basalt type (i.e. phenocryst content and groundmass texture). The composition of the basalts provides an additional line of evidence with which to correlate these lavas.

In Figure 1C.1, Nb is plotted against Ni for some of the lavas in the region of the Grensdalur volcano. For comparison, other eruptive units that occur adjacent to them are also plotted (e.g. DA3, S5, S15). The Stekkás basalts, reported to be from the Hrómundartindur Volcanic System, are also included. This figure shows that the samples within a given eruptive unit are all similar and, although there is some overlap, each units is compositionally distinct. The following conclusions can be drawn from this figure.

Unit S1: The S1 marker lavas are separated from the other units and are very distinct. They are different from the DB10 samples, with which they appear to be laterally continuous in the field; and from the TB6 lavas that they overlie in the Reykir profile (R1; Appendix 1A) (Section 3.3.1).

Stekkás: The Stekkás basalts are similar to the DB10 basalts, and it is suggested here that they may be of the same eruptive unit.

Unit TB1 and TB6: Unit TB6 overlays unit TB1 on the southern slopes of Reykjafjall (Map A2), and the basalt of each unit is very similar in handspecimen; however, they are compositionally distinct.

Unit DB10: This unit is divided into three sub-units on a textural basis (Section 3.3.3), and yet the samples from these sub-units are compositionally indistinguishable.

Appendix 1 - C



Figure 1C.1: Plot of Nb against Ni for selected eruptive units from the Grensdalur Central Volcano, and the Stekkás basalts from the Hrómundartindur Volcanic System.

Appendix 1 - D Compositional Evidence for Unit GB6



Figure 1D.1: A plot of Nb and Ni for samples from the Ingolfsfjall region, below the main Ingolfsfjall lava shield eruptive unit. The plot illustrates how unit GB6 can be split into two (A and B) on compositional grounds. The other units from profiles IU and IX (Appendix 1A) are plotted for comparison. The table below provides sample numbers as detailed here, and additionally, the profile and lava numbers of Léo Kristjánsson et al., (1988), as read directly from the outcrop. All the eruptive units in these profiles above the GB6 lavas are of normal polarity; these lavas therefore mark the 700 ka boundary.





Figure 1D.2: The south face of Ingolfsfjall viewed from Selfoss town: a) photograph; b) line drawing of (a). The location of the samples from unit GB6 are indicated, as are the locations of the profiles IS IX and IU. The profile numbers in brackets are those of Léo Kristjánsson et al., (1988). The small peninsula composed of unit GB4 and the IM1 marker is at the extreme west of the view.

## Appendix 1 - E Postglacial lava from the Grensdalur Central Volcano?

Within the Region occupied by the Grensdalur Central Volcano, near the Tindar high, there are scattered outcrops of basalt that has the texture similar to Postglacial basalt seen elsewhere in the Hengill region. The location of these blocks is indicated in Figure 1AE.1. The basalt blocks are referred to as unit TC3. The scattered blocks of rock are situated on high groun where they protrude through, or are positioned on the clay scree. In thin section (Figure 1AE.2) the rock is completely replaced by hematite, which is reflected in its bulk composition (Table 1AE.1). Vesicles, which display flow texture, suggest that the rock is eruptive material and not a primary alteration product. The location of this material suggests that it is from the Grensdalur Central Volcano, as it is inconceivable that other environmental elements during the Postglacial era could have transported the blocks to their present location. The most abundant occurrence of blocks is on a gentle slope towards the Grensdalur river (Sæmundsson, pers comm., 1990).



Figure 1AE.2: Location of the outcrops composed of the altered Postglacial material.



Figure 1AE.2: Photomicrograph of the material from the Grensdalur Central Volcano that may represent altered Postglacial lava. (Field of viwe is 15 mm).

Sample no	A557
SiO2	21.88
AI2O3	-4.75
FeO (t)	84.16
MgO	4.33
CaO	0.11
Na2O	0.27
K20	-0.01
TiO2	2.22
MnO	0.01
P2O5	0.37
Nb	20.16
Zr	152.48
Y	2.74
Rb	48.05
Sr	46.39
Zn	21.71
Ni	-8.28
V	1935.05
Cr	197.05
Ga	23.49
Sc	14.43
Co	258.72
Total	109.04

Table 1AE.1: Major oxide and trace element compositioin of the material from the Grensdalun
Central Volcano that may represent altered Postglacial lava.

# Appendix 2 - A Feldspar Mineral Compositions

This appendix contains the feldspar mineral compositions from samples within the Hengill-Ingolfsfjall region. Analyses were carried out on the electron microprobe (at the Universities of Durham and Manchester). The analyses are presented in this appendix according to the list below:

Volcanic System	Unit	Sample	Page	Мар
Hveragerði	GB4	<b>A</b> 44	268	<b>B3-4</b>
	BA2	A523	268	<b>B</b> 1-7
	BA3	A122	268	С
	BB5	A545	268	B3-3
	BB7	A36	269	B3-8
	DB7b	A474	269	B1-4
	<b>S</b> 1	A416	269	B2-8
	<b>S</b> 12	A495	270	B2-6
	DB10c	A349	271	B2-5
	ILS	A9	271	С
Hengill	Hæðir	A512	273	<b>B</b> 1-1
	Húsmúli	A502	273	<b>B</b> 1-3
	Nesjaskógur (Nes)	A383	274	<b>B</b> 1-1
	Hahryggur (Há)	A174	274	<b>B</b> 1-1
	Bæjarháls (Bæ)	A368	274	B1-5
	Hengill	A374	275	B1-2

#### Key to Appendix 2 - A:

Sample #	= sample number
Eruptive unit	= refers to the position in the stratigraphy (Figure 3.2 [p.43] & 3.12 [p. 64])
Crystal #	= 1, 2, 3, etc is a means of differentiating between individual crystals
Туре	= Intratelluric (I) or Quench (Q)
Variety	= textural variety of intratelluric crystal, as defined on p.
Spot position	= location of analyses within the crystal
	-C = core
	-I = between the rim and the core
	$-\mathbf{R} = \mathbf{rim}$
Spot reference	= analyses number within sample
Partial total	= total of cations present in the crystal

Hverageröi Volcanic System

Sample #	A44	<b>A44</b>	<b>A44</b>	<b>A44</b>	A523	A523	A523	A523	A523	A523	A122	A122	A122	A122	A122	A545	A545
Eruptive unit	GB4	GB4	GB4	GB4	BA2	BA2	BA2	BA2	BA2	BA2	BA3	BA3	BA3	BA3	BA3	885	BB5
Crystal #	1	1	1	1	1	1	1	2	2	2	1	1	1	2	3	1	1
Туре	L	I.	I	1	1	1	I	Q	Q	Q	1	1	I	1	I.	Q	Q
Variety	Т	T	т	т	т	Т	т				т	т	т	т	т		
Spot position	С	С	С	С	С	С	С	С	С	С	I	t	ŧ	С	С	С	С
Spot reference	A44F1	A44F2	A44FI1	A44F5	A523F1	A523F2	A523F3	A523F4	A523F5	A523F6	A122F1	A122F2	A122F3	A122F4	A122F5	A545F1	A545F2
Si02	47.21	47.68	46.95	48.64	48.36	48.55	47.55	50.51	51.57	51.06	48.46	48.16	48.42	47.86	48.65	54.21	54.01
TiO2	0.02	0.12	0.06	0.10	0.16	0.11	0.00	0.07	0.07	0.16	0.03	0.02	0.02	0.06	0.00	0.09	0.22
A1203	31.39	30.42	31.02	30.84	30.02	30.32	31.05	27.95	28.77	28.43	30.79	31.12	30.05	30.54	31.57	26.05	26.48
Cr203	0.23	0.20	0.16	0.22	0.11	0.08	0.01	0.14	0.07	0.26	0.00	0.15	0.07	0.06	0.15	0.28	0.22
FeO	0.92	0.82	0.68	1.41	0.84	0.73	0.92	1.25	1.34	1.09	0.87	0.74	1.50	0.97	1.16	1.18	1.05
MnO	1.18	1.08	1.17	1.16	0.92	0.77	0.95	1.01	0.76	0.82	0.81	0.87	0.67	0.68	0.65	0.98	1.05
MgO	0.72	0.55	0.78	1.09	0.25	0.22	0.15	0.59	0.64	0.63	0.75	0.44	1.49	0.38	0.95	0.22	0.33
CeO	17.15	16. <b>6</b> 6	16.65	16.51	15.46	15.30	15.77	13.62	13.77	13.76	16.14	16.52	15.30	15.83	16.33	10. <b>09</b>	10.61
Na20	1.62	1.85	1.73	2.09	2.46	2.70	2.21	3.47	3.51	3.36	2.21	2.38	2.34	2.09	1. <b>9</b> 5	5.04	5.05
K20	0.01	0.18	0.00	0.02	0.03	0.05	0.01	0.05	0.03	0.08	0.00	0.05	0.05	0.07	0.00	0.38	0.32
Total	100.43	99.53	99.18	102.06	98.61	98.82	98.62	98.64	100.52	99.64	100.05	100.44	99.91	98.53	101.3 <del>9</del>	98.52	<b>99.35</b>
Atoms per 8 oxys	jens																
SI	2.23	2.27	2.24	2.26	2.31	2.31	2.28	2.41	2.41	2.40	2.28	2.27	2.29	2.29	2.26	2.56	2.53
AI	1.75	1.71	1.74	1.69	1.69	1.70	1.75	1.57	1.58	1.58	1.71	1.73	1.67	1.72	1.73	1.45	1.46
Ca	0.87	0.85	0.85	0.82	0.79	0.78	0.81	0.69	0.69	0.69	0.81	0.83	0.78	0.81	0.81	0.51	0.53
Na	0.15	0.17	0.16	0.19	0.23	0.25	0.21	0.32	0.32	0.31	0.20	0.22	0.21	0.19	0.18	0.46	0.46
Partial totals	4.99	5.00	5.00	4.96	5.02	5.04	5.04	4.99	5.00	4.99	5.01	5.04	4.95	5.01	4.9 <del>9</del>	4.99	<b>4.99</b>
An%	85.39	83.31	84.12	81.40	77.67	75.79	79.70	68.46	68.47	69.39	80.16	79.38	78.37	80.73	82.26	52.52	53.77

Appendix 2 - A

Sample #	A545	A36	A36	A36	A36	A474	A474	A416	A416	A416	A416	A416	A416	A416	A416	A416	A416
Eruptive unit	<b>B85</b>	887	887	887	BB7	D876	DB7b	<b>S</b> 1	<b>S</b> 1	St	<b>S</b> 1	S1	<b>S</b> 1	<b>S</b> 1	<b>S</b> 1	<b>S1</b>	<b>S</b> 1
Crystal #	1	1	1	2	2	1	1	1	1	1	1	1	2	2	3	4	5
Туре	Q	I.	I.	Q	Q	Q	Q	1	E E	ł	I.	ł	1	I.	1	I.	1
Variety		Е	Е					M	М	м	м	Μ	м	М	Е	М	E
Spot position	С	С	С	С	С	С	С	С	С	С	R	R	С	R	С	С	С
Spot reference	A545F3	A36F1	A36F2	A36F3	A36F4	A474F1	A474F2	A416F1	A416F2	A416F3	A416F4	A416F5	A416F6	A416F7	A416F8	A416F9	A416F10
SI02	52.53	54.95	54.43	53.33	52.98	48.69	48.02	47.00	46.96	48.27	48.04	45.91	45.68	45.40	45.74	46.81	44.90
TIO2	0.15	0.22	0.05	0.28	0.15	0.00	0.11	0.03	0.01	0.00	0.00	0.00	0.08	0.02	0.05	0.08	0.10
AI203	<b>26.19</b>	27.54	27.25	<b>26</b> .77	26.87	31.16	31.14	32.64	32.39	32.87	33.64	33.54	33.40	33.28	33.43	32.51	33.37
Cr203	0.13	0.23	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.11	0.00	0.09	0.00	0.10	0.00	0.09	0.04
FeO	1.41	0.95	1.09	1.13	1.04	1.25	1.19	0.65	0.42	0.48	0.29	0.60	0.46	0.65	0.26	0.45	0.60
MnO	0.98	0.16	0.22	0.39	0.30	0.12	0.10	0.06	0.11	0.05	0.00	0.04	0.02	0.11	0.15	0.00	0.21
MgO	0.28	0.52	0.55	0.41	0.44	1.00	1.19	0.70	0.81	0.62	0.74	0.77	0.61	0.63	0.81	0.81	0. <b>79</b>
CeO	10.94	10.88	10.91	10. <del>9</del> 5	11.10	15.57	15.44	16.42	16.49	17.27	17.53	17.68	17.40	17.69	17.28	16.96	17.65
Na20	4.68	5.08	5.15	4.88	4.62	2.44	2.48	1.76	1.99	1.50	1.46	1.40	1.37	1.22	1.51	1.68	1.33
K20	0.25	0.14	0.17	0.21	0.24	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	97.53	100.66	99.83	98.34	97.72	100.22	<b>99.69</b>	99.27	99.21	99.17	99.71	100.00	99.02	<b>99</b> .10	99.21	99.39	98.98
Atoma per 8 oxyg	ens																
SI	2.52	2.48	2.48	2.47	2.46	2.24	2.22	2.18	2.18	2.15	2.13	2.12	2.13	2.12	2.13	2.17	2.10
AI	1.48	1.46	1.46	1.46	1.47	1.69	1.90	1.78	1.77	1.80	1.83	1.83	1. <b>83</b>	1.83	1.83	1.78	1.84
Ca	0.56	0.53	0.53	0.54	0.55	0.77	0.76	0.82	0.82	0.86	0.87	0.87	0.87	0.88	0.86	0.84	0. <b>89</b>
Na	0.43	0.44	0.46	0.44	0.42	0.22	0.22	0.18	0.18	0.14	0.13	0.13	0.12	0.11	0.14	0.15	0.12
Partial totals	4.99	4.91	4.92	4.91	4.91	4.91	5.10	4.93	4.95	4.95	4.96	4.95	4.95	4.94	4.95	4.94	4.95
An%	56.36	54.18	53.90	55.35	57.01	77.92	77.48	83.76	82.08	86.43	86.89	87.49	87.50	88.93	86.35	84.79	88.06

Sample #	A495	A495	A495	A495	A495	A495	A495	A495	A495	A495	A495	A495	A495	A495	A495	A495	A495
Eruptive unit	S12	S12	S12	S12	<b>S12</b>	S12	S12	S12	S12	S12	S12	S12	S12	S12	S12	S12	S12
Crystal #	5	5	5	2	2	2	3	3	3	3	1	1	1	1	1	1	4
Туре	Q	Q	Q	I	I	I	I	1	I	I	I	I.	I	I I	I	1	I
Variety				Е	Е	E	т	т	т	т	м	м	м	м	м	м	м
Spot position	С	Ċ	С	· C	С	С	С	С	С	С	С	С	С	R	R	R	С
Spot reference	A495F17	A495F18	A495F19	A495F7	A495F8	A495F9	A495F10	A495F11	A495F12	A495F13	A495F1	A495F2	A495F3	A495F4	A495F5	A495F6	A495F14
SIO2	47.27	<b>45.60</b>	45.53	49.12	48.35	50.09	47.18	50.28	50.05	50.06	47.46	47.07	46.35	50.23	49.54	48.77	43.82
T102	0.23	0.21	0.23	0.20	0.07	0.06	0.00	0.21	0.22	0.09	0.09	0.03	0.00	0.08	0.12	0.03	0.05
A1203	25.56	23.83	24.25	29.18	28.47	28.88	25.07	27.55	27.04	25.11	31.13	30.85	31.02	29.14	28.64	29.07	29.67
Cr203	0.14	0.18	0.05	0.09	0.15	0.10	0.11	0.07	0.15	0.02	0.06	0.16	0.16	0.16	0.06	0.16	0.17
FeO	1.35	1.56	1.48	1.18	0.90	0.98	1.01	0.84	0.75	1.80	0.76	0.78	0.75	0.97	0.97	0.86	0.46
MnO	1.24	0.81	0.92	1.11	1.17	0.98	0.85	1.27	1.13	1.40	0.93	1.12	1.02	1.04	1.16	1.10	1.20
MgO	0.44	0.44	0.72	0.27	0.18	0.25	0.33	0.1 <del>9</del>	0.33	0.56	0.11	0.06	0.00	0.19	0.11	0.14	0.36
CeO	12.01	11.03	11. <b>64</b>	14.18	14.04	13.50	11.15	12.38	11.99	11.37	16.16	16.20	16.28	13.70	13.78	14.25	14.97
Na20	3.5 <del>6</del>	3.87	3.46	3.27	3.38	3.46	3.56	3.82	4.01	4.38	2.15	2.20	1.96	3.38	3.30	3.14	1.78
K20	0.16	0.19	0.19	0.03	0.12	0.04	0.12	0.11	0.06	0.14	0.03	0.01	0.07	0.09	0.18	0.06	0.04
Total	91.95	87.72	88.47	98.62	96.82	<b>98.3</b> 5	89.38	<b>96.7</b> 1	<del>9</del> 5.72	94.91	98.89	98.48	97.60	<b>98</b> .97	97.86	97.58	92.52
Atoma per 8 oxyg	ena																
Sł	2.42	2.45	2.42	2.35	2.36	2.39	2.47	2.43	2.44	2.49	2.26	2.26	2.25	2.38	2.38	2.35	2.24
AI	1.54	1.51	1.52	1.64	1.63	1.62	1.54	1.57	1.56	1.47	1.75	1.75	1.77	1.63	1.62	1.65	1.78
Ca	0.66	0.63	0.66	0.73	0.73	0.69	0.62	0.64	0.63	0.60	0.83	0.83	0.85	0.70	0.71	0.74	0.82
Na	0.35	0.40	0.36	0.30	0.32	0.32	0.36	0.36	0.38	0.42	0.20	0.20	0.18	0.31	0.31	0.29	0.18
Partial totals	4.98	4.99	4.97	5.02	5.04	5.02	5.00	5.01	5.01	4.98	5.04	5.04	5.05	5.02	5.02	5.04	5.01
An%	65.05	61.15	65.02	70.60	69.69	68.29	63.35	64.17	62.25	59.04	80.61	80.31	82.14	69.15	69.75	71.47	82.33

Appendix 2 - A

Sample #	A495	A495	A495	A495	Xenolith	Xenolith	Xenolith	A349	A349	A349	A349	A349	A349	A349	A9	A9
Eruptive unit	S12	S12	<b>S12</b>	<b>S12</b>	S12	<b>S12</b>	S12	DB10c	DB10c	DB10c	DB10c	DB10c	DB10c	DB10c	ILS	ILS
Crystal #	4	4	6	6	1	1	1	1	1	1	1	2	2	3	1a	1b
Туре	1	I	I	1	1	1	1	1	1	ł	1	1	1	1	1	1
Variety	M	м	м	м	М	м	м	М	М	м	м	М	M	м	т	т
Spot position	С	с	с	R	С	С	R	I	1	ł	Ç	С	С	С	С	С
Spot reference	A495F15	A495F16	A495F20	A495F21	<b>X</b> 1	X2	X3	A349F1	A349F3	A349F4	A349F5	A349F6	A349F7	A349F8	A9F2	A9F1
SIO2	45.81	45.28	45.51	44.89	<b>45.99</b>	46.25	45.51	47.50	45.86	46.98	45.40	46.88	45.98	46.68	47.04	47.47
TIO2	0.15	0.23	0.20	0.06	0.00	0.81	0.14	0.20	0.00	0.00	0.00	0.00	0.00	0.07	0.02	0.00
A1203	30.54	30.56	30.97	30.15	33.67	27.16	32.56	32.59	33.44	32.49	32.50	31.72	31.89	33.72	31.68	31.77
Cr2O3	0.22	0.02	0.12	0.17	0.00	0.13	0.00	0.07	0.07	0.09	0.00	0.06	0.13	0.00	0.17	80.0
FeO	0.63	0.50	0.65	0.67	0.37	4.02	0.48	0.37	0.57	1.02	0.75	0.55	0.47	0.69	0.90	1.05
MnO	0.62	0.99	0.79	1.65	0.00	0.43	0.00	0.23	0.20	0.20	0.33	0.20	0.28	0.20	1.14	1.31
MgO	0.23	0.17	0.30	0.29	0.31	2.39	0.44	0.62	0.45	0.71	0.36	0.57	0.52	0.66	0.55	0.47
CaO	15.58	15.52	16.56	15.93	18.12	16.05	17.37	16.76	17.60	17.13	17.08	16.71	16.31	17.75	16.70	16.43
Na20	1. <b>89</b>	1.55	1.72	1.78	1.65	1.54	1.74	2.00	1.33	1.70	1.44	2.05	1.88	1.56	1.89	2.06
K20	0.02	0.04	0.02	0.06	0.01	0.10	0.04	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	95.68	94.87	96.64	95.65	100.13	98.50	98.28	100.33	99.51	100.32	97.87	<b>98</b> .75	97.46	101.33	100.09	100.65
Atoms per 8 oxyge	ns															
SI	2.26	2.25	2.22	2.23	2.12	2.21	2.14	2.18	2.13	217	2.14	2.19	2.17	2.13	2.23	2.24
Al	1.77	1.79	1.78	1.76	1.83	1.53	1.80	1.76	1.83	1.76	1.81	1.75	1,78	1.81	1.77	1.76
Ca	0.82	0.82	0.87	0.85	0.90	0.82	0.87	0.82	0.88	0.85	0.86	0.84	0.83	0.87	0.85	0.83
Na	0.18	0.15	0.16	0.17	0.15	0.14	0.16	0.18	0.12	0.15	0.13	0.19	0.17	0.14	0.17	0.19
Partial totals	5.03	5.01	5.04	5.01	5.03	5.07	5.03	4.95	4.95	4.93	4.95	4.96	4.95	4.95	5.02	5.02
An%	82.04	84.62	84.13	83.16	85.83	85.20	84.65	82.24	88.03	84.77	86.75	81.80	82.77	86.28	83.04	81.50

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Sample #	A9	A9	A9	A9	<b>A</b> 9
Eruptive unit	ILS	ILS	ILS	ILS	ILS
Crystal #	2	2	2	3	3
Туре	1	I.	1	Q	Q
Variety	т	т	т		
Spot position	С	С	С	С	С
Spot reference	A9F4	A9F5	A9F6	A9F7	A9F8
SIO2	46.79	46.81	47.51	49.51	49.52
TIO2	0.11	0.10	0.10	0.56	0.27
AI203	30.76	31.56	31.70	26.37	28.28
Cr203	0.00	0.00	0.08	0.24	0.09
FeO	0.86	0.80	0.75	2.53	1.57
MnO	1.19	1.12	1.11	1.26	1.14
MgO	0.39	0.40	0.27	1.80	0.82
CaO	16.26	16.74	16.46	13.57	14.10
Na20	2.04	1.88	1.85	3.13	3.20
K20	0.00	0.00	0.00	0.0 <del>9</del>	0.00
Total	98.40	99.41	99.83	99.04	99.00
Atoms per 8 oxyge	<b>ms</b>				
SI	2.25	2.23	2.25	2.37	2.36
AI	1.75	1.77	1.77	1.49	1.59
Ca	0.84	0.85	0.83	0.70	0.72
Ne	0.19	0.17	0.17	0.29	0.30
Partial totals	5.03	5.03	5.02	4.86	4.97
An%	81.48	83.05	83.04	70.52	70.91

Sample #	A512	A512	A512	A502	A502	A502	A502	A502	A502	A502	A502	A502	A502	A502	A502
Eruptive unit	Hæðir	Hæðir	Hæðir	Húsmúll	Húsmúli	Húamúli	Húsmúli	Húsmúll	Húsmúll	Húsmúli	Húsmúli	Húsmúli	Húsmúll	Húsmúli	Húsmúli
Crystal #	1	1	4	1	1	1	2	2	2	3	3	3	4	4	4
Туре	1	1	I	1	I	1	Q	Q	Q	I	I.	1	ł	1	I
Variety	т	Т	Ť	т	т	Т				т	т	Т	(M)	(M)	(M)
Spot position	С	С	С	С	С	R	С	С	С	С	С	R	С	С	R
Spot reference	A512F1	A512F2	A512F5	A502F1	A502F2	A502F3	A502F4	A502F5	A502F6	A502F7	A502F8	A502F9	A502F10	A502F11	A502F12
SI02	48.06	47.97	47.83	47.93	48.42	49.73	51.5 <del>9</del>	53.02	52.69	46.78	46.39	49.75	45.77	45.12	45.98
TIO2	0.13	0.03	0.04	0.06	0.28	0.05	0.57	0.34	0.58	0.10	0.00	0.20	0.17	0.06	0.04
A1203	31.77	31.44	31.59	30.91	31.29	30.87	23.79	26.66	26.09	32.54	32.56	29.70	31.16	30.44	30.76
Cr2O3	0.00	0.00	0.00	0.01	0.00	0.10	0.24	0.05	0.00	0.02	0.00	0.01	0.00	0.00	0.00
FeO	1.14	1.04	1.23	0.58	0.90	1.11	3.07	1.63	1.84	0.64	0.80	0.81	0.51	0.46	0.44
MnO	0.00	0.00	0.01	0.18	0.00	0.02	0.07	0.00	0.14	0.18	0.06	0.02	0.00	0.00	0.00
MgO	0.75	0.77	0.86	0.23	0.38	0.35	1.62	0.71	0.72	0.25	0.30	0.45	0.42	0.18	0.38
CaO	15. <b>65</b>	15.96	15.62	16.02	16.02	15. <b>26</b>	11.31	11.69	11.25	17.33	17.51	14.61	15.77	15.22	15. <b>56</b>
Na2O	2.38	2.55	2.49	2.26	2.24	2.80	3.93	4.44	4.83	1.85	1.17	3.08	1.87	1.85	1.94
K20	0.00	0.00	0.00	0.07	0.02	0.10	0.24	0.20	0.20	0.02	0.00	0.00	0.03	0.02	0.08
Total	99.88	<b>99.75</b>	<b>99.68</b>	98.23	99.54	100.37	96.42	98.74	98.33	99.72	98.79	98.64	95.70	93.34	95.18
Atoma per 6 oxyge	ms														
SI	2.21	2.22	2.21	2.24	2.24	2.28	2.65	2.45	2.45	2.17	2.16	2.31	2.20	2.22	2.22
AI	1.73	1.71	1.72	1.71	1.70	1.67	1.34	1.45	1.43	1.78	1.79	1.63	1.76	1.76	1.75
Ca	0.77	0.79	0.77	0.80	0.79	0.75	0.58	0.58	0.56	0.86	0.87	0.73	0.81	0.80	0.80
Na	0.21	0.23	0.22	0.21	0.20	0.25	0.36	0.40	0.44	0.17	0.15	0.28	0.17	0.18	0.18
Partial totals	4.92	4.95	4.93	4.96	4.93	4.94	4.93	4.88	4.88	4.97	4.97	4.94	4.95	4.96	4.95
An%	78.38	77.60	77.56	79.66	79.76	75.10	61.40	59.26	56.27	83.74	85.25	72.41	82.34	81.99	81.62

Sample #	A383	A363	A383	A383	A383	A174	A174	A174	A368	A368	A368	A368	A368	A368	A368	A368
Eruptive unit	Nee	Nes	Nes	Nes	Nes	Há	Há	Há	Bæ	Bæ	Bæ	Bæ	Bee	Bæ	Bas	Bae
Crystal #	1	1	1	2	2	1	1	1	1	1	1	2	2	3	4	4
Туре	I	ł	I	Q	Q	I.	I	1	i	I	ł	Q	Q	Q	Q	Q
Variety	T	т	т			Ε	E	E	м	м	м					
Spot position	С	С	R	С	С	С	С	С	С	С	R	С	С	С	С	С
Spot reference	A383F1	A383F2	A383F3	A383F4	A383F5	A174F1	A174F2	A174F3	A368F1	A368F2	A368F3	A368F4	A368F5	A368F6	A368F7	A368F8
SI02	48.31	47.33	46.86	50.54	50.33	51.91	50.89	51.14	48.44	46.28	46.20	50.62	50.86	51.08	49.88	49.98
TiO2	0.00	0.07	0.00	0.18	0.01	0.00	0.22	0.03	0.00	0.00	0.16	0.13	0.08	0.06	0.05	0.02
AI203	32.10	31.37	32.14	29.13	29.30	29.64	29.06	29.45	32.52	32.59	32.24	28.64	29.13	29.36	30.30	30.14
Cr203	0.00	0.02	0.00	0.01	0.00	0.00	0.06	0.00	0.00	0.00	0.11	0.00	0.03	0.00	0.05	0.00
FeO	0.68	0.53	0.89	1.28	0.90	0.87	0.70	0.58	0.60	0.59	0.64	1.30	1.15	1.03	0.91	0.80
MnO	0.10	0.09	0.14	0.02	0.04	0.02	0.00	0.04	0.05	0.01	0.04	0.08	0.08	0.32	0.17	0.09
MgO	0.83	0.69	0.60	0.85	0.90	0.57	0.49	0.58	0.30	0.19	0.29	0.36	0.40	0.38	0.37	0.47
CeO	15.44	15. <b>63</b>	15.89	13.15	13.05	13.75	13.68	13.78	17.27	17.59	17.34	14.21	13.85	13.87	14.64	14.92
Na2O	2.65	2.45	2.05	3.65	3.92	3.65	3.40	3.64	1.71	1.38	1.77	3.58	3.61	3.35	3.26	3.12
K20	0.00	0.00	0.00	0.00	0.04	0.03	0.00	0.09	0.07	0.00	0.03	0.09	0.05	0.16	0.12	0.02
Total	100.11	98.18	<b>98</b> .57	<b>98</b> .81	98.48	100.44	<b>98</b> .47	99.33	<b>98.96</b>	98.63	98.80	<b>99</b> .01	99.26	99.60	99.74	<b>99.57</b>
Atoms per 8 oxyge	<b>ns</b>											`				
SI	2.22	2.22	2.19	2.34	2.34	2.36	2.36	2.35	2.17	2.16	2.16	2.35	2.35	2.35	2.30	2.30
Ał	1.74	1.73	1.77	1.59	1.60	1.59	1.59	1.60	1.79	1.80	1.78	1.57	1.58	1.59	1.64	1.64
Ca	0.76	0.78	0.80	0.65	0.65	0.67	0.68	0.68	0.86	0.88	0.87	0.71	0.69	0.68	0.72	0.74
Na	0.24	0.22	0.19	0.33	0.35	0.32	0.31	0.32	0.15	0.13	0.16	0.32	0.32	0.30	0.29	0.28
Partial totals	4.95	4.95	4.94	4.91	4.94	4.94	4.92	<b>4.95</b>	4.97	4.97	4.97	4.94	4.94	4.92	4.95	4.95
An%	76.36	77.93	81.04	66.53	64.77	67.51	68.94	<b>67</b> .70	84.86	87.57	84.45	68.68	67.96	69.62	71.27	72.58

Appendix 2 - A

Sample #	A368	A368	A368	A368	A368	A368	A374	A374	A374	A374	A374	A374	A374	A374	A374
Eruptive unit	Bee	Bæ	Bee	Bæ	Bæ	Bae	Hengili	Hengill	Hengili	Hengill	Hengill	Hengill	Hengill	Hengill	Hengill
Crystal #	5	6	6	6	7	7	1	1	1	2	2	2	3	3	4
Туре	1	I	1	1	I.	1	I	1	I	I.	1	1	Q	Q	i i
Variety	M	M	м	М	м	м	т	т	Т	Т	т	т			(M)
Spot position	С	С	С	R	С	R	С	I.	R	С	С	R	С	I	С
Spot reference	A368F9	A368F10	A368F11	A368F12	A368F13	A368F14	A374F1	A374F2	A374F3	A374F4	A374F5	A374F6	A374F7	A374F8	A374F9
SI02	46.72	46.91	46.88	48.78	46.28	50.68	47.78	50.33	48.01	48.26	46.30	47.23	49.80	50.52	47.23
TIO2	0.00	0.00	0.00	0.09	0.12	0.00	0.09	0.05	0.19	0.13	0.16	0.12	0.15	0.10	0.04
A1203	33.26	32.66	32.54	31.31	32.79	29.81	32.26	30.90	32.65	32.08	31.38	32.06	29.91	30.27	32.61
Cr2O3	0.08	0.10	0.05	0.00	0.00	0.00	0.04	0.00	0.00	0.00	0.04	0.14	0.01	0.00	0.00
FeO	0.59	0.59	0.67	0.81	0.47	0.59	0.38	0.62	0.75	0.43	0.43	0.77	0.93	0.98	0.55
MnO	0.17	0.22	0.04	0.11	0.00	0.11	0.14	0.10	0.14	0.08	0.00	0.00	0.10	0.08	0.00
MgO	0.32	0.22	0.50	0.34	0.24	0.39	0.55	0.48	0.56	0.67	0.45	0.56	0.56	0.62	0.59
CeO	17.49	17.17	17.60	15.75	17.69	14.57	17.18	15.33	16.73	16.42	15. <b>97</b>	16.63	14.88	14.89	17.11
Na2O	1.50	1.81	1.97	2.60	1.39	3.25	1.97	3.06	1.97	2.30	2.00	2.02	2.84	2.90	1.89
K20	0.06	0.01	0.00	0.02	0.00	0.06	0.02	0.06	0.00	0.00	0.03	0.01	0.02	0.00	0.05
Totel	100.18	99.70	100.25	99.83	98.99	99.46	100.39	100.93	101.00	100.37	96.75	99.53	99.16	100.34	100.07
Atoms per 8 oxyge	<b>ns</b>														
Sł	2.15	2.17	2.16	2.25	2.16	2.33	2.19	2.29	2.19	2.21	2.20	2.19	2.30	2.31	2.18
AI	1.81	1.78	1.77	1.70	1.80	1.62	1.75	1.65	1.76	1.73	1.76	1.75	1.63	1.63	1.77
Ca	0.86	0.85	0.87	0.78	0.88	0.72	0.85	0.75	0.82	0.81	0.81	0.83	0.74	0.73	0.84
Na	0.13	0.16	0.18	0.23	0.13	0.29	0.18	0.27	0.18	0.21	0.18	0.18	0.25	0.28	0.17
Partial totals	4.96	4.97	4.98	4.95	4.96	4.95	4.96	4.98	4.94	4.95	4.95	4.94	4.92	4.92	4.96
An%	86.56	83.93	83.17	77.01	87.60	71.20	82.84	73.50	82.36	79.72	81.54	82.01	74.34	73.98	83.32

Sample #	A374	A374
Eruptive unit	Hengili	Hengill
Crystal #	4	4
Туре	1	I.
Variety	(M)	(M)
Spot position	С	R
Spot reference	A374F10	A374F11
SI02	47.60	48.73
TIO2	0.00	0.18
AI203	32.66	31.72
Cr2O3	0.18	0.00
FeO	0.43	0.50
MnO	0.06	0.11
MgO	0.58	0.47
CeO	16.90	16.08
Na20	2.06	2.40
K20	0.00	0.00
Total	100.47	100.18
Atoms per 8 oxyge	ens	
SI	2.18	2.23
A	1.77	1.71
Ca	0.83	0.79

<b>An%</b>	81.93	78.74
Partial totals	4.96	4.95
Ne	0.18	0.21
Ca	0.83	0.79
AI	1.77	1.71

# Appendix 2 - B Olivine Mineral Compositions

This appendix contains the olivine mineral compositions from samples within the Hengill-Ingolfsfjall region. Analyses were carried out on the electron microprobe (at the University of Durham). The analyses are presented in this appendix according to the list below:

Volcanic System	Unit	Sample	Page	Мар
Hveragerði	GB4	A44	278	B3-4
	BA2	A523	278	B1-7
	BA3	A122	279	С
	BB5	A545	279	B3-3
	BB7	A36	280	B3-8
	DB7b	A474	280	B1-4
	S12	A495	281	B2-6
	ILS	A9	281	С
Hengill	Hæðir	A512	283	<b>B1-1</b>
	Húsmúli	A502	283	B1-3
	Nesjæskógur (Nes)	A383	284	<b>B1-1</b>
	Bæjarháls (Bæ)	A368	284	B1-5
	Hengill	A374	284	B1-2

#### Key to Appendix 2 - B:

Sample #	= sample number
Eruptive unit	= refers to the position in the stratigraphy (Figure 3.2 [p.43] & 3.12 [p. 64])
Crystal #	= 1, 2, 3, etc is a means of differentiating between individual crystals
Туре	= Intratelluric (I) or Quench (Q)
Spot position	= location of analyses within the crystal
	-C = core
	- $I =$ between the rim and the core
	-R = rim
Spot ref.	= analyses reference number within sample
Partial total	= total of cations present in the crystal

Hverageröl Voicanic System

	Sample #	A44	A44	A44	<b>A</b> 44	A44	A44	A44	A44	A523	A523	A523	A523	A523	A523
	Eruptive unit	GB4	GB4	GB4	GB4	GB4	GB4	GB4	GB4	BA2	BA2	BA2	BA2	BA2	BA2
	Crystal #	1	1	2	3	4	4	5	5	1	1	1	1	2	2
	Туре	1	I	I	I	1	1	1	I	I	ł	1	I	I	1
	Spot position	С	С	С	C ·	I	С	С	1	С	С	С	R	С	С
	Spot ref.	A440L1	A44OL4	A440L2	A44OL3	A4401.5	A440L8	A440L12	A440L7	A5230L1	A523012	A5230L3	A5230L4	A5230L5	A5230L6
	SI02	40.81	41.09	43.11	41.46	40.19	40.80	41.25	39.22	39.04	38.95	38,90	<b>39</b> .11	39.02	38.66
	TI <b>O2</b>	0.16	0.05	0.10	0.18	0.04	0.04	0.04	0.06	0.00	0.00	0.15	0.07	0.00	0.07
	Ai203	0.54	1.20	5.87	3.00	1.07	1.18	2.94	3.37	0.01	0.25	0.00	0.24	0.00	0.09
	Cr203	0.29	0.22	0.14	0.12	0.17	0.14	0.73	1.61	0.01	0.09	0.15	0.16	0.15	0.00
	FeO	10.92	10.50	10.55	10.25	11.03	11.47	10.71	11.55	17.58	17.76	17.47	17.40	17.26	17.35
	MnO	1.42	1.26	1.37	1.40	1.37	1.42	1.38	1.34	1.35	1.22	1.09	1.19	1.10	1.18
N	MgO	46.39	45.49	33.77	40.60	44.16	<b>44.56</b>	42.10	41.51	41.41	41.21	41.10	40.31	40.72	40.96
78	CaO	0.75	1.02	5.14	2.37	1.24	1.19	1.96	1.72	0.32	0.36	0.33	0.48	0.32	0.27
	Na2O	0.07	0.22	0.57	0.03	0.08	0.06	0.26	0.06	0.00	0.07	0.00	0.02	0.00	0.00
	K20	0.00	0.05	0.01	0.00	0.00	0.00	0.05	0.04	0.00	0.00	0.04	0.10	0.03	0.00
	NIO	0.23	0.22	0.07	0.23	0.58	0.01	0.32	0.36	0.26	0.14	0.04	0.00	0.05	0.33
	TOTAL	101.57	101.31	100.70	99.64	99.91	100.86	101.75	100.84	<b>99</b> .97	100.04	99.26	99.08	98.63	98.90
	Atoms per 4 oxy	gena													
	SI	1.03	1.04	1.09	1.06	1.03	1.04	1.04	1.00	1.03	1.03	1.03	1.04	1.04	1.03
	Fe	0.23	0.22	0.22	0.22	0.24	0.24	0.23	0.25	0.3 <del>9</del>	0.39	0.39	0.39	0.39	0.39
	Mg	1.74	1.71	1.27	1.55	1.69	1.69	1.58	1.58	1.63	1.62	1.63	1.60	1.62	1. <b>63</b>
	Partial total	3.00	2.97	2.58	2.82	2.96	2.97	2.84	2.83	3.05	3.04	3.05	3.03	3.05	3.05
	Fo %	88.35	88.52	85.09	87.60	87.70	87.39	87.52	86.51	80.76	80.53	80.75	80.49	80.79	80.79

Sample #	A523	A523	A523	A523	A523	A122	A122	A122	A122	A122	A545	A545	A545	A545
Eruptive unit	BA2	BA2	BA2	BA2	BA2	BA3	BA3	BA3	BA3	BA3	885	B85	B85	<b>BB5</b>
Crystal #	2	3	4	4	4	1	2	3	3	4	1	1	2	2
Туре	1	Q	Q	Q	Q	1	I	I.	1	1	1	1	I	1
Spot position	С	С	С	С	С	С	С	С	R	I.	1	I	С	С
Spot ref.	A523OL7	A5230L8	A5230L12	A5230L13	A5230L14	A1220L1	A1220L4	A1220L5	A1220L6	A1220L7	A5450L1	A5450L2	A5450L5	A5450L6
SiO2	39.17	39.59	38.34	38.37	38.95	40.44	40.55	39.77	40.38	40.23	33.52	33.07	33.45	33.38
TIO2	0.00	0.09	0.02	0.04	0.19	0.00	0.27	0.15	0.14	0.09	0.07	0.00	0.10	0.15
A1203	0.37	1.13	0.74	0.64	1.06	0.42	0.67	0.14	0.27	1.22	0.43	0.84	0.41	0.39
Cr2O3	0.03	0.00	0.09	0.06	0.13	0.04	0.08	0.06	0.10	0.24	0.09	0.18	0.14	0.19
FeO	17.53	22.33	22.34	<b>21.94</b>	21.81	16.11	16.00	16.06	16.27	16.15	45.72	44.84	46.61	47.96
MnO	1.46	1.35	1.23	1.20	1.35	0.90	1.15	1.12	0.76	0.85	2.14	1.97	2.08	1.86
MgO	41.24	36.20	35.72	36.16	35.44	43.31	41.76	43.06	43.08	40.62	16.44	15.84	15.14	15.21
CaO	0.28	0.95	0.87	0.83	1.10	0.61	0.71	0.62	0.52	0.68	0.77	0.60	0.46	0.45
Na20	0.09	0.12	0.00	0.08	0.02	0.02	0.07	0.05	0.11	0.05	0.29	0.19	0.16	0.33
K20	0.00	0.00	0.00	0.05	0.02	0.04	0.02	0.00	0.00	0.06	0.06	0.03	0.00	0.06
NIO	0.00	0.00	0.00	0.26	0.10	0.13	0.05	0.25	0.05	0.00	0.08	0.00	0.23	0.20
TOTAL	100.16	101.76	99.36	99.62	100.16	102.01	101.32	101.29	101. <b>67</b>	100.17	99.61	97.56	98.78	100.16
Atoms per 4 ox	ygens													
Si	1.03	1.05	1.04	1.04	1.05	1.04	1.04	1.03	1.04	1.05	1.03	1.04	1.05	1.03
Fe	0.39	0.49	0.51	0.50	0.49	0.34	0.34	0.35	0.35	0.35	1.18	1.18	1.22	1.24
Mg	1.62	1.43	1.45	1.46	1.42	1.65	1.60	1.66	1.65	1.58	0.76	0.74	0.71	0.70
Partial total	3.04	2.96	2.99	3.00	2.95	3.03	2.99	3.04	3.04	2.98	2.97	2.96	2.97	2.98
Fo %	80.75	74.29	74.03	74.60	74.33	82.74	82.30	82.70	82.51	81.76	39.05	38.64	36.66	36.12

Sample #	A545	A36	A474	A474	A474	A474	A474	A474	A474	A474	A474	A474	A474
Eruptive unit	BB5	887	DB7b	D87b	D87b	D87b	D87b	D87b	DB7b	DB7b	D87b	DB7b	DB7b
Crystal #	2	1	1	1	1	2	2	2	3	3	· 4	4	5
Туре	1	I	l I	I	I	1	I	I	I	I	I	1	Q
Spot position	С	I	С	С	R	С	С	R	С	С	С	С	С
Spot ref.	A5450L7	A360L1	A4740L1	A4740L2	A474OL3	A474OL4	A474OL5	A474OL6	A4740L7	A474OL8	A474OL9	A474OL10	A474OL11
SiO2	34.00	36.58	40.23	40.63	40.61	40.13	40.33	39.58	40.42	39.40	39.81	39.85	38.50
TIO2	0.07	0.07	0.00	0.00	0.02	0.13	0.05	0.00	0.10	0.20	0.03	0.07	0.00
AI203	0.70	0.25	0.00	0.00	0.30	0.00	0.10	0.12	0.27	0.18	0.00	0.03	0.69
Cr203	0.18	0.09	0.00	0.00	0.00	0.15	0.04	0.27	0.17	0.00	0.06	0.00	0.06
FeO	45.62	33.67	13.90	14.53	14.83	13.59	13.70	13.76	15.50	15.84	14.81	14.62	19.79
MnO	1.90	0.94	0.39	0.40	0.37	0.23	0.43	0.27	0.26	0.19	0.34	0.37	0.47
MgO	14.62	28.33	44.70	44.81	43.99	45.31	45.57	44.81	42.92	43.14	43.69	44.40	37.14
CeO	0.48	0.30	0.23	0.24	0.39	0.23	0.22	0.29	0.57	0.41	0.28	0.31	0.61
Na20	0.19	0.62	0.59	0.43	0.57	0.47	0.56	0.61	0.70	0.55	0.56	0.56	0.64
K20	0.00	0.00	0.00	0.00	0.05	0.02	0.03	0.01	0.00	0.01	0.00	0.00	0.00
NIO	0.00												
TOTAL	97.76	100.86	100.04	101.03	101.13	100.25	101.02	99.71	100.92	99.92	99.58	100.19	97.91
Atoms per 4 oxy	/gens												
SI	1.07	1.02	1.03	1.03	1.03	1.02	1.02	1.02	1.03	1.02	1.03	1.02	1.03
Fe	1.20	0.79	0.30	0.31	0.31	0.29	0.29	0.29	0.33	0.34	0.32	0.31	0.44
Mg	0.68	,1 <b>.18</b>	1.70	1.69	1.66	1.72	1.72	1.71	1.63	1.66	1.68	1.69	1.49
Partial total	<b>2.94</b>	2.99	3.03	3.03	3.01	3.03	3.03	3.02	2.99	3.02	3.02	3.03	2.97
Fo %	36.35	59.99	85.14	84.61	84.10	85.59	85. <b>58</b>	85.32	83.15	82.93	84.01	84.40	76.98

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Sample #	A474	A495	A495	A495	A495	A495	A495	A495	A9	<b>A9</b>	A9	A9	A9	A9
Eruptive unit	DB7b	S12	S12	S12	S12	S12	S12	S12	ILS	ILS	ILS	ILS	ILS	ILS
Crystal #	6	1	1	1	1	2	2	3	1	1	1	2	2	2
Туре	Q	I.	1	I	1	I I	I.	I I	I	I.	1	ł	E E	I
Spot position	F	Α	A	В	В	С	С	С	С	R	С	С	R	С
Spot ref.	A4740L12	A4950L1	A4950L2	A495OL3	A4950L4	A4950L6	A4950L7	A4950L9	A9OL1	A9012	A9016	A901.3	A9OL4	A901.5
5102	39.57	38.39	38.09	38.20	38.12	40.15	39.41	37.04	39.46	39.35	39.85	39.73	39.66	40.04
TIO2	0.05	0.01	0.04	0.07	0.07	0.01	0.00	0.02	0.09	0.00	0.03	0.00	0.00	0.09
AI203	0.13	0.10	0.09	0.07	0.05	0.12	0.10	0.13	0.00	0.15	0.09	0.23	0.15	0.24
Cr203	0.00	0.00	0.24	0.04	0.08	0.19	0.18	0.16	0.08	0.30	0.17	0.19	0.04	0.31
FeO	15.54	22.05	21.88	21.51	22.14	12.74	12.5 <del>9</del>	20.42	16.58	16.53	17.35	16.69	16.70	16.98
MnO	0.49	1.37	1.52	1.31	1.37	0.84	0.91	1.51	1.56	1.28	1.43	1.44	1.35	1.35
MgO	43.59	37.82	37.48	37.78	38.18	44.69	44.49	37.12	42.65	42.73	42.94	43.28	42.27	43.05
CaO	0.47	0.28	0.30	0.33	0.35	0.32	0.38	0.33	0.33	0.47	0.30	0.29	0.39	0.30
Na20	0.56	0.17	0.04	0.03	0.00	0.00	0.00	0.31	0.15	0.21	0.00	0.21	0.11	0.27
K20	0.00	0.09	0.00	0.05	0.02	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
NIO		0.00	0.09	0.00	0.13	0.41	0.29	0.05	0.25	0.15	0.48	0.45	0.42	0.17
TOTAL	100.40	100.28	<b>9</b> 9.76	99.40	100.49	99.48	98.34	97.07	101.15	101.16	102.63	102.52	101.07	102.80
Atoms per 4 oxy	gena													
Sł	1.02	1.03	1. <b>03</b>	1.03	1.03	1.04	1.03	1.03	1.03	1.02	1.02	1.02	1.03	1.03
Fe	0.33	0.50	0.50	0.49	0.49	0.28	0.28	0.47	0.36	0.36	0.37	0.36	0.36	0.36
Mg	1.67	1.52	1.51	1.52	1.52	1.73	1.74	1.53	1.65	1.66	1.65	1.66	1.64	1.64
Partial total	3.02	3.05	3.04	3.05	3.05	3.04	3.05	3.04	3.04	3.04	3.04	3.04	3.04	3.03
Fo %	83.34	75.35	75.32	75.78	75.78	86.21	86.30	76.42	82.11	82.18	81.53	82.22	81.87	81.89

Sample #	<b>A9</b>	A9	A9	A9	A9	A9	A9	A9	<b>A9</b>
Eruptive unit	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS
Crystal #	3	3	4	4	<b>`4</b>	5	5	5	6
Туре	ł	ł	Q	Q	Q	I	ŧ	1	Q
Spot position	С	С	С	С	С	С	С	С	С
Spot ref.	A9017	A90L8	A90L9	A9OL10	A90L11	A90L12	A90L13	A90L14	A90L16
SI02	39.56	39.39	39.40	39.10	39.24	39.64	38.79	39.05	38.70
TIO2	0.00	0.04	0.07	0.10	0.00	0.00	0.02	0.00	0.03
AI2O3	0.13	0.00	0.34	0.00	0.25	0.20	0.12	0.18	0.27
Cr203	0.09	0.16	0.04	0.19	0.22	0.07	0.11	0.18	0.04
FeO	16.41	16.50	18.02	18.41	17.90	16.15	16.48	16.22	17.43
MnO	1.31	1.45	1.43	1.41	1.44	1.26	1.50	1.41	1.36
MgO	42.86	41.86	41.01	40.52	40.64	41.90	42.14	42.54	39.92
CaO	0.33	0.36	0.37	0.44	0.38	0.37	0.25	0.42	0.48
Na2O	0.10	0.14	0.21	0.23	0.31	0.00	0.27	0.06	0.25
K20	0.00	0.00	0.01	0.00	0.03	0.00	0.06	0.00	0.01
NIO	0.06	0.01	0.00	0.07	0.09	0.10	0.36	0.37	0.14
TOTAL	100.85	99.90	100.90	100.45	100.50	99.68	100.08	100.43	98.63
Atoms per 4 oxy	gens								
SI	1.03	1.04	1.03	1.03	1.03	1.04	1.02	1.02	1.04
Fe	0.36	0.36	0.39	0.41	0.39	0.36	0.36	0.36	0.41
Mg	1.66	1.64	1.60	1.60	1.60	1.64	1.66	1.66	1.59
Partial total	3.05	3.04	3.03	3.04	3.02	3.04	3.04	3.04	3.04
Fo %	82.32	81.90	80.22	79.69	80.19	82.21	82.00	82.39	79.49

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Sample # Eruptive unit Crystal # Type	A512 Hæðir 1 I	A512 Hæðir 1 I	A512 Hæðir 1 I	A512 Hæðir 2 Q	A512 Hæðir 2 Q	A512 Hæðir 2 Q	A512 Hæðir 3 Q	A512 Hæðir 3 Q	A512 Hæðir 4 Q/1?	A512 Hæðir 4 Q/I?	A502 Húsmúli 1 I	A502 Húsmúli 1 I	A502 Húsmúil 1 I														
														Spot position	С	С	R	С	С	R	R	R	С	С	С	С	R
														Spot ref.	A5120L1	A5120L2	A512OL3	A512OL4	A512OL5	A512OL6	A5120L7	A512OL8	A5120L9	A512OL10	A502OL1	A5020L2	A5020L3
														SiO2	39.15	39.87	38.79	37.34	37.16	36.81	37.58	37.72	38.07	37.75	38.98	38.30	39.09
TI <b>O2</b>	0.00	0.00	0.00	0.12	0.01	0.11	0.00	0.00	0.00	0.00	0.00	0.05	0.00														
AI203	0.08	0.00	0.25	0.91	0.67	1.21	0.51	0.50	0.13	0.05	0.33	0.08	0.40														
Cr2O3	0.00	0.14	0.04	0.00	0.00	0.05	0.00	0.08	0.00	0.00	0.07	0.00	0.00														
FeO	18.36	18.57	22.87	29.33	28.26	27.30	26.68	26.27	23.26	23.02	20.44	20.24	19.71														
MnO	0.46	0.42	0.48	0.68	0.69	0.65	0.66	0.56	0.60	0.49	0.27	0.35	0.32														
MgO	41.34	41.88	37.19	30.19	30.21	29.79	32.34	32.56	36.17	35.80	39.96	39.40	39.12														
CaO	0.35	0.35	0.53	0.66	0.56	0.92	0.61	0.64	0.39	0.33	0.47	0.43	0.53														
Na2O	0.28	0.45	0.55	0.87	0.64	0.85	0.81	0.65	0.71	0.80	0.11	0.00	0.33														
K20	0.01	0.00	0.00	0.00	0.07	0.00	0.00	0.02	0.00	0.00	0.04	0.00	0.00														
NIO																											
TOTAL	100.01	101. <del>6</del> 7	100.68	100.11	98.27	97.68	99.17	98.97	99.33	98.24	100.68	98.88	99.50														
Atoms per 4 ox)	ygen <b>s</b>																										
Si	1.02	1.02	1.03	1.03	1.04	1.03	1.03	1.04	1.03	1.03	1.02	1.02	1.03														
Fe	0.40	0.40	0.51	0.68	0.68	0.64	0.61	0.60	0.52	0.53	0.45	0.45	0.43														
Mg	1.61	1.60	1.47	1.24	1.26	1.25	1.32	1.33	1.45	1.45	1.56	1.56	1.54														
Partial total	3.03	3.02	3.00	2.95	2.96	2.92	2.97	2.97	3.01	3.01	3.02	3.04	3.00														
Fo %	80.06	80.09	74.35	64.72	65.58	68.04	68.39	68.85	73.49	73.48	77.69	77.63	77.96														

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.
Sample #	A383	A383	A368	A368	A368	A368	A368	A368	A374	A374	A374	A374	A374	A374
Eruptive unit	Nes	Nes	Bæ	Bæ	Bæ	Bæ	Bae	Bæ	Hengili	Hengill	Hengili	Hengill	Hengili	Henglii
Crystal #	1	1	1	1	1	2	2	2	1	1	1	2	2	2
Туре	Q	Q	Q	Q	Q	Q	Q	Q	Q	Q	Q	I	1	E E
Spot position	С	С	С	I	F	С	С	R	С	С	ł	С	С	R
Spot ref.	A383OL1	A3830L2	A368OL1	A3680L2	A368OL3	A368OL4	A368OL5	A368OL6	A3740L1	A3740L2	A374OL3	A3740L4	A3740L5	A374OL6
SiO2	35.96	37.38	39.73	39.12	39.17	40.09	39.99	38.78	39.23	39.22	38.55	39.72	39.55	39.41
TI02	0.19	0.00	0.13	0.06	0.11	0.07	0.04	0.08	0.39	0.00	0.36	0.10	0.00	0.07
AI203	0.23	0.32	0.42	0.28	0.48	0.11	0.00	0.14	2.10	0.67	3.10	0.13	0.21	0.38
Cr2O3	0.10	0.00	0.13	0.16	0.01	0.00	0.05	0.01	0.00	0.06	0.02	0.00	0.04	0.00
FeO	15.16	17.52	20.14	20.46	20.79	15.18	15.14	14.00	17.47	18.20	17.16	17.39	17.12	16.79
MnO	0.33	0.29	0.29	0.30	0.39	0.48	0.38	0. <b>29</b>	0.41	0.41	0.23	0.39	0.32	0.56
MgO	37.10	38.64	39.19	39.68	39.69	44.42	44.60	42.92	36.18	40.48	33.88	42.38	42.53	40.81
CaO	0.43	0.54	0.49	0.47	0.42	0.33	0.30	0.35	1.03	0.73	1.26	0.34	0.34	0.89
Na2O	0.88	0.94	0.38	0.25	0.28	0.18	0.12	0.11	0.70	0.46	0.40	0.48	0.48	0.66
K20	0.08	0.00	0.09	0.00	0.04	0.06	0.00	0.00	0.02	0.07	0.05	0.00	0.03	0.05
NIO														
TOTAL	90.45	95.62	100.99	100.78	101.38	100.90	100.59	<b>96.68</b>	97.52	100.29	<b>9</b> 5.01	100.94	100.61	99.62
Atoms per 4 oxy	gens													
SI	1.03	1.02	1.03	1.02	1.02	1.02	1.02	1.03	1.04	1.02	1.05	1.02	1.02	1.03
Fe	0.36	0.40	0.44	0.45	0.45	0.32	0.32	0.31	0.39	0.40	0.39	0.37	0.37	0.37
Mg	1.59	1.57	1.52	1.54	1.54	1.69	1.70	1.69	1.43	1.57	1.37	1.63	1.63	1.59
Partial total	2.98	3.00	2.99	3.01	3.01	3.03	3.04	3.03	2.87	2.99	2.81	3.02	3.02	2.98
Fo %	81.37	79.72	77.63	77.56	77.28	83.92	84.01	84.53	78.68	79.86	77.86	81.29	81.58	81.23

Sample #	A374	A374	A374	A374	A374
Eruptive unit	Hengill	Henglil	Hengill	Hengill	Henglil
Crystal #	3	3	3	4	4
Туре	I	1	1	I	1
Spot position	С	С	R	С	С
Spot ref.	A3740L7	A374OL8	A3740L9	A374OL10	A374OL11
SI02	39.92	39.76	39.48	40.04	40.01
TIO2	0.00	0.06	0.04	0.17	0.06
A1203	0.13	0.33	0.13	0.13	0.03
Cr2O3	0.06	0.00	0.07	0.08	0.12
FeO	17.76	16.96	17.92	16.93	17.58
MnO	0.32	0.48	0.42	0.25	0.53
MgO	42.99	42.39	42.75	42.08	42.34
CaO	0.39	0.43	0.38	0.38	0.51
Na20	0.47	0.44	0.57	0.39	0.35
K20	0.01	0.00	0.00	0.00	0.00
NIO					
TOTAL	102.05	100.88	101.74	100.43	101.53
Atoms per 4 ox	ygens				
SI	1.02	1.02	1.01	1.03	1.02
Fe	0.38	0.36	0.38	0.38	0.38
Mg	1.63	1.62	1.63	1.61	1.62
Partial total	3.03	3.01	3.03	3.01	3.02
Fo %	81.20	81.67	80.97	81.57	81.10

# Appendix 2 - C Pyroxene Mineral Compositions

This appendix contains the pyroxene mineral compositions from samples within the Hengill-Ingolfsfjall region. Analyses were carried out on the electron microprobe (at the University of Durham). The analyses are presented in this appendix according to the list below:

Volcanic System	Unit	Sample	Page	Мар
Hveragerði	BB7	<b>A3</b> 6	288	B3-8
	DB10c	A349	288	B2-5
	<b>S</b> 5	A153	288	С
	S12	A495	<b>28</b> 8	B2-6
Hengill	Húsmúli	A502	290	B1-3
	Nesjaskógur (Nes)	A383	290	B1-1
	Hahryggur (Há)	A174	290	B1-1
	Bæjarháls (Bæ)	A368	290	B1-5

### Key to Appendix 2 - C:

Sample #	= sample number
Eruptive unit	= refers to the position in the stratigraphy (Figure 3.2 [p.43] & 3.12 [p. 64])
Crystal #	= 1, 2, 3, etc is a means of differentiating between individual crystals
Туре	= Intratelluric (I) or Quench (Q)
Spot position	= location of analyses within the crystal
	-C = core
	-I = between the rim and the core
	$-\mathbf{R} = \mathbf{rim}$
Spot ref.	= analyses reference number within sample
Partial total	= total of cations present in the crystal

Hverageröl V	olcanic S	system												
Sample #	A36	A36	A36	A349	A349	A349	A153	A153	A153	A153	A153	A153	A153	495
Eruptive unit	BB7	887	<b>BB7</b>	DB10c	DB10c	DB10c	<b>S</b> 5	<b>S</b> 5	<b>S</b> 5	S5	<b>S5</b>	<b>S5</b>	S5	S12
Crystal #	1	1	1	1	1	1	1	2	2	2	2	2	2	1
Туре	1	I	1	Q	Q	Q	Q	Q	Q	Q	Q	Q	Q	ł
Spot position	С	С	R	I	I	С	С	С	С	С	R	R	R	С
Spot ref.	A36PX1	A36PX2	A36PX3	A349PX1	A349PX2	A349PX3	A153PX1	A153PX2	A153PX3	A153PX4	A153PX5	A153PX6	A153PX7	A495PX1
SIO2	49.59	48.32	49.83	50.9 <del>6</del>	51.01	51.64	<b>52.38</b>	53,38	<b>52.68</b>	53.73	51.88	51.35	51.13	50.79
TIO2	1.49	1.97	1.22	0.81	0.62	0.80	0.90	0.64	0.82	0.75	0.83	0.93	0.84	0.93
Al2O3	3.77	4.54	2.75	3.56	2.58	3.32	3.76	2.35	2.80	2.42	3.82	3.72	3.63	1.70
Cr2O3	0.02	0.13	0.03	0.32	0.07	0.20	0.57	0.31	0.38	0.30	0.51	0.58	0.46	0.23
FeO	12.08	12.59	9.34	8.36	9.38	7.22	7.37	7.32	7.93	7.47	7.15	7.37	7.13	9.49
MnO	0.75	0.55	0.44	0.43	0.55	0.43	0.35	0.31	0.17	0.27	0.25	0.22	0.29	1.49
MgO	14.33	13.78	14.46	15.12	14.84	15.14	15. <b>94</b>	17.22	16.49	17.12	15.68	15.57	15.81	14.62
CaO	17.04	17.02	18.46	19.91	19.32	20.21	21.31	19.92	20.15	20.55	21.33	21.13	21.04	19.49
Na20	0.76	0.65	0.75	0.69	0.42	0.34	0.58	0.43	0.35	0.27	0.48	0.38	0.44	0.06
K20	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.05	0.00	0.00	0.01
TOTAL	99.84	99.51	97.27	100.14	98.79	99.29	103.16	101.87	101.78	102.87	101.97	101.25	100. <b>76</b>	98.81
Atoms per 6 d	xygena													
SI	1.89	1.85	1.86	1.91	1.94	1.94	1.88	1.93	1.91	1.93	1.88	1.88	1.88	1.92
AI	0.17	0.21	0.13	0.16	0.12	0.15	0.16	0.10	0.12	0.10	0.16	0.16	0.16	0.08
π	0.04	0.06	0.04	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.03	0.02	0.03
F●	0.38	0.40	0.30	0.26	0.30	0.23	0.22	0.22	0.24	0.22	0.22	0.23	0.22	0.30
Mg	0.81	0.79	0.83	0.84	0.84	0.85	0.85	0.93	0.89	0.91	0.85	0.85	0.87	0.83
Ca	0.70	0.70	0.77	0.80	0.79	0.81	0.82	0.77	0.78	0. <b>79</b>	0.83	0.83	0.83	0.79
Partial total	3.99	4.00	3.92	4.00	4.01	3.99	3.96	3.97	3.97	3.97	3.97	3.97	3.97	3.94
Wo	36.73	37.02	40.24	41.93	40.87	43.07	43.2 <del>9</del>	40.18	40.90	40.94	43.80	43.54	43.29	41.24
En	42.94	41.60	43.86	44.32	43.66	44.92	45.04	48.31	46.53	47.43	44.75	44.64	45.27	43.07
Fa	20.33	21.38	15.90	13.75	15.47	12.01	11.67	11.52	12.57	11.62	11.45	11.82	11.44	15.69

Sample #	495	495	495	495	495	495	495	495	495	495	495	495	495
Eruptive unit	<b>S12</b>	<b>S12</b>	<b>S12</b>	S12	<b>S12</b>	<b>S12</b>	<b>S12</b>	<b>S12</b>	S12	<b>S12</b>	S12	<b>S12</b>	S12
Crystal #	1	1	2	2	2	3	4	4	4	5	5	5	6
Туре	1	1	1	I	I	1	I	ł	1	1	1	ł –	I.
Spot position	С	С	С	С	С	С	С	С	С	С	С	С	С
Spot ref.	A495PX2	A495PX3	A495PX4	A495PX5	A495PX6	A495PX7	A495PX8	A495PX9	4495PX10	A495PX11	A495PX12	A495PX13	A495PX14
SIO2	51.11	50.88	48.28	48.07	48.48	48.00	47.99	48.84	48.70	49.51	47.48	49.44	49.78
T102	0.82	1.11	1.38	1.55	1.15	0.92	1.08	0.94	0.81	0.80	1.12	0.79	1.26
AI203	1.89	2.09	3.43	3.74	2.86	2.89	3.00	3.38	3.01	2.41	4.31	2.20	3.38
Cr2O3	0.03	0.22	0.23	0.04	0.15	0.12	0.37	0.37	0.32	0.00	0.20	0.10	0.25
FeO	9.66	9.40	8.86	11.07	10.20	9.21	6.53	6.61	7.14	7.60	7.54	7.42	8.67
MnO	1.28	1.38	1.40	1.49	1.30	1.36	1.15	1.38	1.46	1.22	1.32	1.01	1.30
MgO	14.62	14.47	13.99	12,48	13.30	13.93	14.74	14.43	15.15	15.45	13.91	15.66	14.39
CaO	19.72	19.71	18.94	16.65	17.44	19.20	19.9 <b>7</b>	20.03	20.01	18.83	19.78	19.02	19.87
Na20	0.22	0.28	0.51	0.66	0.52	0.30	0.22	0.52	0.29	0.24	0.39	0.24	0.65
K20	0.05	0.00	0.00	0.1 <del>6</del>	0.09	0.02	0.00	0.01	0.00	0.00	0.00	0.00	0.08
TOTAL	99.40	99.54	<b>97.0</b> 0	95.90	95.48	95.95	95.03	96.50	96.89	96.06	96.05	95.88	99.62
Atoms per 6 c	xygens												
SI	1.93	1.92	1.87	1.89	1.91	1.88	1.88	1.88	1.87	1.91	1.85	1.91	1.87
AI	0.08	0.09	0.16	0.17	0.13	0.13	0.14	0.15	0.14	0.11	0.20	0.10	0.15
ТІ	0.02	0.03	0.04	0.05	0.03	0.03	0.03	0.03	0.02	0.02	0.03	0.02	0.04
Fe	0.31	0.30	0.29	0.36	0.34	0.30	0.21	0.21	0.23	0.25	0.25	0.24	0.27
Mg	0.82	0.81	0.81	0.73	0.7 <b>8</b>	0.81	0.86	0.83	0.87	0.89	0.81	0.90	0.81
Ca	0.80	0.80	0.79	0.70	0.74	0.81	0.84	0.83	0.83	0.78	0.83	0.79	0.80
Partial total	3.96	3.94	3.94	3.90	3.92	3.98	3.96	3.93	3.96	3.96	3.95	<b>3.96</b>	3.94
Wo	41.42	41.78	41.82	39.04	39.73	41.98	43.80	44.25	42.88	40.72	43.98	40.83	42.58
En	42.7 <b>2</b>	42.67	42.94	40.71	42.18	42.32	45.00	44.36	45.17	46.47	<b>42.96</b>	46.74	42.90
Fe	15.87	15.55	15.24	20.25	18.11	15.72	11.20	11.40	11.95	12.81	13.08	12.44	14.51

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			Hengili Vo	icanic Syst	em								
Sample #	495	495	A502	A502	A502	A383	A174	A174	A174	A368	A368	A368	A368
Eruptive unit	<b>S12</b>	S12	Húsmúll	Húsmúli	Húsmúli	Nes	Há	Há	Há	Bæ	Bæ	Bæ	Bæ
Crystal #	6	6	1	1	2	1	1	1	1	1	1	1	2
Туре	I	ł	a	Q	Q	Q	I	1	I	Q	Q	Q	Q
Spot position	С	С	С	С	I	С	С	С	С	С	С	R	С
Spot ref.	A495PX15	6A495PX16	A502PX1	A502PX2	A502PX3	A383PX2	A174PX	I A174PX2	2A174PX3	A368PX1	A368PX2	A368PX3	A368PX4
SI02	49.04	49.12	49.59	49.50	48.39	47.38	52.41	52.26	53.07	49.86	50.07	51. <b>66</b>	47.34
TIO2	1.08	1.33	1.80	1.57	1.92	2.30	0.54	0.51	0.49	1.27	1.36	0.88	2.24
AI203	3.02	3.13	4.34	4.35	4.36	5.95	2.39	2.34	2.54	4.59	4.27	5.7 <del>6</del>	5.17
Cr2O3	0.13	0.00	0.48	0.48	0.00	0.00	0.07	0.10	0.17	0.57	0.61	0.27	0.12
FeO	8.81	8.75	7.87	7.78	12.72	10.12	6.70	6.34	6.75	7.18	7.08	7.15	10.54
MnO	1.09	1.34	0.37	0.33	0.21	0.28	0.16	0.33	0.31	0.26	0.13	0.32	0.16
MgO	14.18	14.55	14.57	14.53	13.40	13.54	17.31	16.68	17.11	14.73	14.95	14.24	12.95
CaO	19.70	20.10	<b>19</b> .52	1 <b>9</b> .57	16.82	17.93	19.69	19.81	19.77	20.73	20.63	18.81	19.30
Na20	0.50	0.63	0.41	0.40	0.78	0.84	0.59	0.49	0.66	0.60	0.44	0.62	0.57
K20	0.00	0.00	0.00	0.12	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	97.52	98.95	98.95	98.61	<b>98.58</b>	98.33	99.85	98.85	100.88	<b>99.79</b>	99.55	<b>99.69</b>	98.39
Atoms per 6 o	xygens												
SI	1.88	1.87	1.88	1.88	1.87	1.82	1.95	1.96	1.95	1.87	1.88	1.92	1.83
AI	0.14	0.14	0.19	0.19	0.20	0.27	0.10	0.10	0.11	0.20	0.19	0.25	0.24
ті	0.03	0.04	0.05	0.04	0.06	0.07	0.02	0.01	0.01	0.04	0.04	0.02	0.07
Fe	0.28	0.28	0.25	0.25	0.41	0.33	0.21	0.20	0.21	0.23	0.22	0.22	0.34
Mg	0.81	0.82	0.82	0.82	0.77	0. <b>78</b>	0. <b>96</b>	0.93	0.94	0.83	0.84	0.79	0.75
Ca	0.81	0.82	0.79	0.80	0.70	0.74	0.78	0.80	0.78	0.83	0.83	0.75	0.80
Pertial total	3.96	3. <b>96</b>	3.99	3. <b>99</b>	4.00	4.00	4.02	4.00	4.00	4.00	4.00	3.95	4.02
Wo	42.55	42.60	42.52	42.69	37.07	40.16	40.17	41.31	40.47	44.25	43.93	42.56	42.37
En	42.60	42.91	44.11	44.07	41.06	42.17	49.15	48.39	48.74	43.77	44.29	44.80	39.56
Fs	14.85	14.49	13.37	13.23	21.86	17.67	10. <b>68</b>	10.31	10. <b>79</b>	11.98	11.78	12.63	18.07

Sample #	A368	A368		
Eruptive unit	Bæ	Bæ		
Crystal #	2	3		
Туре	Q	Q		
Spot position	С	С		
Spot ref.	A368PX5	A368PX6		
SIO2	47.37	48.23		
<b>TIO2</b>	2.66	1.80		
AI2O3	<b>5.68</b>	5.11		
Cr2O3	0.17	0.13		
FeO	12.07	8.52		
MnO	0.37	0.17		
MgO	12.67	13.47		
CaO	18.35	20.13		
Na20	0.64	0.39		
K20	0.02	0.00		
TOTAL	100.00	97.94		
Atoms per 6	oxygens			
SI	1.81	1.85		
AI	0.26	0.23		
п	0.08	0.05		
Fe	0.39	0.27		
Mg	0.72	0.77		
Ca	0.75	0.83		
Partial total	4.00	4.01		
Wo	40.42	44.20		
En	38.83	41.17		

14.63

20.74

•

Fs

# Appendix 2 - D Opaque Mineral Compositions

This appendix contains the opaque mineral compositions from samples within the Hengill-Ingolfsfjall region. Analyses were carried out on the electron microprobe (at the University of Durham). The analyses are presented in this appendix according to the list below:

Volcanic System	Unit	Sample	Page	Map
Hveragerði	BB5	A545	294	B3-3
	BB7	A36	294	B3-8
	DB7b	A474	294	<b>B</b> 1-4
	DB10c	A349	294	B2-5
Hengill	Hàhryggur (Há)	A174	294	<b>B</b> 1-1

## Key to Appendix 2 - D:

Sample #	= sample number
Spot reference	= analyses number within sample

Sample # Spot ref. Mineral	A545 A5450P1 M	A545 A5450P2 M	A545 A5450P3 M	A36 A36OP1 M	A36 A36OP2 M	A474 A474OLI1 C-S	A474 A474OLI2 C-S	A346 A349OP1 M	A346 A349OP2 M	A174 A174OP3 M	A174 A174OP4 M	A174 A174OP5 M	A174 A174OP6 M
SIO2	93	4.9	9.8	4.0	1.6	5.7	2.3	1.0	1.6	19.2	8.8	11.9	11.0
	21.6	23.9	21.3	22.1	23.8	1.3	1.4	23.4	23.4	16.6	20.6	19.3	20.0
A12O2	10	1 9	1.8	22	2.1	20.2	22.1	1.6	2.0	4.8	3.5	3.4	2.7
A1203	1.9	0.1	0.1	0.0	0.0	24.1	26.5	0.1	0.1	0.1	0.2	0.0	0.2
Cr2O3	0.2 53 1	57.5	53.7	58.6	59.8	27.6	27.7	62.4	61.1	45.5	55.4	53.1	53.9
	1.6	15	14	0.8	0.8	0.6	0.0	0.9	1.1	0.4	0.7	0.8	0.5
MaQ	3.1	1.8	3.1	1.7	1.7	15.5	13.5	0.2	0.4	4.6	2.6	3.2	3.4
mg0 CoO	3.1	1.0	27	0.3	0.2	0.6	0.1	0.1	0.3	5.0	1.9	2.8	2.9
	2.5	0.4	03	0.0	0.3	0.5	0.3	0.4	0.4	1.2	0.8	0.8	0.6
Nazo	0.5	0.4	0.5	0.4	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
TOTAL	94.0	93.1	94.3	90.3	90.5	96.1	93.8	90.2	90.2	97.4	94.5	95.3	95.1

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## Appendix 3 Preparation of the geochemical data set

## **Contents**

Part A: Analytical procedure for the X-ray fluorescence (XRF) spectrometer
 A3.1 Introduction
 A3.2 XRF sample preparation
 A3.3 XRF analyses
 A3.4 Fusion disc preparation method; an investigation
 Part B: Microprobe analyses
 Part C: CIPW norms
 Part D: Chondrite normalising values

#### Part A :

#### Analytical procedure for the X-ray fluorescence (XRF) spectrometer

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#### A3.1) Introduction

During this research, samples were mainly analysed by XRF (data tabulated in Appendix 4 and Chapter 6). Major changes and improvements have been made to the geochemical laboratories and workshops at Durham over recent years. These include the purchase and installation of equipment and software that was unavailable at the start of the project. Equipment purchased includes an Agate tema mill, a 'Fritsch Pulverisette' jaw crusher (type 01-704), a fly-press and the Phillips X40 software for the XRF. The workshops and laboratories were subsequently re-organised to accommodate this new equipment. The author contributed to these changes in many ways, one of which was to initiate, and take part in an experiment comparing the various existing XRF fusion disc preparation techniques which resulted in a change of the method used at Durham (see Section A3.4 below).

#### A2) XRF sample preparation

#### Equipment

Due to the transitional nature of the period between 1988 and 1991, the same preparation method could not be applied to all the samples, and consequently the suite was crushed by various peaces of equipment. In this appendix, 'method A' refers to the method used to prepare the first 100 samples, processed during 1988-1989 (samples A1-A160); and 'method B' refers to samples processed after that (samples A161-A560) (Table A3.1). Samples analysed for trace elements by ICP-AES and ICP-MS were prepared using method B.

Methods	Method 'A' - equipment used	Method 'B' - equipment used
Crushing	-Hand splitter	-Hand splitter
	-'Strurtevant' jaw crusher	-'Fritsch Pulverisette' jaw
		crusher (type 01-704)
		-Fly press
Powdering	-Tungsten Carbide Swing Mill	-Fritsch Agate Ball Mill Tema
	(60-120 seconds)	(20-30 minutes)

Table A3.1: Equipment used to prepare the samples processed during this research.

The agate tema was used in preference to the tungsten tema for basaltic rocks to avoid contamination with Ta, Nb and Co (Co 20-40 ppm; Day, 1989). To investigate the degree of contamination of samples processed here, some samples were prepared using both methods 'A' and 'B', and subsequently analysed by XRF (e.g. sample A14, Appendix 4, p.331). The difference in concentrations of elements Nb and Co from the two different powder types was found to be negligible and within analytical error ( $2\sigma$ ).

#### Cleaning

All crushing equipment was thoroughly cleaned between samples, first with water and rag followed by brushes (metal and/or bristle), and finally absolute alcohol (or acetone) before drying with paper towels. The tema pots were cleaned with quartz-sand when other cleaning agents failed to do the job, however, for some of the very altered material, this procedure was carried out between successive samples. Dilute hydrochloric acid and rag were used as a less time consuming substitute for cleaning the agate pots towards the end of the research project (M. A. Morrison, pers. comms., 1991). The whole cleaning process was enhanced between samples of diverse petrographic appearance, and also at the beginning and the end of each crushing and powdering session to avoid cross contamination with rocks of diverse geochemical affinity.

#### Powdered Pellet preparation

Approximately 5g of dried powder was added with 10 drops of PVA 'Mowial' organic binding solution (4% solution), mixed thoroughly, and subjected to 7-9 tons of pressure within a mould. The resulting pellet was then dried in the oven overnight (80°C). The analytical surface was chosen, indicated and isolated accordingly with pellets stored in sealed air-tight plastic bags.

#### Fusion disc preparation

After completion of the geological map and stratigraphy, and after the XRF analyses on powdered pellets was completed and the data had been examined (major and trace elements), a subset of samples was chosen to analyse for more accurate major element compositions using fusion discs. The criteria for selection of this subset included: samples with a tightly constrained stratigraphic location; balanced stratigraphic location relative to the other chosen samples; freshness; and an equal distribution of samples between fissure- and lava shield-type eruption events. Fifty samples were selected.

The fusion discs were made following the Leicester University method of preparation (see Section A3.4 below). The 'Loss On Ignition' (LOI) of each sample was calculated first by weighing approximately 5g of powder for each sample into a porcelain crucible and heating it in a furnace (used solely for LOI) at 900°C for two hours. The crucibles were allowed to cool completely before being re-weighed. Due to the fresh nature of most of the samples chosen, many actually increased in weight, constituting a 'gain'. It is possible that samples may have taken up atmospheric moisture during the short cooling period, so increasing in post-ignition weight and decreasing the LOI%, but this could not be avoided.

The fusion disc preparation method used (the 'Leicester method') involves the use of Johnson Matthey Spectroflux 100B (Lithium Metaborate and Lithium Tetraborate flux, with no La oxide present). The ignited powders were added to the flux in the proportions of 1:5 (0.45g powder:2.25g flux) and were very carefully and thoroughly mixed with a polythene rod. The crucibles, with their lids, were then placed in the furnace (used solely for fused disc preparation) at 1050°C for twenty minutes. One at a time, the four crucibles were removed from the furnace (whilst stirring gently) and poured on to a hot graphite disc and immediately flattened with a stainless steel plunger. They were then covered with a heated glass jar to prevent quench-fragmentation of the discs, and allowed to cool. The analytical surface on the disc was chosen and labelled. This was consistently the side that was in contact with the plunger, because it is completely flat. The discs were immediately placed and sealed in plastic bags. The used crucibles are placed in a boiling solution of dilute hydrochloric acid for 5-10 minutes to remove any remaining glass before being used for the next batch of samples.

It is standard practice to make three discs of each sample to provide a cross check on the method of preparation. However, to economise on time, it was decided to make two and only a third if there was inconsistency. The two discs were analysed in succession on the XRF to minimise the error from machine drift. Only one sample out of fifty had to have a third disc made . The major element oxide data from fused discs in Appendix 4 represent mean composition of the two discs.

#### A3.3) XRF analyses

#### Introduction

Whole-rock major oxide and trace element concentrations were acquired on the Phillips PW1400 wavelength dispersive X-ray fluorescence spectrometer with automatic sample changer

### Appendix 3

PW1051/10 and using a Rh X-ray tube (accelerating potential=80KV, 35mA; or =55KV, 50mA when analysing for REEs). Various other types of software were investigated during the course of this research, but the final data set tabulated in Appendix 4 is data acquired using the X40 software.

#### Software

Prior to 1988, the Durham XRF laboratory analysed only powder pellets for both major and trace elements, and used a variety of software to process the raw counts of the peaks and backgrounds (e.g. Pearce, 1988). In 1988, the first fusion discs were made at Durham (made in the Durham-Edinburgh technique) and analysed for major element oxides with the raw counts processed by the CBI software (detailed in Day, 1989). This is a program written by C. Watson (in 1972) and later modified and implemented for microcomputer use by D. Stevenson (in 1988). The results for major element oxide abundances acquired using this software were so poor that the 'fusion+CBI' technique was abandoned (Day, 1989). The Leicester method of fusion preparation used here was adopted later in 1989 and produced much better results using the CBI software (see Section A3.4 below).

The CBI program works on the principal of a cubic-curve fit and an iterative mass absorption correction process. Each iteration normalises the total composition including the LOI to 100%, and the calculation proceeds until the resultant compositions change by less than a predefined tolerance level. The program is unsatisfactory on many accounts which are outlined below but are described in more detail in Day (1989). The major element oxide result files from the CBI software are then used in a program written by Dr, N. J. Pearce (1988) (and implemented for microcomputer use by D. Stevenson in 1988), to analyse the trace elements in syenites (a method adapted from the Greenland Geological Survey as described by Bailey (1976)). The settings of this program are fixed, but were not suitable for the range of tholeiitic basalts encountered in this research. However, in 1988 & 1990, in the absence of other suitable software, the CBI and Pearce programs had to be used to analyse 200 samples from this research. This data is considered unacceptable for reasons outlined here below and in Day (1989), and has not been used in this thesis. By 1991, the new software (the more widely used Phillips X40 Alpha on-line software) had been installed and the previously analysed 200 pellets, plus an additional 225 pellets, were all analysed in one almost continuous session in February to May, 1991. The Phillips software has many advantages over the CBI and Pearce software previously used at Durham, and these are outlined briefly below. The new X40 software is still being explored by Durham's Senior Experimental officer in Geology (R. Hardy) and other research workers involved in the use of this facility.

1) There is much more user-control over the numerous options available within the Phillips package. These options include the incorporation of a Monitor and the LOI into the calculations, and a choice on whether the results are normalised to 100%. The detection limits for all the elements in an individual run can be calculated.

#### Preparation of geochemical data set

2) The CBI software works by analysing the standards and the unknowns together (i.e. acquiring the data) and then calculating the best-fit calibration curve for the standards. The operator has little control over this process because the standards can only be removed from the input file *after* the data has been acquired (D. Stevenson, pers. comms., 1991). In the Phillips software package, the calibration of the machine can be manipulated before the data is acquired. This is done by subtracting standards from the calibration line where needed so that the line is improved. This calibration was then tested by analysing International Reference Material (IRM) samples as unknowns. When the results of such a test were not satisfactory for a single element, the line could be altered and improved for that element. In this way, each element is totally independent of the others.

3) The CBI and Pearce programs governed that the major elements had to be determined first followed by the trace elements, so that the pelleted samples had to be put through the machine twice. The Phillips software allows all the elements and any combination of elements, so that a few select elements can be analysed at any one time.

4) The major element oxide data acquired from fusion discs is very sensitive to the method of preparation. If the discs are incorrectly made the effects are likely to be demonstrated by poor totals. The CBI program would mask this problem because the iterative process continually normalises to 100%. This is probably the greatest drawback of the CBI software in its present form.

#### Pellet analyses using X40 software

The powdered pellets were analysed for 7 major element oxides (FeO\*, CaO, Na<sub>2</sub>O, K<sub>2</sub>O, TiO<sub>2</sub>, MnO and P<sub>2</sub>O<sub>5</sub>), and a minimum of 11 trace elements (Nb, Zr, Y, Sr, Zn, Ni, V, Cr, Ga, Sc and Co) for all samples; the operating conditions for which are outlined in Table A3.2. SiO<sub>2</sub> Al<sub>2</sub>O<sub>3</sub> and MgO were analysed, but were rejected on the basis of low accuracy and precision. An initial small subset of pelleted samples and IRM samples were analysed to test and explore the X40 software. From this a suitable set of International Reference Materials were selected, and subsequently analysed using the calibration program of the Phillips X40 software. This program calculates the best straight line fit (y=mx+c) of counts against composition for each element. The operator may delete standards that distract the line to minimise the rout-mean-square (RMS) error of the regression in order to obtain an optimum calibration.

After calibration, a subset of standards were analysed as unknowns to test the quality of the calibration. Harvey & Atkin (1988) set out guide lines for acceptable errors on trace elements on IRM samples (Table A3.3), and using this guide, each element either passed or failed the test. The calibration lines of any problematic elements were re-examined and improved as far as possible. This procedure was carried through to all subsequent calibrations during the analyses on the XRF. It would have been preferable to analyses IRM samples that had not doubled as standards, however, with the limited number of suitable IRM samples available this was not possible, and therefore, the test probably gives a slightly optimistic estimate of the degree of accuracy.

Appendix 3	3
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Element	line	crystal	columator	angle	bkg (+)	bkg (-)	time (pk)	time
								(bkg)
Trace								
Nb	Ka	1	С	30.46	0.6	0.6	80	20
Zr	Ka	1	С	32.115	1.1	0.9	80	20
Y	Ka	1	С	33.91	0.9	0.8	80	20
Sr	Ka	1	С	35.875	1.1	1	80	20
Rb	Ka	1	С	38.01	0.7	0.92	80	20
Zn	Ka	1	С	60.6	1.2	0.8	80	20
Ni	Ka	1	С	71.32	2.2	1.3	80	20
V	Ka	1	F	123.315	2	2.2	80	20
Cr	Ka	1	С	107.31	2	2	100	20
Ga	Ka	1	С	56.215	0.9	0.26	80	20
Sc	Ka	2	С	97.74	2	2	100	20
Co	Ka	1	С	77.98	2	3.5	100	20
LREE								
Nd	La	1	С	112.8	1.72	0.38	200	200
La	La	1	С	139.075	3.54	2.84	200	200
Ce	Lb	1	С	111.685	0.68	2.42	200	200
Majors								
Fe	Ka	2	F	57.48	2.6	1.2	20	4
Ca	Ka	2	С	113.265	3.8	3.84	20	10
Na	Ka	6	С	28.1	2.92	1.14	100	40
К	Ka	2	С	136.85	5	3.8	40	10
Ti	Ka	2	С	82.265	3.8	1.24	20	4
Mn	Ka	1	С	95.4	4.64	1.38	20	10
Р	Ka	5	С	141.025	4	3.5	80	20

**Table 3.1:** XRF operating conditions for the analyses of powdered pellets for the acquisition of trace and major elements data. Abbreviations: bkg(+) = positive background; bkg(-) = negativebackground; time (pk) = time in seconds spent analysing the peak; time (bkg) = time spent analysingthe background.

During the collection of all the data analysed here, two IRM samples were analysed as unknowns (BHVO-1 and W-2) to monitor precision (i.e. machine drift, expressed as  $2\sigma$ ; Appendix 4, p. 308). Additionally, two unknowns were treated as internal monitors (samples A413 and A416). The IRM samples used as monitors were chosen on the grounds that they were of similar composition to the majority of basalts encountered in the research, and that their concentrations are well reputed and reliable (Geostandards Newsletter, 1987). The internal monitors were chosen because they are different to each other in composition and they cover the range of trace element compositions encountered in this research. Additionally, they are fresh and have abundant finely temaed agate powders available. Six pellets and four fusion discs were made of each internal monitors. The IRM samples were repeatedly analysed every 24 hours and the internal monitors at various intervals between these, the frequency of which depended on the performance of the machine at that time. It was usually after every five unknowns, but this varied between four and ten. This method and pattern of monitoring was also used for both the REE and the fusion disc major element analyses. During the latter, BIR-1 was substituted for BHVO-1 (except for the acquisition of Na<sub>2</sub>O and K<sub>2</sub>O which used BHVO-1 again).

Preparation of geochemical data set



**Figure A3.1:** Variation of internal monitors (International reference materials: BHVO-1 & W-2) during the analyses of 425 unknowns over a period of two and a half months. The new calibration indicates the interval at which the Rh tube was changed. The labelled horizontal lines indicate the "expected" concentrations of these elements for each of the IRM samples: a)  $TiO_2$ - a good example; b)  $Na_2O$  - a poor example.

The analyses of 425 unknowns (plus monitors and repeats) took place over a period of two and a half months during two main runs (referred to as R#1 and R#2). The session was interrupted at the end of R#1 by a Rh tube collapse and replacement. The machine was allowed to settle and was recalibrated under exactly the same conditions, as far as possible. The monitors (BHVO-1, W-2, A413, A416) demonstrate, however, that the precision and accuracy for a few elements were affected by the tube change (e.g.  $Na_2O$ ), whereas most were not (e.g.  $TiO_2$ ) (Figure A3.1).  $Na_2O$  was the worst affected, and this has been taken into account in the discussions and arguments presented in this thesis (Chapter 5).

Abundance of element (ppm)	% of error allowed
10	20
100	7
1000	<1

**Table A3.3:** Harvey & Atkin (1988) guide lines for absolute maximum errors acceptable for trace elements from IRM samples.

The LREE (La, Ce and Nd), Rb and Ba were analysed using a separate calibration set up (Table A3.2). REE for a subset of 21 samples have also been analysed by ICP-AES at the Royal Holloway and Bedford College, University of London (courtesy of K. Darke and J. Cann); and these and other trace elements were acquired for two samples by ICP-MS at the Durham laboratories (courtesy of Mark Wharton, Ian Parkinson and Julian Pearce). The results for the LREE from the XRF have high precision errors ( $2\sigma$ ) (significant for some samples with lower concentrations) but a comparison of data acquired by ICP-AEC and XRF for 21 samples shows that the difference between these two methods is minimal (Figure A3.2).



Figure A3.2: A comparison of La, Ce and Nd analysed by XRF and ICP-AES.

#### Fusion disc analyses using X40 software

The fusion disc analyses ran without any mechanical or electrical problems. The operating conditions are summarised below in Table A3.4.

Element	Line	crystal	columator	Angle	bkg (+)	bkg (-)	Time	Time
							(pk)	(bkg)
Si	Ka	3	С	109.175	4.8	4.2	20	10
Al	Ka	3	С	145.23	1.4	5.1	20	10
Fe	Ka	2	F	57.48	2.6	1.2	20	10
Mg	Ka	6	C	23.3	3.2	2.48	100	40
Ca	Ka	2	С	113.265	3.8	3.84	20	10
Na	Ka	6	С	28.1	2.92	1.14	200	80
K	Ka	2	С	136.85	5	3.8	40	20
Ti	Ka	2	С	86.265	3.8	1.24	20	10
Mn	Ka	1	C	95.4	4.64	1.38	20	10
P	Ka	5	С	141.025	4	3.5	80	40

**Table A3.4:** XRF operating conditions for the analyses of fusion disc major element oxide data. Abbreviations are the same as in Table A3.2:

#### A3.4) Fusion disc preparation methods: an investigation

As an independent check on the quality of the major element oxide data from fusion discs, twelve discs were made and analysed at a separate institution; the University of Leicester. The differences in the major element abundances for the samples and the differences in the method of preparation discovered by visiting Leicester initiated an investigation into the details of the preparation procedure. The fusion discs methods of Durham (based on that of Edinburgh) and Leicester differed in two ways:

1) The Leicester method fuses ignited powders whereas the Durham/Edinburgh method uses hydrous powders. For the latter, the crucible with 1:5 ratio of powder to flux would be fused for 3-5 minutes, removed from heat, cooled, weighed and the loss in weight would be considered as the loss on ignition. The weight would be reimbursed with flux before fusing was continued. Later, Durham modified the technique by not replacing the lost weight. This will affect the ratio of flux to powder and consequently produce lower count rates and low accuracy.

2) Leicester use a light flux, the Johnson Matthey Tetraborate Spectroflux (100B) and Durham used a heavy La-oxide Spectroflux (105). The latter has been known to cause problems in analysing elements such as Na and other heavy absorbers (N. J. Marsh, University of Leicester, pers. comms., 1989). The Na K alpha peak overlaps with the La L lines and measurement of this element can be problematic when using the 105 flux (Day, 1989).

#### Appendix 3

The knowledge of these differences raised the question of how important the accuracy of any one method is and what effect these differences have on the final data produced? With the assistance of R. Hardy, and geochemical technicians at Durham, an experiment was constructed and carried out to try to answer these questions. Four IRM were chosen, and the three methods of preparation applied to each. The discs were analysed on the XRF and the counts processed by the CBI software (the only software installed at Durham at the time). The flux type had the most dramatic affect on MnO,  $TiO_2$ , CaO and FeO(t). These results demonstrated that the difference in technique had significant effects on the data, and based on this conclusion and the general logic behind the Leicester method, the Durham geochemical laboratories made the decision to convert to the Leicester method.

#### Part B:

#### Microprobe analyses

Volcanic glass and minerals were analysed by electron microprobe using a Cambridge Instruments Geoscan model at the University of Durham. For the glass analyses, the abundances of  $Cr_2O_3$ , MnO and  $P_2O_5$  were below the detection limit, as assessed by the random scatter of this data when it is plotted against the Mg # of the glass (Figure A3.3).



**Figure A3.3:**  $Cr_2O_3$  plotted against Mg # for volcanic glass samples.  $Cr_2O_3$  is one of the three elements ( $Cr_2O_3$ ,  $P_2O_5$  and MnO) that is thought to be below detection limit of the microprobe, estimated from the random scatter of the data points in this figure.

Compositions of feldspar crystals from a xenolith (unit S12) were analysed on the microprobes at the University of Manchester (courtesy of Mark Wharton). These are indicated accordingly in Appendix 2-A.

## Part C: CIPW norms

In this thesis, CIPW-normative compositions have been calculated for 47 fresh whole-rock samples and 24 glass samples from the Hengill to Ingolfsfjall region, and 2 whole-rock samples from the Reykjanes Ridge ( $63^{\circ}10'$ N). This was done using the computer programs "NORM" and "CIPWNORM" from the IGPET1 software package (Carr, 1985). The Fe<sub>2</sub>O<sub>3</sub>/FeO ratio was assumed to be 0.15 in all calculations (Irvine & Baragar, 1971).

#### Part D:

#### **Chondrite normalising values**

Element	Value	Reference
La	0.310	1
Ce	0.808	1
Pr	0.122	1
Nd	0.600	1
Sm	0.195	1
Eu	0.0735	1
Gd	0.259	1
Dy	0.322	1
Ho	0.0718	1
Er	0.210	1
Yb	0.209	1
Lu	0.0322	1
Ba	6.9	2
Rb	0.35	2
К	120	2
Nb	0.35	2
Sr	11.8	2
Р	46	2
Zr	3.87	3
Ti	620	2
Y	2.0	2

### References

1 = Boynton (1984)

2 = Thompson *et al.* (1984)

3 = Sun & McDonough (1989)

Appendix 3

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## Appendix 4 Whole-rock geochemical data set

## **Explanation**

The following appendix contains the geochemical data set for the whole-rock samples. The analyses are grouped primarily in stratigraphic order for the three volcanic systems. The intrusive rocks are grouped separately at the end.

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0	Bjarnarfell (B)	311
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## Key to Appendix 4

= stratigraphic unit or lava shield, as described in Chapter 3.
= refers to one of four varieties, as defined in Figure 2.11 (p.30). Those dykes
that do not belong to one of these groups have titles presented in brackets. If
this title is the name of an eruptive unit, it is inferred that the intrusive material
is therefore of a similar basalt type and is associated with that eruptive unit.
= refers to the map location from where the sample was collected from.
= the data set has been compiled by various analytical techniques (Figure 5.1
and Appendix 3), and only data acquired by the most accurate method are
presented in this thesis. The data source therefore refers to the method used to
acquire the data for any particular sample. Major elements from fixed discs are marked
= refers to petrological types, as defined in Chapter 4 (Table 4.3, p. 87)
= refers to the variety of element mobility detected in the sample.

#### Appendix 4

Abbreviations and explanations for the variables explained above are summarized in Table A4.1 below. The Mg # for data acquired from fusion discs are quoted to 1 decimal place, whereas the Mg # from powder pellet are not quoted to 1 decimal place, and are provided only as a rough estimation of the true Mg #. Estimates of analytical error, expressed as 20, are listed in Table A4.2.

Information type	Abbreviation	Meaning
Data source - Major	Р	- Powdered pellet XRF data
-	F	- Fusion disc XRF data
- REE	D	- REE data from the XRF at Durham
	Е	- ICP-AES data from RHBC, London University
	<u>M</u>	- ICP-MS data from Durham University
Petrological type	1a	- Aphyric with olivine in the groundmass
	1b	- Aphyric with no olivine in the groundmass
	2a	- Feldspar megacrysts in significant proportions
	2b	- Feldspar megacrysts + pyroxene
	2c	- Feldspar megacrysts + pyroxene + olivine
	3a	- Olivine and feldspar phyric with radiating
		glomerophyric clusters
	3Ъ	- Olivine phenocrysts only
	<u>3c</u>	- Radiating glomerophyric clusters of feldspar only
Alteration type	<b>A</b> 1	- Haematite and other secondary opaques minerals causing scatter in $EeO^*$ V and Co
	A2	- Mobility of CaO, Na-O, K-O, Sr. Rb, Ba, La and Ce
	A1+2	- Alteration of both types A1 and A2 above

 Table A4.1: Abbreviations used in whole-rock data set.

Major	Error	Error	Trace	Error
	F	P		P
SiO2	0.53	(1.78)*	Nb	1.41
Al2O3	0.14	(1.01)#	Zr	6.22
FeO *	0.16	0.22	Sr	10.05
MgO	0.23	(i · 05) <b>¥</b>	Zn	8.64
CaO	0.21	0.63	Ni	12.03
Na2O	0.25	0.27	<b>v</b>	9.84
K20	0.00	0.02	Cr	31.2
TiO2	0.03	0.12	Ga	3.36
MnO	0.01	0.04	Sc	4.88
P2O5	0.02	0.01	Co	2.55
	1	ļ	Rb	2.31
			Y	2.84
			Ba	21.1
			Ce	8.46
			La	5.22
			Nd	4.37

\* These three elements are <u>not</u> used in this thesis

**Table A4.2:** Precision errors (20) for major and trace elements, for samples analysed by XRF (University of Durham). F = fusion disc; P = powdered pellets; FeO \* = total iron. The errors are derived from replicate analyses on four samples, where n=79 for powdered pellets and n=15 for fusion) discs.

Whole-rock data set - Hveragerői Volcanic System

				*				*		*	
Sample #	<b>A5</b> 4	A403	A405	A522	A537	A525	A526	A406	A407	A529	A240
Unit	GA 1	GA 4	GA 4	<b>GA</b> 5	GA 6	GA 9	GA 9	GA 10	GA 10	GA 11	GA 12
Map ref	B1-8	B1-8	B1-8	B1-8	B1-8	B1-8	B1-8	B2-8	B1-7	B1-7	B2-8
0.00	1	(			6.0	· Go a	110 0		(10 N		LIDA
SICZ	(47.0)	(49.0)	(48.0)	47.97	(48.9).	(44 · · · ·	(48·L)	47.45	(48.8)	47,64	(48.5)
	(15.5)	(4.4)	$(14.1)^{-1}$	13.58	(15.+)	(14-1)	· (14·1)	13.72	(14-6)	13.07	(14.5)
reu!*	12.24	14.60	14.00	14.25	15.04	14.73	13.80	14.31	14.01	14.28	14.29
MgU	(12.0)	(0.9)	(0.1)	7.23	(10.9).	(10.5)	(9.0)	1.34	(4.8)	7.59	(8.9)
	11.28	11.70	11.09	11.01	102	1.07	0.00	0.07	0.19	0.00	11.30
	1.75	2.05	2.11	2.34	1.80	1.95	2.22	2.21	2.10	2.20	2.74
	1.42	1.06	1.02	1 01	1.08	2.04	1.02	1.05	1 09	1.02	1.02
MaQ	0.12	0.01	1.85	1.91	1.84	2.03	1.80	1.85	1.80	1.82	1.85
	0.10	0.21	0.23	0.22	0.22	0.22	0.21	0.22	0.22	0.21	0.22
Tatel	(1 a ( d)	0.17 (101 m)	0.14	0.10	(a) A	(131.2)	(101 2)	0.19	(10 T)	0.19	0.21
	(101.6)	(104.9)	(104.0)	.0.20	(1045)	104.5	(0)	0.61	(105.+)	99.10 0.77	(102.4)
				-0.20				-0.01		0.77	
Mg#	(67)	(61)	(61)	51.5	(60)	(60)	(58)	51.8	(58)	52.7	(56)
Nb	7	12	12	12	12	13	11	13	12	11	14
Zr	54	86	96	92	93	98	94	93	96	90	106
Y	21	28	31	32.1	31	33	32	31	33	31.3	32
Sr	159	151	157	155	148	163	160	158	159	148	185
Zn	<b>8</b> 5	99	111	107	109	107	96	104	112	106	100
Ni	119	93	<b>96</b>	<b>8</b> 5	90	96	82	91	90	87	67
V	292	402	419	399	407	411	365	414	456	403	386
Cr	271	164	155	147	155	150	133	144	153	145	139
Ga	20	17	19	19	20	20	19	19	21	17	19
Sc	34	48	48	42	42	44	38	47	45	48	45
Co	56	64	<b>6</b> 6	67	<b>6</b> 5	<b>6</b> 6	62	67	64	<b>6</b> 6	64
Rb				1.8				2.3		0.0	
Ba				40				43		49	
La				7.8				6		7.5	
Ce				19.3				16		18.8	
Pr				2.8						2.7	
Nd				14.8						14.2	
Sm				4.1						3.9	
Eu				1.5						1.5	
Gd				5.3						5.1	
Dy				6.2						6.0	
Но				1.2						1.1	
Er				3.5						3.3	
Yb				3.3						3.2	
LU				0.5						0.5	
Data source	B										
Majors	Ρ	Р	Р	F	Р	Ρ	Ρ	F	Р	F	Ρ
REE				Е				D		Ε	
Pet. type Alteration	1a	3a	3a	1a	1a	1a	1a	1a	1a	1a	1a

. -

	*					*					*
Sample #	A28	A27	A29	A30	<b>A</b> 5	<b>A55</b> 0	<b>A55</b> 1	A32	<b>A3</b> 1	<b>A4</b> 5	A33
Unit	GB 1	GB 1	GB 2	GB 3	GB 4	GB 4	GB 4	GB 6a	GB 6a	GB 6a	GB 6b
Map ref	B3-8	B3-8	B3-8	B3-8	<b>B3</b> -5	B3-4	<b>B3-4</b>	B3-8	B3-8	B3-4	<b>B3-4</b>
SiO2	47.30	(48.0)	(48.2)	(47.4)	(48.6)	48.08	(48.6)	(46.6)	(47.))	(46.7)	48.51
AI203	14.94	(14.6)	(15.3)	(16.2)	(16.7)	15.63	(15.3)	(14.9)	(14.8)	(14.8)	13.47
FeOT*	12.23	11.98	11.41	11.62	10.11	9.66	9.45	11.36	12.11	11.02	13.76
MgO	8.82	(9.4)	(8.5)	(9.9)	$(\mathbf{q}\cdot7)$	10.69	(8.9)	(7.1)	(7.8)	(7.6)	6.98
CaO	12.61	12.45	12.72	11.94	13.72	13.38	13.69	11.75	12.28	12.21	11.83
Nazu	1.84	1.19	2.24	1.87	1.75	1.52	1.42	2.58	2.38	2.10	2.38
K20	0.05	0.14	0.09	0.07	0.01	0.02	0.04	0.21	0.10	0.13	0.20
1102	1.31	1.32	1.15	1.10	0.60	0.62	0.63	1.42	1.47	1.37	1.91
MINU	0.19	0.19	0.10	0.17	0.17	0.15	0.15	0.18	0.19	0.18	0.21
P200	0.11	0.10	0.10	0.10	0.02	0.04	0.04	0.15	0.12	$\left  ac z \right $	0.10
	99.37	(49.4)	(44.4)	(100-5)	(161.4)	99.76	(48.2)	(96.2)	(48.5)	(96-5)	89.39
LUI	0.40					-4.79				-0.10	-0.45
Mg#	60.2	(62)	(61)	(64)	(67)	<b>6</b> 9.9	(66)	(57)	(57)	(59)	51.5
Nb	7	6	7	7	3	2 ′	2	7	7	7	11
Zr	62	62	52	52	29	30	29	66	62	71	100
Y	23	23	21	20	17	15.6	15	26	24	24	29
Sr	143	155	135	136	106	93	105	157	156	168	177
Zn	78	80	76	79	61	64	63	73	79	73	93
Ni	128	114	124	142	170	<b>19</b> 5	168	99	119	103	52
V	288	309	291	289	220	216	236	292	314	275	363
Cr	184	194	205	197	428	442	527	214	254	260	146
Ga	14	14	16	16	15	16	14	14	14	19	18
Sc	42	42	43	41	45	44	44	38	38	40	41
Co	59	57	55	58	48	48	44	55	58	51	63
Rb	0.5					1.8					2.0
Ba	42	•	•	•		9		-	•	•	87
	1	6	8	6		1.9	4	5	6	6	8
Ce Dr	10	9	5	9		3.6	-1	12	11	13	26
Pr Na	•	F	•	•		0.5	•	•	•	40	40
NG Sm	0	5	o	0		J.4 1 2	U	9	9	12	10
En						1.3					
Gd						21					
Dv						2.1					
Ho						06					
Er						17					
Yb						1.7					
Lu						0.3					
Data source											
Majors	F	Р	Ρ	Р	Ρ	F	Р	Ρ	Ρ	Р	F
REE	D	D	D	D		Ε	D	D	D	D	D
Pet. type Alteration	Зb	Зb	1a	3a	<b>3a</b>	<b>3</b> a	3a	2c	2c	2c	2c

				*		+	*	*			*
Sample #	A555	A553	A538	A539	A540	A523	<b>A54</b> 1	A524	A237	A238	4398
Unit	GB 6b	GB 6b	GC 1	GC 1	GC 2	BA 2	BA 3	BA 3	BA 3	BA 3	BA 3
Map ref	B3-8	B3-8	B3-3	B3-3	B3-3	B1-7	B3-3	B1-7	B2-8	B2-8	R2-8
			200	200		5	000	0.1			
SiO2	(49.1)	(50-8)	(48.0)	47.23	(48.7)	47.66	45.95	46.88	(47.2)	(46.9)	47.41
AI203	(11.2)	hite in	(13.9)	13.57	(14.1)	13.60	13.78	14.49	(1, -2)	(C A	15.17
FeOT*	13.80	13.99	14.93	14 49	14.52	14 27	13.86	12 41	12 71	11 69	11 34
MaQ	(7.1)	(2.7)	(a, c)	7.09	(8.6)	7.31	8.34	8 91	60.0	(1,2)	8 68
CaO	11.38	12 13	11.85	11 70	11 80	11 70	10 54	11.80	12 17	12 66	12 54
Na2O	2 79	2 45	2 21	234	2 49	2 12	2.30	1.81	1 90	1 67	1 00
K2O	0.25	0.10	0 10	0.16	014	0.09	0.31	0.14	0.07	0.06	0.18
Tim	2 10	2 03	2.02	2 15	2 00	1 04	235	1.62	1 65	1 13	1 20
MnO	0.23	0.23	0.22	0.21	0.22	0.21	0.20	0.10	0.10	0.19	0.10
D2OE	0.20	0.20	0.17	0.21	0.10	0.21	0.20	0.10	0.18	0.10	0.18
Tetal	(a) 2)	0.14	(1,1,1)	00.12	0.10	0.20	07.05	0.10	U.20		0.10
	(01.2)	(1024)	(05.0)	39.13	(102.8)	89.07	87.00	80.00	(101.2	)((0))))	4 51
				0.01		0.34	-0.42	-0.02			1.51
Mg#	(52)	(54)	(57)	50.7	(55)	51.8	55.8	60.1	(64)	(67)	61.6
-											
Nb	13	11	14	13	13	12	18	13	12	8	10
Zr	115	104	97	98	<b>9</b> 3	<b>9</b> 5	127	90	<b>93</b>	61	72
Y	34	33	32	32	30	31	30	28	27	20	21
Sr	190	178	199	197	191	158	249	165	171	148	178
Zn	99	104	108	103	102	111	108	94	91	83	85
Ni	80	<b>6</b> 6	92	86	84	84	147	170	141	161	169
V	<b>400</b>	395	403	431	391	409	351	331	313	273	267
Cr	146	169	94	91	89	150	353	408	345	388	394
Ga	21	18	20	16	18	18	19	19	17	17	17
Sc	42	44	44	45	45	42	41	42	43	40	44
Со	63	63	68	64	65	67	66	61	60	56	57
Rb				1.2		21	4.2	1.4			2.8
Ba				50		37	<b>8</b> 5	54			54
Le	10	6	8	10	9	6	11	5		5	3
Се	32	21	19	23	20	17	14	19		14	13
Pr											
Nd	20	13	12	14	14		17			8	
Sm											
Eu											
Gd											
Dv											
Ho											
Er											
Yb											
Lu											
Data source	e										
<b>Ma</b> jors	P	Р	Р	F	P	F	F	F	Р	Ρ	F
REE	D	D	D	D	D	D	D	D		D	D
Pet. type Alteration	2c	2c	1a	1a	1a	1a	<b>3a</b>	<b>3a</b>	3a	3a	<b>3</b> a

						*			*		
Sample #	A399	A400	A547	A548	A549	A530	A46	A545	A546	A35	A517
Unit	BA 3	BA 3	BB 4	BB 4	BB 4	BB 4	<b>BB</b> 5	BB 5	BB 5	BB 6	<b>BB 6</b>
Map ref	B2-8	B2-8	B3-3	B3-3	B3-3	<b>B3-1</b>	B3-4	<b>B3-3</b>	B3-3	B3-8	B3-8
•											
SiO2	(49.8)	(49.1)	(47.8)	(47.3)	(47.5)	45.38	(49.8)	(49.7)	48.03	(47.9)	(46.7)
A1203	(13.9)	(11.3)	(14.2)	(16.6)	15.0	14.21	(14.0)	(13.0)	12.83	(16·1)	(15.1.)
FeOT*	11.56	11.71	13.96	13.72	12.31	12.71	15.15	16.03	17.22	13.34	12.66
MaO	(8.8)	(9.2)	(7.1)	(7.9)	(8.7)	10.43	(7.2)	16.1)	5.23	(9.0)	(a.2)
CaO	13.38	13.18	11.74	11.63	12.29	12.16	11.37	9.99	9.77	11.82	11.43
Na2O	1.65	1 75	2 70	2.57	2.23	1.54	2.91	3 29	2.66	2.50	2 37
K20	0.08	0.12	0.16	0.16	0.09	0.11	0.25	0.47	0.45	0.17	0.10
TiO	1 4 4	1 37	2 22	2.07	1 48	1 54	2 38	3.00	3.52	1.67	1 53
MaO	0.20	0.10	0.21	0.21	0.10	0.20	0.24	0.00	0.02	0.01	0.20
	0.20	0.18	0.21	0.21	0.18	0.20	0.24	0.27	0.25	0.21	0.20
Tatal	0.14	0.11	0.22	0.21	0.15	0.10	0.20	0.30	0.45	0.18	0.17
	(100-9)	(101-0)	(100.3)	(100.3)	(100.1)	86.43	(103-6)	)(102.4)	100.41	(103-5)	(998)
LUI						2.23			-0.80		
Mg#	(61)	(62)	(52)	(55)	(60)	63.3	(50)	(44)	38.9	(60)	(60)
•		~ /					- /	•			
Nb	11	10	15	14	10	10	18	30	31	11	11
Zr	78	75	107	101	72	80	123	194	209	88	81
Y	26	20	34	30	25	29	35	51	49	27	25
Sr	187	180	198	196	157	126	203	231	243	170	161
7n	94	87	101	97	88	92	111	123	184	93	87
Ni	128	154	85	115	170	228	50	31	37	222	212
V	318	200	365	348	207	330	441	427	07 AGE	207	210
· ·	267	260	343	040	201 AEE	604	76	421	400	507	280
	307	348	40	200	400	40	70	40	50	507	340
	19	17	10	10	10	10	21	17	20	14	17
30	40	41	43	42	42	40	40	41	39	41	38
	52	54	62		5/	64	70		48	65	63
HD						3.3			6.1		
13 <b>8</b>			_	-	_	33			586	_	_
			7	6	5	6	11	20	113	9	7
Ce			25	24	16	14	25	47	280	23	18
Pr									33.53		
Nd			17	14	7		13	33	201	14	15
Sm									42.83		
Eu									15.77		
Gd									<b>56.25</b>		
Dy									<b>6</b> 5.51		
Но									12.81		
Er									38.56		
Yb									41.51		
Lu									5.80		
<b>_</b> .											
Data sourc	;e										
Majors	P	Ρ	Ρ	P	P	F	Р	Р	F	Р	Ρ
REE			D	D	D	D	D	D	D	D	D
Det tors	0-	0-	0L	0L	<b>0</b> L	<b>6</b> L	4 -	4 -		-	-
Alteration	38	38	30	30	30	30	18	18	18	30	3b

		- ★									
Sampie #	A516	A34	A515	A514	A513	<b>A5</b> 0	A49	A48	<b>A4</b> 7	<b>A</b> 42	<b>A3</b> 7
Unit	BB 6	BB 6	BB 6	BB 6	BB 6	BB 6	BB 6	BB 6	BB 6	BB 7	BB 7
Map ref	B3-8	B3-8	B3-8	B3-8	B3-8	B3-4	<b>B3-4</b>	B3-4	B3-4	B3-8	B3-8
SiO2	(48.5)	44.73	(47.q)	(47.1)	(48.2)	(46·4)	(47.9)	(456)	(47.3)	(48.8)	(48.9)
A1203	(16.5)	13.06	(16.2)	(IE.A)	(161)	(154)	(14.8)	(14.7)	(15.4)	(11.9)	(2.8)
	13 20	12.00	12 79	11.99	12.46	13.17	13.55	13.06	13.07	17.08	16.31
MaO	10.20	14 47	(98)	(8.5)	(0.11)	(9.0)	(8.0)	(8·4)	(8.3)	$(\varsigma \cdot z)$	(5.5)
my Co	( 4.%) 11 74	10.21	11 95	12 19	12.27	11.17	12.03	11.53	11.55	9.54	9.40
	2 20	1 50	2 20	2 28	2.26	2.47	2.51	2.29	2.51	3.37	2.92
Nazu Kao	0.11	0.10	0.07	0 10	0 10	0.14	0.13	0.15	0.16	0.65	0.53
	1.62	1 21	1.52	1 44	1 43	1 77	1.92	1.80	1.76	3.92	3.35
1102	0.01	0.10	0.20	0.10	0.10	0.21	0.21	0.20	0.20	0.31	0.26
MnO	0.21	0.19	0.20	0.13	0.13	0.21	0.17	0.17	0.19	1.01	0.52
P206	0.15	0.12	0.14	(a 2)	(102.0)	(ag g)	$(\alpha, \beta)$	(92.9)	(100.5)	101.8	(100.4)
Total	(104.2)	98.59	(102:9)	(49.5)	(102.5)	(110)		(17-1)	(	(	
LOI		-0.38									
Mg#	(61)	70.2	(62)	(60)	(61)	(59)	(55)	(57)	(57)	(41)	(41)
Nb	11	8	10	8	9	14	14	13	12	44	34
Zr	84	65	77	71	71	93	103	98	96	333	<b>2</b> 67
Y	27	23	26	26	26	29	30	28	29	61	56
Sr	160	137	158	152	150	177	186	183	176	306	249
7n	93	84	91	84	84	89	<b>9</b> 9	93	90	157	143
Ni	234	381	242	180	218	206	128	140	185	21	23
v	299	254	294	285	295	<b>29</b> 5	316	303	305	306	378
Cr	617	997	552	494	559	512	343	377	493	23	<b>56</b>
Ga	17	14	19	16	17	18	18	19	17	23	20
Sc.	41	35	41	39	40	38	38	35	36	31	34
Co	67	70	62	57	59	69	<b>6</b> 6	64	68	68	74
Rh	0.	2.5	-		•						
Re		51									
i e	5	7	7	8	8	7	7	7	9		22
Ce	14	15	14	15	11	19	24	23	19		56
Dr	14		••								
Nd	٩	10	9	15	11	10	14	16	14		33
Sm	Ŭ		•								
SHL -											
Cd											
Ly Lo											
nu E-											
				•							
111											
Data sour	ce				_	_	_	-	-	-	~
<b>Majors</b>	Р	F	P	Ρ	Ρ	P	P	P -	4	۲	4
REE	D	D	D	D	D	D	D	D	D		D
Pet. type Alteration	3b	Зb	Зb	3b	Зb	3b	<b>3</b> b	3b	3b	1a	18

	*										
Sample #	A36	<b>A67</b>	A74	<b>A8</b> 4	A73	A109	A242	A107	A69	A70	A75
Unit	BB 7	<b>TA 1</b>	TA 1	<b>TA</b> 1	TA 1	TA 2b	TA 3	TA 3	TA 3	TA 3	TA 3
Map ref	<b>B3-</b> 8	B2-4	B2-4	B2-3	B2-8	B2-7	B2-3	B2-7	B2-8	B2-8	B2-3
<b>S:</b> 00	47 00	(52.4)	(11.0)	(10.2)	000	(1.6.2)	(10 -1)	114 2	(107)	lina	(1.0 Z)
5102	47.20	(5LL)	(44.0)	(48.5)	(450)	(44.0)	(48.+)	(4+++)	(48.7)	(43.7)	(47.1)
AI2U3	12.40	(10.6)	(15 - 4)	(12.6)	(14.5)	(13.0)	(13.7)	(10.7)	(13.6)	(13.5)	(15.8)
Feur	17.42	11.30	15.30	15.33	17.30	10.00	14.3/	12.92	14.47	14.55	14.70
MgU	4.07	((0.5)	(13.4)	(9.9)	(18.6)	(11q)	· (9·6)	·[9·5)	(9.5)	(40)	(9.7)
	9.32	15.65	9.97	12.12	8.79	10.69	11.56	10.44	11.59	11.84	11.74
Nazo	2.95	1.28	2.92	1.99	1.15	1.78	2.15	1.42	2.19	2.21	2.03
K20	0.69	0.08	0.07	0.10	0.07	0.07	0.12	0.05	0.10	0.11	0.09
TIO2	3.98	1.27	2.44	2.13	2.05	2.21	2.18	2.16	2.27	2.35	2.14
MnO	0.28	0.18	0.25	0.23	0.22	0.23	0.21	0.25	0.22	0.23	0.21
P2O5	1.12	0.07	0.26	0.22	0.18	0.24	0.23	0.26	0.23	0.26	0.23
Total	100.08	(103.0)	(104.0)	(102.8)	(107.9)	(105.6)	(102.8)	(100.8)	(102.9	)(1029)	) (104:4)
LOI	-0.11										1.51
Mg#	36.0	(66)	(65)	(57)	(69)	(60)	(58)	(61)	(58)	(56)	(58)
Nb	48	6	18	15	15	16	20	16	19	19	18
Zr	371	56	136	115	117	115	126	121	130	132	128
Y	60	21	25	32	23	33	32	33	34	37	34
Sr	341	138	160	191	152	242	189	213	204	189	197
Zn	213	79	115	102	117	135	107	113	102	106	109
Ni	14	99	90	66	80	64	88	83	83	69	75
V	287	358	462	415	426	451	401	432	400	410	397
Cr	14	229	231	117	137	104	172	214	172	179	183
Ga	27	20	17	18	20	16	19	17	16	16	19
Sc	33	48	46	45	45	48	42	48	44	46	42
Co	55	48	75	68	84	72	63	49	67	66	64
Rb	14.8			~					0/		04
Ba	761										
1.8	175										
Ce	443										
Dr.	53 46										
Nd	304										
Sm	65.36										
Fu	22.04										
Gd	80 18										
Dv	82 04										
Ho la	15.02										
Er	13.02										
Ci Vh	44.12									*	
1	40.24 6 00										
	0.00										
Data source											
Majors	F	Р	Р	Ρ	Ρ	Ρ	Р	Р	Ρ	Ρ	Р
REE	D										
Pet. type	1a	1a	1a	1a	1a	1a	1a	1a	1a	18	18
Alteration		A1+2	A2		A1+2	A1+2		A1+2			

Sample #	A102	<b>A212</b>	<b>A226</b> TA 3/4	<b>A227</b> TA 3/4	<b>A531</b> TA 3/4	<b>A432</b> TA 3/4	<b>A257</b> TA 4	★ A247 TA 4	<b>A428</b> TA 4	<b>A8</b> 0 TA 4	A250
Map ref	B2-7	B2-7	B2-6	B2-6	B2-1	B2-1	B2-3	B2-3	B2-2	B2-3	B2-3
SiO2	(475)	(49.5)	(49.3)	(48.8)	(48.5)	(49.6)	.(49.0)	48.56	(41.6)	(49.0)	(49.5)
A1203	(13 0)	(12.8)	(13·4)	(13.8)	(13.4)	(12.1)	(13.2)	13.29	(9.8)	(12.4)	(12·7)
FeOT*	14.63	14.49	14.16	14.41	15.26	15.27	14.92	13.59	14.48	14.04	15.52
MgO	(10.0)	(8-6)	(g.7)	. (10.1)	£11·9)	.(9.5)	(0.8)	5.76	(10.4)	(9.3)	(9.6)
CaO	11.49	12.42	11.80	11.48	11.35	12.99	11.7 <del>9</del>	12.00	18.35	12.61	11.92
Na2O	2.25	2.05	2.15	2.27	1.44	1.64	1.83	2.11	1.40	1.89	2.13
<b>K2O</b>	0.17	0.11	0.18	0.12	0.09	0.05	0.09	0.22	0.06	0.16	0.16
TiO2	2.28	2.18	2.05	2.06	2.06	2.42	2.25	2.16	2.76	2.17	2.17
MnO	0.22	0.22	0.20	0.20	0.21	0.17	0.22	0.22	0.35	0.24	0.25
P205	0.25	0.25	0.24	0.23	0.24	0.20	0.24	0.25	0.26	0.22	0.19
LOI	(101.8)	(1026)	((03.2)	(103.4)	((044)	(105.4)	0.35	98.15 2.93	.[99•4)	(102.()	((04-1)
Mg#	(59)	. (55)	(59)	(60)	(62)	(57)	(60)	47.1	(60)	(58)	(57)
Nb	20	18	17	18	16	19	18	18	21	18	15
Zr	136	130	123	125	128	124	122	129	143	126	118
Y	35	33	33	32	33	34	33	31.3	39	33	36
Sr	194	193	192	190	168	212	208	188	336	189	185
Zn	106	112	105	105	115	114	103	107	145	108	116
Ni	75	80	80	92	70	76	92	77	79	92	68
V	383	397	391	397	429	452	396	394	527	397	436
Cr	162	170	180	169	89	118	159	186	156	189	93
Ga	19	17	19	19	20	10	19	19	18	17	19
50 Co	42 69	41 61	40 57	41 62	44 67	49 67	39 67	41 81	50 50	44 62	40 67
Dh	00	01	57	ΨZ	07	07	07	25	JZ	UZ.	07
Ra								64			
La		10						11.2		10	
Ce		32						27.1		31	
Pr								3.7		•	
Nd		16						17. <del>9</del>		18	
Sm								4.6			
Eu								1.6			
Gd								5.6			
Dy								6.1			
Но								1.1			
Er								3.3			
Yb								3.1			
								0.5			
Data sourc	e	<b>C</b>	-	•		-	-	-	-	<b>n</b>	-
Majors	۲	ч р	۲	۲	۲	۲	۲	r	۲	۲ D	٣
	4 -	U	4 -	4 -	4-	4-	4-	E 4-	4-	U	46
Per. type Alteration	18	18	18	18	18	18 A2	18	18	18 A2	18 A2	ID

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Sample #	A251	<b>A88</b>	A127	A128	A129	A130	<b>A8</b> 5	A256	A92	<b>A96</b>	<b>A96</b>
Unit	<b>TA 6</b>	TA 6	TA 7	TA 7	TA 7	TA 7	<b>TA 7</b>	TA 9	TA 9	TA 9	TA 9
Map ref	B2-3	B2-3	B2-3	B2-3	B2-3	B2-3	B2-3	B2-3	B2-3	B2-3	B2-3
SiO2	(49.4)	(49.5)	(48-9)	· (48·5)	(49.2)	(48.6)	(48.5)	(48.1)	(48·0)	(48.7)	(49.5)
A1203	(13.2)	(12.9)	(13.7)	(13.8)	(13.5)	(12.2)	(13.4)	(13.3)	(12.6)	(13.2)	(12.1)
FeOT*	15.10	15.10	14.78	15.29	14.83	17.99	15.18	15.31	16.10	15.38	15. <b>6</b> 2
MgO	(9.2)	(9.6)	(8:4)	(9.8)	(9.5)	(q.8)	(10.1)	(9·5)	(10.5)	(8.8)	(8.8)
CaO	11.40	11.60	11.54	11.21	11.44	10.79	11.45	11.06	11.55	11.39	11.47
Na20	2.34	2.19	2.53	2.36	2.28	2.24	2.06	2.31	2.26	2.40	2.07
К2О	0.16	0.14	0.18	0.19	0.21	0.13	0.19	0.27	0.20	0.19	0.22
TiO2	2.00	2.00	2.30	2.21	2.26	2.66	2.27	2.27	2.37	2.3 <del>9</del>	2.37
MnO	0.23	0.23	0.23	0.25	0.25	0.29	0.23	0.23	0.25	0.25	0.25
P2O5	0.23	0.1 <del>9</del>	0.20	0.22	0.19	0.28	0.20	0.21	0.22	0.24	0.23
Total	(103.3)	(10 <b>3</b> .4)	(102.8)(	(103.8)	(103-6)	(105.10)	(1036)	(102.6)	( 103.9)	(102.9)	(102.6)
Mg#	<b>(56)</b> .	(57)	(54)	(57)	(57)	(53)	(58)	(57)	(57)	(55)	(54)
Nb	15	14	18	16	17	20	19	18	19	19	18
Zr	117	111	119	116	114	147	123	127	127	126	128
Y	33	34	31	32	32	38	33	32	34	34	33
Sr	181	183	225	200	208	215	212	211	207	214	218
Zn	112	107	105	105	110	123	109	111	114	114	105
Ni	63	67	78	75	70	49	74	63	73	70	63
V	428	420	425	416	441	507	429	426	430	420	438
Cr	81	97	116	80	94	67	113	100	127	103	115
Ga	20	18	17	17	20	21	19	16	19	19	20
Sc	41	45	42	40	46	39	42	40	42	42	47
Co	64	71	72	67	64	84	69	66	69	66	68
Rb											
<b>B8</b>											
PT NH											
nu Sm											
SIII											
Gd											
Но											
Er											
Yb											
Lu											
Data sourc	e										
<b>Ma</b> jors REE	Ρ	Ρ	P	Ρ	Р	Р	Ρ	Ρ	Ρ	Ρ	Р
The	41	4 L	4.5	4 -	4 -	<b>.</b>	<b>A</b> .	<b>A</b> .	4	4	
Alteration	10	10	18	18	18	18	18	18	18 A1	18	18

Sample #	A190	A299	A401	A402	G46	<b>G60</b>	A404	A394	A392	A123	A192
Unit	TA 10	TA 10	TA 10	TA 10	TB 1	TB 1	TB 1	TB 1	TB 1	TB 1	TB 1
Map ref	B1-8	B2-8	B1-8	B1-8	B1-8	B2-8	B1-8	B2-8	B2-8	B2-8	B1-8
									<b>.</b>		<i>(</i>
SiO2	(49.8)	(51.2)	(50.1)	(49.9)	(51.1)	(49.6)	(48.3)	(49.5)	(49.5)	(49.7)	(4 <b>8</b> ·4)
Al203	(13.8)	(141)	(13.9)	(14:0)	(13·7)	(12.6)	(14.5)	(14-2)	(13.9)	(13.3)	(14.0)
FeOT*	14.88	13.08	15.04	15.03	10.75	10.55	10.62	10.82	10.76	`10. <b>8</b> 5	10.85
MgO	(9.2)	(6·7)	(8.4)	(9.0)	(7.4)	(5:0)	(8.0)	(6.4)	(6·5)	(6.3)	(7.0)
CaO	11.25	11.63	11.35	11.32	14.25	13.63	12.47	12.91	13.24	13.44	13.21
Na2O	2.52	2.48	2.62	2.57	1.77	2.42	2.05	2.48	2.28	2.08	2.22
K20	0.28	0.17	0.25	0.24	0.10	0.20	0.20	0.24	0.22	0.21	0.16
TiO2	2.11	2.22	2.14	2.13	1.20	1.57	1.39	1.61	1.59	1.61	1.63
MnO	0.23	0.24	0.22	0.23	0.17	0.15	0.14	0.17	0.17	0.17	0.16
P205	0.21	0.23	0.24	0.22	0.10	0.20	0.18	0.22	0.20	0.17	0.18
Totai	(104.2) (	(102.1) 1	(1042)	(104.6)	(100.5)	(95.8)	(97.9)	(98.6)	(98.4)	(97.9)	(97.8)
LOI											
Mg#	( <b>56</b> )	(52)	(54)	<b>(56</b> )	( <b>59</b> )	(50)	(61)	(55)	(56)	(55)	(58)
Nb	19	18	18	19	10	14	13	15	14	13·	13
Zr	130	131	131	128	64	94	89	104	93	94	89
Y	33	41	33	31	24	23	19	23	22	20	20
Sr	219	178	221	218	181	257	250	250	257	268	263
Zn	107	123	106	111	77	73	70	76	73	75	76
Ni	68	66	<b>68</b>	75	77	67	75	63	65	73	71
V	399	451	413	412	282	280	281	302	287	283	277
Cr	92	124	94	87	425	182	155	156	172	169	170
Ga	20	18	18	18	15	16	1/	19	19	18	19
5C	40	48	40 67	45	5U 47	40	37	30	41 45	40	39
	00	30	07	04	4/	40	40	****	40	49	4/
De la											
Le Le			13						10		
			28						21		
Dr .			20						21		
Nd			19						14		
Sm									••		
Eu											
Gd											
Dy											
Но											
Er											
Yb											
Lu											
Data source	8	~	<b>C</b>	<b>c</b>	-	~	-	-	~	-	<b>P</b>
Majors Dec	۲	۲	۲ D	۲	٢	٢	۲	۲	۲ 0	۲	۲
REE			U						U		
Pet. type Alteration	1a	1a	1a	1a	2c	2c	<b>2</b> b	<b>2</b> b	2b	<b>2</b> b	<b>2</b> b

Sample #	G45	A417	A418	A396	A395	A393	A124	A397	A150	A411	G204
Unit	TB 1	TB 3	TB 3	TB 6	TB 6	TB 6	TB 6	TB 6	TB 6	TB 6	TB 6
Map ref	B1-8	B2-8	B2-8	B2-8	B2-8	B2-8	B2-8	B2-8	B2-8	B2-8	B2-8
SiO2	(50.3)	(50.1)	(48.8)	(4B·3)	(47.8)	(49.1)	(52.1)	(50·I)	(50.5)	(49.7)	(57.3)
AI2O3	(12.5)	(13.0)	(12.9)	(150)	(15.6)	(14.5)	(13.1)	(14.5)	(IS-4)	((S·2)	(13.1)
FeOT*	11.38	14.01	14.27	9.72	9.84	10.09	8.94	10.17	9.80	10.75	9.53
MgO	(S·2)	(8-8)	(76)	(8.4)	(9.5)	(8.0)	(4:7)	(7.4)	(5.8)	(7:8)	(4.5)
CaO	13.76	10.74	10.34	13.40	13.03	13.94	14.31	14.17	14.27	13.38	13.71
Na2O	2.38	1.81	2.01	1.79	1.63	1.82	1.89	1.72	1.85	1.68	1.93
K20	0.21	0.29	0.38	0.11	0.11	0.10	0.10	0.12	0.04	0.10	0.15
1102	1./1	2.03	2.12	0.93	0.93	1.04	1.17	1.01	1.34	1.17	1.29
MnU	0.17	0.22	0.22	0.14	0.15	0.10	0.15	0.10	0.14	0.17	0.10
P200	(17.6)	(120)	(aa, a)	0.11 Gir a	(10	0.00	0.13	60.08	0.11 (cr a)	0.12 (\com_mail)	0.10
LOI	(1+-4)	(101.1)	(98.8)	(47.9)	(98.7)	(48-8)	(96.6)	(99.5)	(47.3)		(45.8)
Mg#	(49)	(57)	(53)	(64)	(67)	(62)	(52)	(60)	(55)	(60)	(50)
Nb	14	24	23	8	8	9	8	9	8	8	11
Zr	<b>9</b> 6	145	141	55	55	60	62	57	53	58	73
Y	24	33	30	18	19	18	20	19	20	20	22
Sr	270	243	241	172	168	180	208	178	187	161	191
Zn	80	121	109	57	64	71	69	69	71	72	77
Ni	63	86	64	110	111	117	85	114	75	98	68
V	311	419	383	243	244	253	267	255	309	278	268
Cr	168	193	160	324	333	331	332	354	313	2/6	368
Ga So	17	10	17	10	10	17	18	19	19	18	19
SC Co	44	41 62	୍ୟ ଟ	42	41	41	43	41	40 97	45	40
Rh		UZ	8			40			57	40	
Ba											
La								6			
Се								5			
Pr											
Nd								8			
Sm											
Eu											
Gd											
Dy											
Ho											
Lu											
Data source	1										
Majors REE	Ρ	Р	Ρ	Ρ	Ρ	Р	Ρ	P D	Ρ	P	Ρ
Pet. type Alteration	2b	2b A1+2	2b A1+2	2c	2c	2c	2c	2c	2c	2c	2c

		`						-			
Sample #	A556	A254	A447	A316	A228	A231	A356	A357	A286	A287	A419
Unit	TC 1	TC 2	TC 2	DA 1	DA 1	DA 1	DA 1	DA 1	DA 1	DA 1	DA 2
Map ref	B2-3	B2-3	B2-3	B2-7	B2-6	B2-6	B2-5	B2-5	B2-3	B2-3	B2-6
SiO2	(50.6)	(53.3)	(51.8)	4.8.9)	(48·3)	(47.3)	(47.8)	46.76	(49.3)	(50-4)	(50.5)
AI203	(12.a)	(10.9)	(12.6)	(14.2)	((2.3))	(12.0)	(16.6)	14.49	(15.3)	(13·7)	(14.4)
FeOT*	14.52	13.11	16.02	11.90	12.36	12.20	12.79	12.29	12.10	11.88	12.51
MaO	(7.0)	(2·I)	(1, 5)	(8-5)	(8.7)	(7.6)	$(11 \cdot q)$	8.41	(10.2)	(9.1)	(8.0)
CeO	11 51	12.29	945	12 62	14.86	15.38	11.39	13.03	11.99	14 42	13.64
	200	1 75	1 95	1 80	1.38	1.38	1 38	1 77	1 60	1 55	1 54
	0.12	0.12	0.25	0.00	0.04	0.03	0.06	0.09	0.12	0.07	0.05
	1.06	2 22	2.16	1 42	1.96	1 79	1.57	1.69	1 22	1 29	1 55
1102	0.01	3.33	3.10	1.40	1.00	1.70	1.57	1.50	1.30	1.30	1.55
MnO	0.21	0.23	0.34	0.22	0.10	0.21	0.22	0.20	0.10	0.20	0.27
P206	0.18	0.38	0.31	0.14	0.17	0.18	0.18	0.16	0.17	0.15	0.13
Total .	(101.0)	(991)	(100-5)	(49.8)	(100·1)	(98-0)	(104.0)	98.75	(102.3)	(102.4)	)((03+1)
LOI					5.36			2.67			
Mait	(50)	(37)	(37)	(60)	(60)	(57)	(66)	59.0	(64)	(62)	(59)
	(00)		(0.)	(00)	(00)	(0.7	(00)	00.0		(44)	(00)
Nh	12	26	25	٩	12	11	10	10	10	10	11
7,	00	176	170	70	01	97	70	70	75	70	70
	20	45	17Z A1	22	06 06	07	78	78 20 0	75	70	~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~
T E=	30	40	41	20	20	21	21	22.2	404	21	23
Sr Ta	220	259	259	106	200	196	170	201	101	179	201
21	99	233	135	92	99	87	86	86	80	78	88
Ni	62	76	82	102	125	106	235	173	83	123	98
V	391	586	551	305	326	285	261	252	276	274	346
Cr	168	106	106	305	329	170	400	308	212	240	224
Ga	15	22	22	18	22	19	16	16	19	16	17
Sc	44	49	45	39	43	41	34	36	36	42	44
Со	63	60	77	51	49	52	63	58	54	52	54
Rb								0.7			
Ba								32			
La								6.8			
Се								16.9			
Pr								2.3			
Nd								12.1			
Sm								32			
Fu								12			
Gd								1.2			
								4.0			
Uy He								4.4			
<b>no</b>								0.8			
Er			•					2.4			
YD								2.2			
Lu								0.3			
Date ecured											
Jaia BUUIG		D	Ð	D	Б	Б	D	E	Б	Б	Б
majura	Г	Г	г	r	r	r	٣	г г	٣	r	٣
REE								E			
Det turne	<b>2</b> h	1h	16	39	39	30	39	30	30	30	20
Attornation	20	A1 + 2	10 A1 ( 0	Ja	<b>M</b> 2		<u>مم</u>	0a 80	0a 82	<u>مم</u>	20 A1
MILEIGUON		AI+Z	MI+2		~~	~~	~~	~~	~~	nc.	~1

Sample #	A420	A448	<b>A42</b> 1	<b>A46</b> 7	A468	A450	<b>A45</b> 1	A272	A433	A274	A296
Unit	DA 2	DA 2	DA 2	DA 3	DA 3	DA 3	DA 3	DB 1	DB 1	DB 2	DB 2
Map ref	B2-6	B2-6	<b>B2-5</b>	B2-6	B2-6	B2-6	B2-6	<b>B2-2</b>	B2-2	B2-1	B2-2
SiO2	(18.5)	(50.1)	(46.5)	(48.7)	(51.9)	(51.8)	(536)	(45.3)	(42.6)	(48.7)	(50·2)
A1203	(12.4)	(12.1)	(12.7)	(13.8)	(13.6)	(13.2)	(z.a)	(111)	(120)	(13.3)	(11.5)
FeOT*	11.09	10.97	11.24	17.36	12.35	13.34	11.00	17.17	20.71	12.99	13.45
MgO	(10.3)	(7.7)	(1.1)	(9.5)	(5.1)	(5.7)	(3.1)	(12.1)	(19.8)	(9.4)	(11.6)
CaO	14.16	14.27	13.66	10.11	10.99	11.00	10.06	10.19	10.14	13.10	13.74
<b>Na2O</b>	1.28	1.42	1.03	1.52	2.03	2.10	1.78	0.80	0.61	1.44	0. <del>9</del> 4
K20	0.19	0.02	0.04	0.21	0.18	0.46	0.20	0.21	0.03	0.04	0.05
TiO2	1.44	1.50	1.69	2.13	2.36	2.09	2.40	2.61	2.54	1.73	1.92
MnO	0.20	0.21	0.15	0.33	0.22	0.22	0.20	0.23	0.14	0.20	0.21
P205	0.19	0.16	0.18	0.27	0.34	0.23	0.30	0.25	0.22	0.17	0.19
Total	(100.7)	(99.7)	(A9.3)	(103.9)	(99.0)	(100.1)	(96.4)	(103-0)	(110.7	(101.7)	(103.8)
LOI	• • • •		C 9			- /	(~ //				
Mg#	(66)	(59)	(67)	(53)	(46)	(47)	(37)	(60)	(67)	(60)	(64)
Nb	14	11	15	22	24	23	25	21	21	12	14
Zr	86	77	94	141	145	144	145	137	138	<b>8</b> 6	<b>9</b> 3
Y	22	24	21	34	36	32	38	35	30	25	27
Sr	252	172	244	242	265	253	284	185	161	190	208
Zn	72	90	103	122	101	120	123	140	221	86	93
Ni	81	110	82	73	64	67	72	96	70	91	82
V	274	325	355	433	436	391	437	521	508	338	389
Cr	213	320	310	177	198	177	202	156	179	186	138
Ga	22	19	22	18	20	18	21	20	20	19	16
Sc	34	44	40	46	45	38	42	54	52	40	48
Co	45	46	44	74	53	60	50	77	92	56	59
KD											
Nd											
Sm											
Eu											
Gd											
Dv											
Ho										<i>,</i>	
Er											
Yb											
Lu											
Data source	e										
Majora	P	Р	Р	P	Р	P	Ρ	р	P	Р	P
REE	-	•	•	•	•	•	•	1	•	•	•
Pet. type	2a	3a	<b>2</b> a	<b>2a</b>	2c	2c	2c	1b	1b	<b>2a</b>	<b>2a</b>
Alteration				A1	A1		<b>A</b> 1	A1+2	A1+2	A1	A1
				*							
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Sample #	A519	A446	A521	A281	A282	A518	A429	A277	A390	A391	A132
Unit	DB 2	DB 2	DB 3	DB 4	DB 4	DB 4	DB 5	DB 6	DB 6	DB 6	DB 7a
Map ref	B2-2	B2-2	B2-3	B2-3	B2-3	B2-3	B2-2	B2-2	B2-2	B2-2	B2-2
SiO2	(44.0)	(53.2)	(SI · I)	49.03	(48·9)	(51.0)	(50.3)	(50.0)	) (48.8)	(48.8)	(49.8)
AI2O3	(12.8)	(4.1)	(12.7)	14.39	(13.6);	(130)	(13·7)	(12.7)	(12.7)	(13.2)	( 15 · I)
FeOT*	15.87	9.44	<b>ì</b> 4.7Ó	12.75	13.06	11.15	12.52	11.74	12.27	11.64	13.19
MgO	(15.0)	(2·5)	(5.5)	6.15	(8.5)	(4·3)	(7.7)	(C · 8)	(7.6)	(7.3)	(87)
CaO	13.72	11.02	11.02	12.89	11.25	13.05	12.29	13.96	14.05	13.49	, <b>12.0</b> 7
Na2O	0.91	2.04	2.45	2.15	1.85	2.38	1.78	1.90	1.95	2.00	2.33
K20	0.01	0.12	0.38	0.29	0.23	0.27	0.13	0.16	0.22	0.22	0.20
TiO2	1.7 <del>9</del>	2.14	2.39	1. <b>6</b> 3	1.53	2.13	1.61	1.60	1.56	1.55	1. <b>6</b> 5
MnO	0.26	0.15	0.24	0.20	0.18	0.20	0.18	0.19	0.20	0.18	0.20
P205	0.15	0.15	0.23	0.17	0.16	0.24	0.14	0.14	0.12	0.13	0.16
Total	(104.5)	(94:9)	(( <i>∞</i> .7)	99.63	(99.3)	(97.7)	(1∞-3)	(99.2)	(49.5)	(98.5)	(103 .4)
LOI				4.65							
Mg#	(66)	(35)	(44)	50.3	(58)	(45)	(56)	(55)	(57)	(57)	( <b>58</b> )
Nh	16	17	20	11	12	15	12	12	11	11	13
7r	111	108	127	91	84	102	85	82	83	81	84
V	30	33	34	25	23	32	25	23	24	23	26
' Sr	179	229	244	204	205	229	108	216	213	214	176
7n	109	87	113	04	89	107	93	87	85	82	92
Ni	79	60	62	104	117	84	82	65	64	60	91
V	448	431	425	327	304	402	363	330	315	302	331
Cr	202	270	112	194	234	262	238	165	159	147	215
Ga	19	22	20	19	18	19	17	18	19	20	18
Sc	48	<u> </u>	39	38	35	44	47	41	40	40	46
Co	69	38	65	57	60	47	53	45	50	47	59
Rb				6.1							
Ba				77							
La				6							6
Ce				18							22
Pr											
Nd											12
Sm											
Eu											
Gd											
Dy											
Но											
Er											
Yb											
Lu											
Data sourc	e										
Majors	Р	Р	Р	F	Р	Р	Ρ	Ρ	Ρ	Ρ	Р
REE	-	-		D		•					D
				-							-
Pet. type	<b>2a</b>	2a	1b	<b>2a</b>	<b>2a</b>	<b>2a</b>	<b>2</b> b				
Alteration	A1	A1+2									

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Sample #	A133	A142	A322	A323	A325	A477	A476	A475	A474	A473	A472
Unit	DB 7a	DB 7b	DB 7b	DB 7b	DB 7b	DB 7b	DB 7b				
Map ref	B2-2	B2-2	B2-2	B2-2	B2-2	B1-4	B1-4	B1-4	B1-4	B1-4	B1-4
<b>-</b>						-					
SiO2	48.67	48.67	(50.0)	(49.1)	(49.5)	(19.0)	(48.9)	(48.7)	46.06	(19.2)	1(Q.7)
A1203	14 84	14 87	(15.2)	(1,1)	(12-4)	· (12.1)	(5.4)	(IST)	13.92	(16.2)	(i, i)
FeOT*	12.08	12 11	13 18	12 07	13 16	13 60	13.08	12.90	13 26	13.39	14 02
MaQ	6.82	6.89	(a.G)	(2.0)	(q.)	(72)	(7.9)	$(q_{1})$	9.87	(9.5)	(2.8)
CaO	12 74	12 65	12 25	12 93	13.32	11.94	11.96	11 72	11 27	11.79	11.54
Na2O	216	2 17	2.00	2 17	211	2 21	2 44	2 27	2 04	2.33	2 29
K2O	0.18	0.21	0.18	017	0.13	0.09	0.12	0 10	0.16	0 14	0.20
TiO	1 51	1.51	1.62	1 40	1 57	217	1 99	2.07	1 77	1 79	2.37
MnO	0.19	0.19	0.21	0.21	0.23	0.19	0.21	0.19	0.20	0.21	0.20
P205	0.16	0.15	0.14	0.13	0.20	0.78	0.21	0.10	0.18	0.18	0.23
Total	99.34	QQ 41	(101.7)	(100)	(101)	$\sqrt{(m)}$	(102.2)	(101.a)	0.10	(104.8)	(101.)
	0.15	1 05	(ioq.)		( २२ क	) (100.1)			-0.22	0040)	
	0.15	1.05							-0.22		
Mg#	54.2	54.4	(61)	(57)	(58)	(53)	(56)	57)	61.0	(60)	(54)
								,			
Nb	11	11	13	9	11	16	14	16	13	14	21
Zr	81	79	84	75	83	107	105	108	91	<b>8</b> 5	124
Y	26	23.8	26	24	28	32	33	32	25.0	27	32
Sr	189	179	174	187	188	211	194	191	183	184	222
Zn	87	87	93	<b>8</b> 5	93	98	96	102	93	<del>9</del> 7	107
Ni	90	84	89	76	90	138	182	258	209	222	136
V	336	319	350	302	338	360	361	378	352	346	416
Cr	232	221	234	233	228	<b>29</b> 7	455	664	557	701	420
Ga	15	18	19	17	17	20	18	21	16	16	17
Sc	46	46	46	43	42	41	43	40	38	43	43
Со	<b>6</b> 5	60	60	54	58	60	61	61	66	66	63
Rb	1.6	6.3							3.2		
Ba	69	59							51		
ها	7	7.6				11	8	7	8.7	6	12
Ce	21	18.0				26	25	24	21.0	21	38
Pr		2.5							2.8		
Nd		12.5				19	16	15	14.2	14	24
Sm		3.3							3.6		
Eu		1.3							1.4		
Gd		4.1							4.5		
Dy		4.7							5.0		
Но		0.9							0.9		
Er		2.5							2.6		
Yb		2.4							2.5		
Lu		0.4							0.4		
Date											
Data source	•	-	~	~	~	-	_	_	_	_	_
Majors DEC	F	F	۲	Р	۲	4	4	P	F	P	P
REE	U	E				D	D	D	E	D	D
Pet. type Alteration	<b>2</b> b	3a	3a	Зb	Зb	3b	3b				

			¥		¥						
Sample #	<b>A47</b> 1	A470	A480	A469	A479	A327	A283	A284	A285	A289	A290
Unit	DB 7b	DB 7b	DB 7b	DB 7b	DB 7b	DB 8a	DB 8a	DB 8a	DB 8a	DB 8a	DB 8a
Map ref	B1-4	B1-4	B1-4	B1-4	B1-4	B2-1	B2-3	B2-3	B2-3	B2-3	B2-3
,	()	· C		· (		(		(1	(100)	(, , )	1
SiO2	(48.6)	(48 9)	46.56	(480)	46.93	(49.7)	(48.4)	(49.6)	(49.3)	.(49.6)	(49.6)
AI203	(15.1)	(14.9)	12.97	(14.4)	13.65	(13.6)	(15-1)	(14.5)	(14:7)	(14.0)	(14.7)
FeOT*	14.33	14.68	14.50	14.65	14.19	12.89	12.77	12.88	12.69	12.97	12.92
MgO	(8.8)	(8.4)	7.25	(9.2)	5.54	(9.3)	(11.4)	(9.6)	(q.4)	(9.2)	(9.4)
CaO	10.96	11.30	10.84	11.05	11.11	13.36	11.36	11.97	11.65	12.44	11.89
Na2O	2.42	2.43	2.44	2.21	2.41	1.83	1.63	1.92	2.02	1.98	1.96
K20	0.31	0.29	0.39	0.29	0.26	0.08	0.10	0.08	0.13	0.09	0.13
TIO2	2.26	2.35	2.46	2.23	2.75	1.61	1.42	1.66	1.51	1.67	1.61
MnO	0.22	0.23	0.24	0.22	0.22	0.22	0.19	0.18	0.19	0.21	0.20
P205	0.23	0.23	0.29	0.18	0.29	0.14	0.16	0.16	0.17	0.16	0.16
Total	$(103 \cdot 2)$	(loz·5)	97.92	(102.5)	97.33	LO27)	(026)	(102.5)	(101.8)	)(102.4)	(102-6)
LOI			0.09		1.75						
Mg#	<b>(56</b> )	(55)	51.2	(57)	45.0	(60)	( <b>65</b> )	(61)	(61)	(60)	(60)
Nb	19	21	22	20	24	12	10	13	11	12	11
Zr	119	125	142	127	152	83	77	85	82	84	82
Y	32	33	34.1	33	40.1	23	23	28	27	26	28
Sr	208	216	212	211	230	192	179	195	196	195	178
Zn	107	108	119	109	122	88	91	92	91	92	94
Ni	196	142	116	151	90	95	95	97	87	90	111
V	395	405	431	400	489	327	297	323	291	332	317
Cr	484	408	306	407	250	212	183	195	176	196	197
Ga	17	20	16	16	16	18	17	19	15	15	16
Sc	40	39	40	40	45	42	38	44	39	40	42
Со	69	69	68	67	<b>6</b> 5	55	59	60	59	58	58
Rb			9.0		5.5						
Ba			112		103						
La	11	12	15.4	14	16.6						
Се	30	25	35.7	34	37.8						
Pr			4.7		5.2						
Nd	17	17	22.2	19	24.6						
Sm			5.5		6.0						
Eu			1.9		2.1						
Gd			6.3		7.1						
Dy			6.7		7.6						
Но			1.3		1.4						
Er			3.6		4.1						
Yb			3.4		3.8						
Lu			0.5		0.6						
Data source	•										
<b>Majors</b>	Ρ	Р	F	Р	F	Р	Р	Р	Р	Р	Р
REE	D	D	Ε	D	Е						
Pet. type Alteration	Зb	3b	3b	3b	3b	2b	2b	<b>2</b> b	<b>2</b> b	2b	2b

		¥									
Sample #	A288	A478	A139	A292	A297	A348	A336	A338	A279	A329	
Unit	DB 8a	DB 8b	DB 9	DB 9	DB 9	DB 10a	DB 10a	DB 10a	DB 10b	DB 10b	
Map ref	B2-3	B1-4	B2-2	B2-3	B2-2	B2-5	B2-1	<b>B2-5</b>	<b>B</b> 2-1	B2-1	
•							~		6		
SiO2	(50.9)	46.74	(49.8)	(49.3)	(490)	(48-6)	(46.6)	(4 <b>8</b> ·2)	(49.6)	(48:4)	
A1203	(13.0)	11.32	(14.5)	(145)	(14-4)	(15.0)	(13·7)	(14.7)	(14.4)	(13 · 3)	
FeOT*	12.64	18.79	11.95	12.14	12.60	12.11	12.11	12.36	12.95	14.00	
MgO	(6.8)	4.67	(7.3)	(7.4)	(8.2)	(9.9)	(9 <i>·8</i> )	(10.2)	(9.5)	(8.8)	
CaO	13.54	9.81	12.81	12.69	12.42	11.92	12.40	11.70	12.36	11.47	
Na20	2.13	2.49	2.24	2.20	2.01	1.69	1.73	1.68	2.13	2.14	
K20	0.16	0.46	0.21	0.21	0.08	0.05	0.07	0.08	0.19	0.21	
TiO2	1.76	3.46	1.49	1.65	1.66	1.49	1.47	1.48	1.56	1.73	
MnO	0.22	0.27	0.20	0.19	0.18	0.16	0.21	0.18	0.20	0.23	
P2O5	0.15	0.37	0.17	0.17	0.16	0.15	0.14	0.14	0.14	0.18	-7
Total	(101.2)	98.35	(100-6)	) ((∞·4)	(100.7	)(101.0)	(98.3)	(100.6)	(03.1)	( (00 2	>)
LOI		-0.15									
Mo#	(53)	34.3	(56)	(56)	(58)	(63)	(63)	(63)	( <b>61</b> )	(57)	
	- /		```		- /		-	-			
Nb	14	30	11	12	12	12	11	10	12	12	
Zr	90	197	79	86	<b>8</b> 5	79	77	78	82	94	
Y	27	49.7	23	27	26	22	22	22	23	28	
Sr	199	205	209	212	209	186	182	185	182	193	
Zn	96	162	78	87	83	85	78	83	87	93	
Ni	72	42	72	62	69	94	83	97	83	79	
V	362	536	303	343	324	302	296	298	342	354	
Cr	214	48	166	157	165	198	198	193	227	173	
Ga	17	20	22	19	18	17	20	17	17	16	
Sc	45	41	42	44	40	38	39	34	48	40	
Co	53	87	55	53	55	53	52	55	58	60	
Rb		10.2									
Ba		132									
La		19.9									
Ce		46.9									
Pr		6.4									
Nđ		30.3									
Sm		7.6									
Eu		2.6									
Gd		9.0									
Dy		9.8									
Но		1.8									
Er		5.3									
Yb		4.9									
Lu		0.7									
Data sour	ce										
Majors	Р	F	Р	P	P	Р	Р	P	Ρ	Р	
REE		Ε									
Det hrist	0h	10	9h	Эh	2h	20	20	29	29	29	
Alteration	20	ia	٤U	עש	20	20	20	20	24		

Sample #	A330	A334	A335	A341	A344	A343	A342	A349	A415	A414
Unit	DB 10b	DB 10b	DB 10b	DB 10b	DB 10c	DB 10c	DB 10c	DB 10c	S 1	S 1
Map ref	B2-1	B2-1	B2-1	B2-1	<b>B2</b> -5	<b>B2</b> -5	<b>B2-1</b>	<b>B</b> 2-5	B2-8	B2-8
SiO2	(49.1)	(48.6)	(48.6)	(49-9)	.(47.8)	(48·9)	(4.8-4)	(47.8)	(50.5)	(51.6)
AI2O3	(14:0)	(13.6)	(13.6)	(14-3)	(14.0)	(14-6)	(4.7)	(15.2)	(14.0)	(13.9)
FeOT*	13.34	12.15	13.70	12.52	11.67	12.25	12.49	12.22	9.88	10.55
MgO	(10-0)	(q.5)	(10·2)	(9.2)	. (8.6)	(9.0)	(9.1).	((0.7)	(+.+)	(7.+)
CaO	12.55	12.69	12.19	12.57	13.03	12.37	12.06	11.48	14.51	13.00
<b>Na2O</b>	2.02	1. <b>8</b> 5	1.91	1.75	1.92	2.06	2.19	1.78	1.00	1.40
K20	0.16	0.13	0.13	0.12	0.15	0.20	0.21	0.18	0.02	0.03
TiO2	1.52	1.51	1.7 <del>9</del>	1.54	1.29	1.4/	1.54	1.33	0.07	0.09
MnO	0.21	0.19	0.20	0.18	0.20	0.19	0.19	0.10	0.10	0.17
P2O5	0.14	0.12	0.16	0.15	0.14	0.18	0.17	0.15	(aa)	(0.05)
Total LOI	(103.0)	(100.3)	(102.4)	(102.2)	. (98-8)	( [0(·Z)	(101-0)	(01.0)	(49.4)	(14.5)
Mg#	(61)	(62)	(61)	(61)	(61)	(61)	(60)	(65)	(62)	(61)
Nb	10	12	14	12	11	11	11	11	3	4
Zr	76	80	97	82	72	81	81	75	38	45
Y	23	22	28	22	20	23	24	22	18	19
Sr	181	184	187	183	191	184	181	1/5	144	133
Zn	86	90	97	85	82	81	87	82	440	102
NI	<b>9</b> 3	81	79	90	93	79	00	200	119	281
V	336	331	359	338	257	308	313	290	406	A12
Cr	225	225	171	221	1/1	19	104	10	15	17
Ga	18	18	19	19	34	37	38	35	42	47
SC	43	44 55	40 61	42 57	52	55	53	54	41	47
	30	55	01	57	Ű.	•••				
Ro Ro										
i e I e		8				5				
Ce :		19				17				
Pr										
Nd		13				11				
Sm										
Eu										
Gd										
Dy										
Но										
Er										
Yb										
Lu										
Data sou	rce	-	п	D	D	P	P	P	Р	Р
Majors	۲	r n	Г	F	r	, D	•	•	-	
KEE		U			_		-	<b>A</b> -	0-	0-
Pet. type Alteratio	2a n	2a	2a	2a	2a	2a	28	28	28	28

	¥										
Sample #	<sup>5</sup> A416	A409	A410	A151	A406	A345	<b>A35</b> 1	A352	A355	A233	A465
Unit	S 1	S 1	S 1	S 1	S 1	S 1	S 1	S 1	S 1	S 1	S 2
Map ref	B2-8	B2-8	B2-8	B2-8	B2-8	<b>B2-</b> 5	B2-5	B2-5	B2-5	B2-6	B2-6
•											
SiO2	49.79	(50.1)	(4.9.7)	(507)	(51.8)	(510)	(50.1)	(49.1)	(50.4)	(48.4)	(56.5)
AI203	16.63	(14.6)	(16.1)	(5.5)	(13.5)	(14.0).	(13.8)	(15.0)	(14.1)	(11.3)	(12.6)
FeOT*	9.74	10.51	10.54	10.59	10.32	9.34	Ì0.49	10.31	9.59	10.13	13.84
MgO	6.85	(8.4)	(9.3)	(8.2) .	(7.3)	(5.8)	(8-6)	(9.1)	(7-1)	(10·5)	(5.7)
CaO	14.57	14.41	13.11	13.71	14.65	14.10	14.67	13.94	14.12	13.68	11.58
Na20	1.72	1.60	1.34	1.40	1.51	1. <b>6</b> 6	1.55	1.53	1.67	1.31	2.32
K20	0.04	0.01	0.03	0.05	0.06	0.04	0.03	0.04	0.04	0.01	0.26
TiO2	0.78	0.88	0.78	0.80	0.86	0.84	0.79	0.79	0.79	0.75	2.78
MnO	0.15	0.16	0.16	0.18	0.18	0.13	0.17	0.16	0.13	0.16	0.30
P2O5	0.06	0.05	0.05	0.06	0.07	0.06	0.05	0.06	0.05	0.05	0.74
Total	100.34	(100.9)	(101.1)	(101.2)	(100.2)	(97.0)	(100.3)	(100-0)	(97.9)	(99.3)	(100.7)
LOI	0.78						< J	(()		- (( ))	
			(c_)	(-)	(			(r.)		()	/\
Mg#	59.6	( <b>63</b> )	(65)	(62)	(60)	(57)	( <b>63</b> )	(65)	(61)	(68)	(47)
	_			•			•			•	•
ND	4	5	4	3	4	4	3	4	4	3	24
	36	34	34	36	39	38	36	35	34	34	184
Y	17.3	18	17	19	18	17	15	16	18	16	48
Sr	141	121	106	113	130	143	140	131	137	114	362
Zn	67	63	65	67	71	66	65	60	63	65	144
NI	105	119	124	131	120	100	127	127	116	121	64
V	269	261	253	263	279	287	251	255	257	252	412
Cr	394	433	410	425	445	439	374	364	390	342	124
Ga	16	17	18	17	16	19	15	18	18	17	19
Sc	48	49	45	45	49	45	43	41	43	41	39
C0	43	46	49	50	45	36	45	45	39	43	59
Rb	0.6										
Ba	18										
	2.5						1				
Ce	6.2						5				
Pr	0.9										
Nd	5.4						4				
Sm	1.7										
Eu	0.7										
Gd	2.4										
Dy	3.2										
HO	0.6										
Er	1.8										
YD	1.8										
Lu	0.3										
	•										
Maiore	F	Р	P	P	P	Р	P	P	P	P	D
REF	F	I	T.		ſ	F	n	ſ	F	F	r
ططا :	-						U				
Pet. type Alteration	2a	2a	<b>2a</b>	2a	<b>2a</b>	<b>2a</b>	2a	2a	<b>2a</b>	2a	2b

<sup>5</sup>Sample A416 has been used as an internal monitor

								¥			
Sample #	<b>4466</b>	A496	A300	A481	A307	A305	A306	A301	A303	A482	A311
Unit	S2	S 2	S 3	S 3	S 4	S 4	S 4	S 4	S 4	S 4	S 5
Map ref	B2-6	B2-6	B2-8	B2-8	B2-8	B2-8	B2-8	B2-8	B2-8	B2-8	B2-8
									<b>.</b> .		4
SiO2	(50.3)	(50.5)	(50.1)	(uq.b)	(49.7)	(So.7)	(49.5)	48.93	(51.3)	(48.6)	(4.4.2)
AI203	(12.5)	(12.0)	(13.2)	(13.2)	(12.6)	(12.2)	(13.3)	13,67	(12.1)	(13.7)	(3·7)
FeOT*	12.84	14.23	13.60	15.84	14.10	14.18	13.99	13.28	14.23	13.71	11.58
MaO	(4.0)	(6.9)	(6.9)	(8.7)	(8.5)	ほみ	(9.4)	6.92	(6-9)	(0.1)	(6.6)
CaO	11.66	11.35	11.64	10.81	11.72	<b>11.6</b> 5	11.75	12.01	11.80	11.76	11.08
Na2O	2.61	1.99	2.42	2.03	2.10	2.09	2.21	2.32	2.03	2.33	1.98
K20	0.35	0.49	0.22	0.27	0.11	0.11	0.19	0.28	0.18	0.27	0.38
TiO2	2.92	2.90	2.13	2.03	1.89	1.87	1.79	1.75	1.78	1.78	1.88
MnO	0.22	0.23	0.22	0.27	0.20	0.16	0.20	0.20	0.18	0.21	0.16
P2O5	0.52	0.42	0.24	0.25	0.23	0.22	0.18	0.19	0.17	0.16	0.17
Total	(97.9)	(102.0)	(00.9)	) (102.9)	(101.1)	(100.9	)(102.5)	99.52	(1006	)(102.6,	) ( <b>9</b> 4· <b>7</b> )
LOI		- •						1.07			
Mg#	(40)	(50)	(51)	(54)	(56)	(53)	(59)	52.2	(50)	(61)	(54)
	06	05	10	18	14	14	13	14	13	13	21
ND 7-	170	191	132	130	105	103	101	98	104	103	117
	20	101	30	27	26	25	26	24	26	25	21
Y 6-	363	315	261	274	257	240	219	204	233	217	315
Sr Za	303	100	111	113	94	98	95	84	98	<b>9</b> 5	83
	27	100	68	86	76	54	74	69	62	72	114
NH M	303	370	390	388	361	363	360	339	362	349	283
v C-	115	96	88	85	88	91	94	91	101	93	195
Ga	22	18	20	20	16	19	16	17	17	15	21
Gia Sc	36	32	41	43	42	45	45	41	44	39	33
эс Со	52	57	60	72	62	60	62	59	61	60	48
Rh	¥.	0,						4.3			
Re								79			
La .					11			11			
Ce					26			17			
Pr											
Nd					15						
Sm											
Eu											
Gd											
Dv											
Ho											
Er											
Yb											
Lu											
Data sour	'Ce					-	-	-	~	~	P
<b>Majors</b>	Р	Р	Ρ	Р	P	P	Р	F	۲	۲	٣
REE					D			U			
Pet. type	2b	<b>2</b> b	1a	1a	1a	1a	18	1a	1a	1a	2b

				_ ₩			*				
Sample #	A315	A105	A159	°A413	A497	A498	<b>A48</b> 6	<b>A48</b> 7	A488	A489	A490
Unit	S 5	S 5	S 5	S 5	S 6	S 6	S 7	S 7	S 7	S 8	S 8
Map ref	B2-7	B2-7	<b>B2</b> -7	B2-8	B2-6	B2-6	B2-7	B2-7	<b>B2-7</b>	B2-7	B2-7
SiO2	(4.8.8)	(49.9)	(49.5)	49.82	(45.2)	(40.2)	48.64	(50.8)	(51.3)	(49.1)	(49.2)
Al2O3	(13.4)	(14:8)	(13.4)	15.65	(S·7).	(3.6)	16.11	(12.9)	(13.4)	(13.1)	(13.0)
FeOT*	12.52	12.46	14.21	11.80	13.88	12.27	8.71	12.49	10.13	13.26	13.27
MgO	(6-9)	(6.7)	(8-0)	4.07	(8.6)	(q·2)	4.84	(5.4)	(4·6)	(6 2)	(5.9)
CaO	11.02	11.82	10.60	12.15	20.39	27.14	12.51	12.47	12.27	11.06	11.33
Na2O	2.17	2.28	2.00	2.33	1.14	0.94	2.39	1.85	2.19	2.36	2.37
K20	0.49	0.44	0.42	0.44	0.11	0.13	0.33	0.22	0.36	0.41	0.42
TiO2	2.40	2.14	2.15	2.09	2.07	1.82	2.29	2.10	2.27	2.61	2.63
MnO	0.18	0.20	0.21	0.16	0.34	0.53	0.14	0.18	0.18	0.19	0.19
P2O5	0.24	0.23	0.24	0.24	1.35	1.00	0.27	0.22	0.21	0,36	0.31
Total LOI	(98.1)	(101.0)	(100.7)	98.76 1.89	(98.8)	(૧૯・૧)	96.22 2.12	(૧૪.૧)	(97.0)	(98.7)	(986)
Mg#	(53)	(53)	(54)	42.0	(57)	(61)	53.8	(49)	(49)	(50)	(48)
Nb	24	21	23	23	10	17	25	21	23	25	25
7r	130	124	142	133	197	110	140	128	141	163	150
V	28	20	31	26.2	30	36	20	24	28	28	20
' Sr	318	23	228	320.2	273	287	356	346	340	321	23
7n	89	05	112	04	76	70	87	96	102	101	102
Ni	91	76	60	102	58	70	58	91	77	72	73
V	335	356	388	317	377	333	355	346	346	353	377
Cr	176	200	149	225	67	62	259	249	239	178	204
Ga	21	18	19	19	22	18	21	18	18	20	20
Sc	32	35	37	36	44	43	39	35	39	34	34
Co	54	53	59	53	51	41	33	51	42	54	53
Rb		•••	•••	9.2	•••	•••	6.5	•••		•	•••
Ba				120			157				
La				16.0			15		16		
Ce				37.0			45		37		
Pr				4.8					•••		
Nd				22.3					22		
Sm				5.2					—		
Eu				1.7							
Gd				5.5							
Dy				5.5							
Ho				1.0							
Er				2.7							
Yb				2.4							
Lu				0.3							
Data source	e										
Majors	Р	Р	P	F	Ρ	Р	F	P	P	Р	P
REE				Е			D		D		
Pet. type Alteration	2b	<b>2</b> b	<b>2</b> b	2b	2a A1+2	2a A1+2	2b	<b>2</b> b	<b>2</b> b	2c	2c

<sup>6</sup>Sample A413 has been used as an internal monitor

Somale #	A 1 6 6	A 200	A 494	A 405	A 402	A954	A 482	A 482	A 402	A 404	
	A 100	<b>A300</b>	<b>8</b> 0	<b>A400</b>	S 10	<b>R-304</b>	<b>A402</b>	<b>A4103</b>	6 10	C 10	S 12
Man ref	30 22.7	09 82.9	59 82.9	82.8	3 IU 82-6	B2.5	3 I I 82-5	82.5	B2-6	B2-6	B2-6
mep rei	D2-1	D2-0	D2-0	D2-0	D2-0	D2-3	DZ-3	Dz-5	D2-0	D2-0	D2-0
SiO2	(1.9.0)	(1.0.)	(1.9.2)	(i.a 2)	(50.9)	150.0	(4.9.0)	(48.8)	(44.8)	(50.4)	46 64
A1203		(43.₩)	(40.7)	(41.3)	(22)	(11.0)		(-1)	621)	(ic.i)	14.96
FeOT*	12 38	978	9.81	13.82	15 74	11 34	10.88	14 75	13.32	11 72	11 75
MaO	(6.9)	(1, c)	(5.9)	(10.0)	(5.8)	(2.1)	(2.2)	(S.S)	(Co)	(2, 7)	5 13
CaO	11.55	13.92	13.09	11.86	10.43	12.66	12.98	11.39	11.61	11.14	11.51
Na2O	2.66	2.13	2.20	2.36	2.47	2.57	2.49	2.49	2.32	1.27	2.84
K20	0.43	0.17	0.18	0.21	0.37	0.30	0.22	0.19	0.40	1.59	0.50
TiO2	2.30	1.30	1.30	1.73	3.18	2.41	2.29	2.83	2.65	1.87	2.64
MnO	0.18	0.15	0.15	0.21	0.26	0.15	0.14	0.19	0.20	0.16	0.20
P2O5	0.31	0.12	0.14	0.17	0.39	0.31	0.31	0.41	0.37	0.24	0.40
Total	(qq.7)	(92.8)	(94.6)	(103.5)	(101.4)	(45. i)	(92.8)	(99.6)	(92.8)	(a.2)	96.56
LOI			\ + <b>-</b>	(		Cong		019	(19)		-0.14
NA-#	(50)	(50)	(50)	(00)	(40)	(00)		(44)	(10)	6	47 0
MG#	(54)	(50)	(30)	(60)	(43)	(39)	(39)	(44)	(48)	(58)	47.8
	~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~	10	44	12	20	04	04	07	20	21	07
7,	142	70	75	00	100	24 1 <i>4 4</i>	122	167	174	107	170
ZI V	142 97	20	10	99 26	188	32	30	107	36	ז בו סס	24
l Sr	311	20	248	20	330	333	348	316	309	508	327
31 7n	<u>9</u> 0	72	66	213 96	134	104	107	135	107	04	108
Ni	78	48	44	71	40	26	28	40	93	161	64
V	347	247	256	362	494	370	353	448	389	315	378
Cr	195	69	68	99	40	82	75	88	125	456	107
Ga	19	20	19	17	19	19	21	18	18	22	19
Sc	36	40	37	47	38	36	38	39	36	29	33
Co	55	39	38	63	67	44	40	61	53	50	49
Rb										5.9	8.7
Ba										175	146
La										16	18
Ce										32	45
Pr											
Nd										19	
Sm											
Eu											
Gd											
Dy											
Но											
Er											
Yb											
Lu											
Deta sourc	е										
Majors	P	Р	P	Р	Ρ	Ρ	Р	Ρ	P	Ρ	F
REE	-									D	D
Pet. type	2c	2c	2c	1a	1a	<b>2</b> b	<b>2</b> b	<b>2</b> b	2c	2c	2c
Alteration										A2	

					¥					
Sample #	<b>A</b> 454	A455	<b>A46</b> 4	A461	A346	<b>A45</b> 7	A458	A350	A459	<b>A460</b>
Unit	S 13	S 13	S 13	S 14	S 15	S 15	S 15	S 16	S 16	S 16
Map ref	B2-6	B2-6	B2-6	B2-5	B2-5	B2-5	B2-5	<b>B2-5</b>	B2-5	<b>B</b> 2-5
SiO2	47.52	(50.5)	(50.9)	(50.7)	47.86	(49.6)	(48.6)	(49.9)	(510)	(49.6)
AI2O3	13.56	(12.2)	(12.3)	(12.4)	19.78	(12.5)	(15.3)	(12.2)	14.1)	(12.2)
FeOT*	14.09	15.36	14.39	14.29	8.42	9.82	9. <del>9</del> 9	10.83	9.59	<b>9.49</b>
MgO	4.12	(4.9)	(5.3)	(70)	3.43	(4.0)	(5.8)	(5.1)	(3.7)	(3.6)
CaO	10.48	10.48	11.14	11.42	13.47	13.32	10.36	14.53	11.89	13.44
Na20	2.59	<b>2.6</b> 5	2.88	2.03	2.46	2.48	1.95	2.09	2.31	2.41
K2O	0.34	0.31	0.30	0.33	0.27	0.26	0.17	0.10	0.32	0.27
TiO2	3.17	3.27	2.95	3.00	1.46	1.77	1.75	2.07	1.94	1.73
MnO	0.20	0.25	0.17	0.25	0.16	0.18	0.18	0.17	0.15	0.17
P2O5	0.47	0.45	0.44	0.43	0.24	0.29	0.32	0.25	0.30	0.24
Total	96.52	(100.3)	(100.7)	(101.8)	<b>97.54</b>	(94.2)	(94.4)	(933)	(95.3)	(93.1)
LOI	3.08	(		( ( · · · )	0.98	~~~~~	<- <del>+</del> +)			
Mg#	38.0	(40)	(43)	(51)	46.1	(46)	(55)	(50)	(45)	(44)
Nb	29	29	<b>2</b> 5	28	15	17	18	17	19	19
Zr	211	206	188	185	115	126	120	111	134	128
Y	45	50	40	37	21.8	25	26	26	32	27
Sr	342	340	349	286	370	360	346	306	386	365
Zn	132	133	109	131	76	76	80	93	92	86
Ni	36	42	43	58	55	45	47	64	49	63
V	430	432	404	472	218	235	267	326	282	224
Cr	36	29	136	78	125	143	165	83	177	143
Ga	22	20	24	22	17	20	20	20	21	15
Sc	40	37	40	39	33	35	33	40	40	35
Co	63	<b>6</b> 5	61	61	33	39	37	45	34	36
Rb	8.6				4.5					
Ba	110				95					
La	20				12.1					
Ce	48				27.9					
Pr					3.8					
Nd					17.4					
Sm					4.1					
Eu					1.5					
Gd					4.4					
Dy					4.5					
Но					0.8					
Er					2.3					
Yb					2.0					
Lu					0.3					
Data source	;									
Majors	F	Р	Р	Р	F	Р	Р	Ρ	Р	Р
REE	D				Е					
Pet. type Alteration	1b	1b	1b	1b	<b>2</b> b	<b>2</b> b	<b>2</b> b	2c	2c	2c

	*			-	7			*			
Sample #	A443	A444	A445	<sup>′</sup> A14 (a)	Á14 (t)	A17	A19	A20	A21	A23	A24
Unit	S 17	S 17	S 17	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS
Map ref	B2-2	B2-2	B2-2	B3-8	B3-8	B3-8	B3-8	B3-8	B3-8	B3-8	B3-8
										20.0	000
SiO2	50.02	(51.5)	(46.3)	(45-9)	(45.7)	(4.3.2)	(45.1)	45.62	(45.2)	(46.1)	(46.8)
A1203	15.81	(13.6)	(12.3)	(15.0)	(15.1)	(15.1)	(40,4)	14.04	(15 1)	(15.1.)	(5.0)
FeOT*	8 84	825	10.13	14.44	14.05	13.01	12.95	13.67	11.63	13.15	1316
MaQ	5 41	().I)	(4.6)	(11.3)	(8.1)	(8.1)	(7.8)	8.84	((,q))	$(7, \epsilon)$	(7.9)
CaO	12 20	10.51	11 50	11 04	11.15	10.74	10.99	10.82	11.12	11 07	11 01
	2 11	280	2 00	214	2 20	2 20	2 35	2 21	240	2 46	2 52
K2O	0.23	0.59	0.24	0.18	0 17	0.17	0.19	0.20	0.19	0.18	0.17
	2 01	2.39	1.54	2 24	2 27	2.08	217	2 21	1.82	2 13	2.02
MnO	0.12	0.17	0.15	0.21	0.21	0.10	0.10	0.20	0.17	0.10	0 10
P2O5	0.12	0.17	0.15	0.21	0.21	0.18	0.10	0.20	0.17	0.17	0.18
Tatal	0.22	(0.40	(a 2)	$(\alpha_2 \beta)$	(9a.1)	$(a \leq 1)$	(a1 1)	0.22	(al =)	(00.17	(20 Y)
	90.97	(42.8)	GD-S)	(102.6)	C(T,T)	(15-4)		0.02	042	(18.4)	(99.7)
LUI	3.34							-0.51	-0.43		
Mg#	56.2	<b>(38</b> )	(49)	(62)	(57)	(57)	(56)	57.6	(55)	(55)	(56)
Nb	21	39	18	16	17	15	14	16	14	15	14
7r	120	275	125	117	119	111	113	110	101	112	103
Y Y	20	53	30	27	30	26	28	26	24	28	26
Sr.	267	334	294	250	252	236	244	246	255	255	255
7n	63	149	91	112	108	94	94	88	83	99	95
Ni	50	25	37	164	167	140	138	133	135	150	182
V	381	212	283	342	330	201	317	204	265	314	202
	210	35	200	372	370	201	351	234	200	342	202
G	218	30	10	10	15	20	10	12	19	15	10
Se	42	24	19	20	20	20	30	27	24	24	22
SC	40 20	25	30	59	69	64	64	SI BA	50	54 66	55
	50	20	30	00	00	04	04	20	39	00	00
	5. <del>8</del> 142							0.U 00			
	142			o	7	0	44	10	10	10	44
	24			0 25	7 77	34	25	20	12	20	21
Ce D-	34			20	21	24	25	32	22	30	21
				16	40	10	20	10	01	20	14
Na Sm				10	10	13	20	10	21	20	14
Sm Fu											
EU											
Ga											
Dy											
Ho											
Er											
Yb											
Lu											
Data sourc	e										
Majors	F	Р	Р	Ρ	Р	Р	Ρ	F	Ρ	Ρ	Ρ
REE	D			D	D	D	D	D	D	D	D
	-			-	·	-	-	-	-	-	
Pet. type Alteration	2c	2c	2c	3a	3a	<b>3</b> a	3a	3a	3a	<b>3a</b>	3a

<sup>&</sup>lt;sup>7</sup>Sample A14 has been powdered in both agate and tungsten carbide tema pots, and analyses from each of these powders are labelled A14(a) and A14(t) respectively.

						¥					
Sample #	A26	A542	A543	A544	A552	A554	A3	<b>A66</b>	A63	A43	<sup>1</sup> A2
Unit	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS LL
Map ref	B3-8	B3-3	B3-3	B3-3	B3-4	B3-8	B3-6	B3-8	B3-4	B3-8	B3-6
SiO2	(44.1)	(17.7)	(46.4)	(47.6)	(47:3)	45.05	(17.9)	(44.5)	(47.4)	(46.3)	(4.3.1)
AI203	(11.8)	(11.5)	(15.1)	$(\mu, \Lambda)$	(15.5)	13.51	$(1, \mathbf{I})$	(11-8)	(15.6)	115.2)	(G.L)
FeOT*	13.44	14.29	14.15	14.22	14.38	14.23	12.66	12.68	13.94	12.60	20.29
MaO	(8.1)	(14.3)	(8.9)	(7.1)	(3.1)	9.48	(63)	(4.0)	(7-5)	12.1	(6.5)
CaO	10.73	11.90	11.15	11.65	11.18	10.44	11.45	11.09	10.87	10.71	10.89
Na2O	2.26	2.34	2.21	2.37	2.38	2.19	2.79	2.38	2.66	2.45	1.99
K20	0.18	0.16	0.16	0.19	0.18	0.20	0.25	0.19	0.22	0.30	0.17
TiO2	2.19	2.19	2.34	2.59	2.32	2.36	2.08	2.25	2.31	2.18	5.40
MnO	0.19	0.23	0.21	0.21	0.21	0.20	0.20	0.19	0.21	0.19	0.27
P205	0.15	0.22	0.16	0.16	0.15	0.22	0.29	0.20	0.20	0.22	0.22
Total	(ar.1)	linio	(100.8)	(100.6)	(101.9)	97 87	(99.6)	(95.l)	(in a	Varu	( ag . ( )
		(0+.0)	-0 28	((00 0)		-0.41	( 10.0 )	(() +)	$(\omega)$	JC77.4,	- <b>0.32</b>
			0.20			-0.41					0.02
Mg#	<b>(56</b> )	<b>(56</b> )	(57)	(51)	(55)	58.3	<i>(</i> 51)	(54)	<b>(53</b> )	(55)	(40)
Nb	16	16	15	16	15	17	15	18	15	17	12
Zr	115	122	117	127	116	123	145	117	122	117	86
Ŷ	29	32	28	32	29	28.4	31	29	30	27	27
Sr	250	203	252	257	250	249	231	229	252	243	133
Zn	99	107	108	111	111	111	88	90	100	92	128
Ni	157	127	178	126	200	188	132	115	190	181	44
V	317	367	345	370	338	335	286	313	312	303	1433
Cr	357	291	409	356	427	441	330	327	419	428	72
Ga	19	17	16	18	19	18	21	17	15	20	16
Sc	30	43	39	40	38	37	36	35	34	33	57
Со	66	64	68	64	70	72	62	61	70	62	87
Rb						2.3					
Ba						53					
لم	10	12	6	10	8	10.4	12	12	9		
Ce	25	28	23	24	17	25.7	37	27	30		
Pr						3.6					
Nd	18	15	12	18	15	18.0	22	17	21		
Sm						4.7					
Eu						1.8					
Gd						5.6					
Dy						5.8					
Ho						1.1					
Er						3.0					
Yb						2.7					
Lu						0.4					
Data source	ce										
Maiors	P	P	Р	Р	Р	F	Р	P	Р	Р	P
REE	ח	D	D	D	ח	F	י ח	, n	י ח	•	•
	-	-	2	5	2	-	~	5	0		
Pet. type Alteration	<b>3a</b>	<b>3a</b>	3a	<b>3a</b>	<b>3a</b>	3a	<b>3a</b>	3a	3a	3a	

<sup>1</sup>ILS LL = Lava lake sample

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Sample #	Δ11	<b>A13</b>	<b>A6</b> 5	A64
Sample #	2 11 5 6	II S d	II S d	ILS cl
Unit Mon rof		R2.7	R3-8	R3-7
Map ret	D3-0	D3-7	0.0	
SiO2	(44.7)	(44.3)	44.79	(46.3)
A1203	(16.4)	(18.1)	13.71	(17.5)
FeOT*	11.84	13.29	13.94	13.69
MaQ	(9.8)	(15.5)	9.87	(11.9)
CaO	10.86	9.29	11.78	10.0Ó
No2O	2 12	2.03	2.24	2.33
KOU	0.18	0.17	0.21	0.19
	1 40	1 1 4	2 29	1 48
	0.47	0.20	0.20	0.20
MNU	0.17	0.20	0.20	0.20
P205	0.20	U.24	0.23	(a z d)
Total	(4+.8)	(04.5)	97.25	105.9
LOI	-0.38		-0.63	
Mg#	<b>(63</b> )	(71)	59.8	( <b>65</b> )
Nb	11	7	17	10
Zr	71	64	122	78
Y	22	20	29	24
Sr	257	175	247	199
Zn	79	85	103	94
Ni	197	306	235	235
V	208	202	340	240
Cr	292	294	508	432
Go	17	16	20	17
Sc	33	29	38	30
Co	66	81	67	79
Dh		0.	4.8	
Re			90	
	10		10	
	21		30	
	21		~	
Nd	15		18	
Sm				
Fu				
Gd				
Dv				
Ho				
ПU Е-				
LU				
Data sour	ce			
<b>Majors</b>	Р	Р	F	Р
REE	D		D	
Dat hart	30	30	30	<u>ao</u>
Alternation	38	56	38	Ja
AIGLATIOU	I			

 $<sup>^{2}</sup>$ ILS cl = samples from the coarse lava flow that caps the Ingolfsfjall table mountain

Appendix 4

	*								¥	*	
Sample #	A512	A511	<b>A</b> 510	A509	A508	A507	A506	A505	-A384	A499	A500
Мар	B1-1	B1-1	B1-1	B1-1	B1-1	B1-1	<b>B1-1</b>	B1-1	B1-1	B1-3	B1-3
Unit	Hæðir	Hæðir	Hæðir	Hæðir	Hæðir	Hæðir	Hæðir	Hæðir	Hæðir	Húsmuli	Húsmuli
		、		<i>(</i> ,			<i>C</i>				
SiO2	<b>47.9</b> 6	(43.9)	(48.9)	(482)	(49,9)	(48.7)	(49.3)	(49.2)	47.40	47.72	(49.7)
A1203	15.26	(15.2)	(15:4)	(60)	((5.5)	(155)	(15.0)	(15.7)	15.53	15.01	(13.2)
Fe2O3	10.39	10.40	10.49	9.82	11.12	11.21	11.58	10.94	10.43	13.15	12.43
MgO	9.33	(7·4)	(7.6)	(7.7)	(7.8)	(11.2)	(8.0)	(8.0)	9.64	5.05	(G · J)
CaO	13.31	13.75	13.69	13.59	13.63	13.49	13.54	13.48	13.16	12.03	12.40
Na2O	1.80	2.24	2.16	2.04	2.01	1.92	2.26	2.13	1.87	2.27	2.18
K20	0.10	0.09	0.09	0.08	0.09	0.08	0.07	0.09	0.10	0.15	0.16
TiO2	1.04	1.13	1.10	0.89	1.26	1.21	1.38	1.23	1.06	2.19	2.02
MnO	0.16	0.17	0.17	0.16	0.18	0.18	0.18	0.18	0.16	0.20	0.19
P2O5	0.09	0.08	0.08	0.07	0.09	0.07	0.09	0.09	0.09	0.27	0.28
Total	99.42	(qq.3)	(99.7)	(98.5)	(01.5)	(103.6)	) (iov4)	(101.0)	99.43	98.03	(aq·1)
LOI	-0.35								-0.41	1.31	
Mg#	<b>6</b> 5.3	(60)	(60)	(62)	(59)	<b>(63</b> )	( <b>59</b> )	(61)	<b>6</b> 6.0	44.6	(51)
Nb	6	6	6	4	6	7	7	6	5	20	19
Zr	51	55	53	43	60	59	65	59	51	129	126
Ŷ	18.3	20	20	16	22	21	25	22	19	30.2	34
Sr	173	175	170	170	175	177	178	175	173	246	249
Zn	67	69	68	67	77	80	78	73	66	101	89
Ni	140	140	157	180	132	154	126	150	143	60	62
V	256	265	259	229	288	287	312	275	257	352	330
Cr	406	415	416	465	377	402	360	388	399	126	121
Ga	17	16	16	14	17	15	17	17	19	20	19
Sc	43	42	47	43	48	46	47	43	43	43	43
Со	49	49	49	46	51	51	54	51	45	56	53
Rb	1.1								3.5	2.9	
Ba	31								73	79	
La	3.9	4	5	6	6	4	4	4	8	12.6	11
Ce	9.8	12	11	10	7	8	9	11	14	30.7	33
Pr	1.4									4.2	
Nd	7.9	8	4	6	7	3	12	8	7	20.4	18
Sm	2.2									5.1	
Eu	0.9					•				1.9	
Gd	3.0									5.8	
Dy	3.5									6.1	
Но	0.7									1.1	
Er	1.9									3.2	
Yb	1.9									2.9	
Lu	0.3									0.4	
Data sour	'Ce										
<b>Majors</b>	F	Ρ	Ρ	Р	Р	Ρ	Ρ	Р	F	F	Р
REE	Е	D	D	D	D	D	D	D	D	E	D
										_	-
Pet type Alteration	3a 1	3a	<b>3a</b>	3a	3a	3a	<b>3a</b>	3a	3a	3c	Зс

							*		*			
Sample #	A503	A504	<b>A5</b> 01	A502	A175	A382	A383	A385	A386	A169	A358	A364
Мар	B1-3	B1-3	B1-3	B1-3	B1-1	B1-1	B1-1	B1-1	B1-1	B1-2	B1-2	B1-2
Unit	Húsmuli	Húsmuli	Húsmuli	Húsmuli	Nes	Nes	Nes	Nes	Nes	Lóm	Lóm	Lóm
						<i>c</i>						<b>,</b> ,
SiO2	(49.7)	(48.9)	(49.7)	(49.7) (	47.7)	(49.0)	47.77	(47.9)	47.56	(46.6)	(49.5)	(47.9)
Al2O3	(13.5)	(12.8)	([3·l)	(3.2)	(13.7)	(3.9)	13.25	(13.7)	13.84	(10.6)	(14.7)	(13.7)
Fe2O3	12.94	14.16	14.55	13.74	14.31	15.04	14.69	16.04	13.70	<b>6.9</b> 6	11.71	13.96
MgO	(63)	(7.5)	(7-5)	(7.6)	(7.5)	(7·5)·	6.38	(7.6)	6.29	(3.6)	(8.4)	(7.6)
CaO	12.21	11.46	11.51	12.06	11.45	11.32	10.96	10.77	11.33	14.77	13.19	11.11
<b>Na2O</b>	2.33	2.09	2.19	2.20	2.62	2.72	2.38	2.65	2.63	1.92	2.24	2.32
<b>K2O</b>	0.14	0.15	0.14	0.12	0.28	0.29	0.44	0.31	0.32	0.08	0.11	0.23
TiO2	2.23	2.53	2.51	2.44	2.25	2.41	2.34	2.67	2.32	1.04	1.17	2.02
MnO	0.19	0.19	0.21	0.19	0.22	0.24	0.22	0.25	0.21	0.10	0.19	0.20
P2O5	0.30	0.34	0.28	0.31	0.23	0.20	0.27	0.23	0.26	0.10	0.08	0.17
Total	(99 <i>.</i> 8)	(101·7)	(101.0)	(100 <u>j</u> )	(100-3)	(102.6)	<b>98</b> ,70	(102·1)	98.44	(85.7)	(101.3)	(49.2)
LOI			2.22	•			-0.10		4.11			-
Mať	(51)	(52)	(52)	(52)	(52)	(51)	A7 7	(50)	46.1	(52)	(m)	(53)
un Am	$(\mathbf{u})$	(12)	(		(02)	C)	-1.1			(32)		
Nb	21	22	22	22	22	21	21	25	25	6	8	10
7	127	135	141	131	133	130	140	163	164	47	58	118
V	35	38	39	35	31	37	34.7	A1	<u>41</u>	14	21	30
Sr.	248	233	231	241	228	211	211	236	235	307	155	223
31 7n	100	100	112	103	103	113	110	123	121	50	77	102
Ni	51	52	50	88	70	60	65	62	58	45	8	71
V	330	384	374	361	388	428	422	426	424	182	288	382
Cr	130	204	184	102	148	80	01	420	04	146	263	72
Ge	10	10	18	18	17	18	10	19	20	18	16	20
5~	47	40	44	47	30	43	46	40	30	35	47	38
Co	55	50	61	60	61	69	68	73	71	21	52	61
Rh			01		01	~	17	10	71			01
Re							03		103			
	11	14	14	13	10	15	14 5	13	15	5	7	10
	37	43	35	31	35	38	14.0 22 g	37	42	14	8	27
Dr	57	÷		01			A 5	07	76	14	U	<b>L</b> 1
	24	25	23	21	18	21	7.5 21 0	21	23	ß	A	18
Sm	24	25	20	21	10	21	55	21	20	U	-	10
Sin Eu							1.0					
Cd							6.4					
							6.0					
Ho.							13					
Er Cr							3.7					
EI Vh							34					
т <b>и</b> Гн							0.4					
							0.5					
Data sourc	ce											
Majors	Р	Р	Ρ	Р	Ρ	Ρ	F	Ρ	F	Ρ	Ρ	Ρ
REE	D	D	D	D	D	D	Е	D	D	D	D	D
Pet type Alteration	Зс	Зс	3c	3c	1a	1a	1a	1a	1a	<b>2a</b>	2a	<b>2a</b>

Sample # Map	<b>A365</b> B1-2	<b>A387</b> B1-2	<b>A381</b> B1-1	<b>A174</b> B1-1	<b>A376</b> B1-1	<b>A377</b> B1-1	<b>A378</b> B1-1	<b>A380</b> B1-1	<b>A172</b> B1-2	<b>A359</b> B1-2	<b>A360</b> B1-2	<b>A36</b> 1 B1-2	<b>A362</b> B1-2
Unit	Lóm	Lóm	Há	Há	Há	Há	Há	Há	Kross	Kross	Kross	Kross	Kross
SiO2 Al2O3 Fe2O3	(49·5) (14·0) <b>15.10</b>	(48.8) (13.7) 14.96	(50-8) (14·6) <b>13.96</b>	50.21 13.36 13.79	(50·3) (13·7) 14.48	(50 4) (13 8) <b>14.67</b>	(So.8) (14.3) <b>13.87</b>	(49.7) (12.8) <b>15.30</b>	(47.3) (13·2) <b>13.13</b>	(46·6) (12·8) <b>13.40</b>	(45-8) (13-2) <b>11.99</b>	(49.4) (13.0) 12.85	(47.9) ((3·4) 13.45
MgO	(8.6)	(8.4)	(7.4)	5.77	(6·7)	(7.2)	(6·5)	(5.6)	(5.6)	(5·7)	(5.3)	(4.6)	(50)
Na2O	2.57	244	2.99	2.83	2.76	2.81	3.02	3.27	2.14	2.33	2.15	2.43	2.50
K20	0.19	0.18	0.47	0.46	0.38	0.40	0.42	0.74	0.33	0.42	0.35	0.36	0.45
TiO2	2.13	2.23	1.83	1.92	2.10	1.98	2.08	2.45	2.29	2.55	2.29	2.40	2.68
MnO	0.22	0.23	0.23	0.23	0.25	0.23	0.24	0.28	0.21	0.21	0.19	0.22	0.22
P205	0.22	0.21	0.28	0.29	0.22	0.21	0.22	0.61	0.21	0.25	0.21	0.22	0.27
Total LOI	(1041)	(102.8)	(19.1)	99.11 -0.50	(101.6 <u>)</u>	(10,2.3)	((02.3)	(98-7)	(95.8)	(95.1)	(92.5)	(98·2)	(96 <i>·7</i> )
Mg#	<b>(54</b> )	(54)	(52)	46.7	(49)	(51)	(49)	(44)	(47)	(47)	(48)	( <b>43</b> )	(44)
Nb	18	18	27	25	24	23	25	48	20	23	21	21	25
Zr	125	122	204	1 <b>9</b> 8	153	153	172	310	132	149	131	137	1 <b>6</b> 9
Y	35	35	46	44.5	43	38	44	74	29	34	30	31	35
Sr Z-	225	227	188	189	198	197	192	252	307	283	292	313	304
ZN Ni	107	108	118 50	56	63	113 56	51	104	97 77	90 61	69 68	97 60	107 54
V	417	402	324	310	369	380	363	159	324	342	310	331	334
Cr	92	89	45	45	75	65	75	35	147	119	155	157	107
Ga	17	19	20	17	19	19	18	21	17	22	16	21	21
Sc	47	47	42	<b>39</b>	39	38	43	27	33	35	34	36	32
Co	68	67	61	58	<b>6</b> 5	63	62	60	56	58	51	54	59
Rb				11.3									
1581 Lo		15	10	128	16	15	16	20	15	16	10	10	12
Ce		26	27	42.7	36	37	41	81	38	35	40	33	43
Pr		20		5.7		0.		01			-10	~	~~
Nd		17	17	26.8	23	20	25	47	23	22	23	20	23
Sm				6.6									
Eu				2.2									
GQ				7.6									
Ho				0.0									
Er				4.7									
Yb				4.5									
Lu				0.7									
Data sour Majore	r <b>ce</b> P	р	P	F	P	P	Р	P	P	Þ	P	P	P
REE	ı	D	D	Ē	D	D	D	D	D	D	D	D	D
Pet type Alteration	2a 1	2a	1a	1a	1a	1a	1a	1a	2b	<b>2</b> b	2b	<b>2</b> b	<b>2</b> b

					*		$\star$	*			
Sample a	F A363	A173	A366	A367	A368	A388	A379	A167	A168a	A168b	A170
Мар	B1-2	B1-6	B1-5	B1-5	B1-5	B1-5	B1-1	B1-2	B1-2	B1-2	B1-2
Unit	Kross	Bæ	Bæ	Bæ	Bæ	Bæ	Hvithlið	Hvithlið	Hvithliö	Hvithlið	Hvithlið
SiO2	(47.1)	(48.0)	(48.0)	(47.9)	47.33	(49.1)	47.29	48.05	(49-3)	(47. i)	(4.9.1)
A1203	(12.2)	(12.9)	(ino)	(iz.x)	14.46	(4.1)	17.50	16.37	(11.5)	(IE A)	(14.2)
Fe203	14.04	13.95	12.49	10.98	12.69	12.57	9.99	11.18	11.16	11.63	11 01
MaO	(5.0)	(G.I)	(5.8)	(5.2)	6.18	(5.3)	7.03	7.08	(6.9)	(9.0)	(7.7)
CaO	10.69	11.66	12.04	12.93	11.65	11.46	13.04	12.95	12.91	12.07	13.90
Na2O	2.02	2.49	2.53	2.32	2.53	2.68	1.97	1.90	2.44	2.23	2.10
K20	0.40	0.36	0.36	0.26	0.30	0.50	0.16	0.14	0.15	0.14	0.09
TiO2	2.51	2.61	2.26	1.91	2.30	2.49	1.35	1.26	1.26	1.24	1.11
MnO	0.23	0.22	0.19	0.16	0.19	0.20	0.15	0.17	0.19	0.17	0.18
P205	0.22	0.22	0.21	0.18	0.25	0.31	0.13	0.11	0.16	0.13	0.07
Total	(91, .7)	(100.1)	(97.9)	(95.7)	97.90	(986)	98.60	99.20	(99.0)	(ag.7)	la r
LOI	C 14 17	0000	Cr y	(12-1)	0.08		0.18	0.06	(119		(17.6)
Mg#	(43)	(50)	(49)	(50)	50.5	(47)	59.6	57.1	(56)	(62)	(59)
Nb	25	24	21	17	20	20	44	10	0	0	7
7,	20 164	24 156	122	112	140	129	וו גיד	69	0 70	9 70	7
v	36	36	30	25	20.3	32	20	22	24	20	20
Sr.	312	295	208	20	29.0	200	20	210	24 107 ·	102	160
31 7n	107	112	250	9/ 8/	204	250	2 <del>40</del> 76	73	74	81	70
Ni	67	60	_ <del>5</del> 5 _ 68	78	30 72	79	08	84	60	86	70
V	327	354	323	70 275	340	330	260	204	284	208	78 275
Cr	128	161	152	158	153	152	200	192	154	180	273
Ga	18	21	23	20	18	22	19	18	17	18	15
Sc	34	40	34	36	38	33	36	42	42	38	45
Co	60	60	51	44	56	53	43	48	50	50	45
Rb			•••		9.9		1.2	3.2			
Ba					125		79	75			
La	15	11	12	9	14.5	13	9	7	8	9	5
Ce	42	26	39	29	34.4	43	18	15	14	14	12
Pr					4.7						
Nd	21	19	21	17	22.4	29	15	13	11	9	7
Sm					5.5						
Eu					1.9						
Gd					6.1		•				
Dy					6.2						
Ho					1.1						
Er					3.1						
Yb					2.7						
Lu					0.4						
Data sour	ce	·									
<b>Majors</b>	Ρ	Ρ	Ρ	Ρ	F	Ρ	F	F	Р	Р	Р
REE	D	D	D	D	Ε	D	D	D	D	D	D
Pet type Alteration	2b	2c	2c	2c	2c	2c	2c	2c	2c	2c	2c

					*
Sample #	A171	A371	A372	A373	A374
Map	B1-2	B1-2	B1-2	B1-2	B1-2
Unit	Hvithliö	Hengill	Hengill	Hengill	Hengill
		Ŭ	J	Ū	Ū
SiO2	(47.1)	(ຣາ 🗉	(47.3)	(44·8)	47.57
A1203	(i. 3)	(12.4)	(11.0)	(15.0)	14.75
Fe2O3	10.13	13.35	13.83	14.29	13.06
MaO	(7.3)	(4:7)	(7.2)	(12.8)	5.96
CaO	13.74	12.51	12.02	10.36	12.15
Na2O	1.94	2.00	2.34	1.64	2.10
K20	0.08	0.24	0.26	0.26	0.22
TiO2	1.02	2.31	2.34	1.81	2.20
MnO	0.16	0.22	0.21	0.20	0.18
P205	0.10	0.19	0.16	0.14	0.10
Total	(45.8)	(aq.1)	(ag. 7)	(m 2)	
	(13-0)	(troj	(1)+)	((01.5)	0.09
LUI					0.80
Mg#	(60)	(42)	(52)	(65)	<b>48.9</b>
Nb	6	20	20	15	19
Zr	51	120	112	90	112
Y	19	32	30	23	29.1
Sr	176	267	254	238	250
Zn	69	111	108	98	104
Ni	75	144	91	266	101
V	254	358	343	270	333
Cr	212	284	265	551	280
Ga	17	17	19	18	18
Sc	45	40	41	34	44
Co	42	68	61	73	58
Rb					2.9
Ba					89
La	6	10	12	10	12.5
Ce	13	32	36	27	29.6
Pr					4.1
Nd	10	18	21	17	19.8
Sm			-		4.9
Eu					1.8
Gd					5.6
Dv					5.9
Ho					1.1
Er					3.1
Yb					2.8
Lu					04
					0.7
Data sou	rce				
<b>Majors</b>	Ρ	Ρ	Р	Р	F
REE	D	D	D	D	Е
Pet type Alteration	2c n	<b>3a</b>	<b>3a</b>	3a	3a

Sample # Unit Map ref	<b>A161</b> B1-6 Stekkás	<b>A163</b> B1-6 Stekkás	<b>A533</b> B1-6 Kyllisfell	<b>A534</b> B1-6 Kyllisfell	A164 B1-6 Stórhóll	A165 B1-6 Stórhóll	A166 B1-6 Stórhóll
5102	(19.2)	(48.4)	(52.7)	(50.8)	(49.7)	(49 · 1)	(47.9)
A1203	(40.2)	(10,1)	(12.6)	(13·4)	(13.0)	(14.4)	(4.2)
	11 55	11.58	13.13	9.81	14.14	13.09	12.47
	(7.1)	(6.5)	(5.9)	(3.8)	· (6 · 4)	(8.4)	(7.0)
myo CoO	12.33	12.50	10.77	11.92	11.32	12.72	12.29
	2.54	2.59	1.39	1.89	2.27	2.19	2.36
K2O	0.20	0.20	0.13	0.13	0.35	0.12	0.18
TiO2	1.36	1.45	2.15	2.47	2.16	1.65	1.57
MnO	0.18	0.18	0.18	0.11	0.23	0.20	0.19
P205	0.17	0.18	0.24	0.32	0.20	0.12	0.13
Total	(97.8)	(97.8)	(99 2)	(94:7) ·	(99 · 8)	(101.9)	(98·3)
Mg#	(56)	(54)	(48)	(45)	( <b>49</b> )	(57)	(54)
Nb	10	11	22	24	22	12	11
Zr	80	79	119	136	140	82	83
Ŷ	23	25	28	54	35	25	24
Sr	194	208	282	316	243	186	182
Zn	78	80	104	134	112	95	89
Ni	69	73	86	57	72	87	78
V	296	301	354	401	400	333	320
Cr	172	166	250	262	160	230	217
Ga	18	17	19	18	16	17	18
Sc	41	38	41	40	38	45	43
Co	51	51	53	36	64	58	53
Rb							
Ba							
La	9	9	14	30			
Се	29	20	27	61			
Pr							
Nd	15	14	21	42			
Sm							
Eu							
Gd							
Dy							
Но							
Er							
Yb							
Lu							
Data sou	rce _	_	-	-	D	D	P
Majors	P	P	4	r n	г	r	•
REE	D	D	D	U			
Pet. type	2c	2c	2c	2c	2c	2c	2c
Alteratio	n		A1+2	A)+2			

			*				*			
Sample #	A76	A77	A146	A562	<b>A9</b> 1	<b>A90</b>	A249	A262	A260	A71
Map ref	B2-3	B2-3	B2-2	B2-2	B2-3	B2-3	B2-3	B2-3	B2-3	B2-8
Dyke type	<b>'A'</b>	<b>'A</b> '	<b>'A'</b>	<b>'</b> A'	<b>'A</b> '	<b>'A'</b>	'B'	<b>'</b> B'	'C/B'	<b>'C'</b>
SiO2	(i.e.)	(1.0.2)	48 59	(11.2)	(47.6)	(49.8)	62 08	(56.8)	(5(.))	(43.2)
A1203	(4-0-2)	$(4^{-1}, 5)$	17 20	(++)	(11.9)	(1,2)	13 46	(12.1)	(12.1)	(10.9)
	10.61	11 45	11 00	14 10	12 79	13.90	9.80	10.25	10.96	17 86
MaO	(5.9)	(K·W)	3 79	6	(4.4)	(q.a)	1.07	(2.0)	6 2	(a.d)
myo CaO	14 64	13.85	12 75	17.33	13.24	1287	5 53	5 20	654	13 20
Na2O	1 87	2 04	2 07	0.24	1 71	1 64	3.97	3.30	1 76	1.09
K2O	0.17	0.08	0.32	0.16	0.16	0 13	1.63	1 47	2 17	0.02
TIO2	2.00	2.08	1.86	1.95	1 64	1 74	1 29	1.06	1 50	3.71
MnO	0.17	0.18	0.17	0.35	0.21	0.22	0.23	0.26	0.26	0.24
P205	0.31	0.29	0.23	0.12	0.15	0.17	0.34	0.17	0.30	0.33
Total	(ac.2)	60.20	97.96		67.1)	(100 K)	99.38	62 0	(0(.1)	199.0
		(10.1)	1 53	(05)	.(		6 65	(95.6)		CT (3)
			1.00				0.00			
Mg#	(54)	(54)	42.0	(64)	(56)	(57)	18.6	(29)	(39)	( 51)
Nb	16	18	16	12	10	12	7 <del>9</del>	89	77	29
Zr	111	120	113	88	71	84	658	1002	627	195
Y	27	27	28	28	22	26	116	128	114	48
Sr	304	306	283	182	201	188	209	224	170	200
Zn	80	<b>8</b> 5	89	107	81	94	165	195	167	127
Ni	80	<b>6</b> 5	56	90	78	78	19	27	20	43
V	278	300	347	436	335	372	63	41	57	565
Cr	166	139	154	269	160	172	1	3	-3	75
Ga	17	17	18	16	21	20	23	25	22	18
Sc	41	42	37	46	41	44	18	16	18	48
Co	45	48	52	55	55	62	34	32	38	74
Rb			5.0				38.7			
Ba			107				132			
La	8	11	11	5	6	9	46	53	49	13
Ce	26	34	32	19	18	18	121	136	122	33
Pr										
Nd	16	23	18	11	10	11	71	76	68	20
Sm										
Eu										
Gd										
Dy										
Но										
Er				•						
Yb										
Lu										
Data sour	ce									
Majors	P	Р	F	Р	P	P	F	Р	Р	Р
REE	D	D	D	D	D	D	D	D	D	D
	-	-	-	-	-	-	-	-	2	~
Pet. type	<b>2</b> b	<b>2</b> b	<b>2</b> b	1b	<b>2a</b>	1a	1b	1b	1b	1b
Alteration				A2		A2			A2	A2

							$\star$		*	
Sample #	A72	A78	A103	A216	A560	A217	A183	A340	A225	A312
Map ref	B2-8	B2-3	B2-7	B2-7	<del>B</del> 2-2	B2-7	B2-6	B2-1	B2-6	B2-8
Dyke type	'C'	'C'	<b>'C'</b>	'C'	<b>'C</b> '	<b>'C</b> '	<b>'C</b> '	<b>'C</b> '	'D'	'E'
SiO2	(48.6)	(48.7)	(46.1)	(50 z)	(46.3)	(47.8)	47.61	(480)	63.48	(49.2)
A1203	(12.2)	(10.a)	(9.7)	(11.2)	(15.2)	(10.7)	13.53	(11-7)	14.24	$(12, \neq)$
FeOT*	13.93	17.79	19.02	17.74	14.68	19.51	16.12	15.84	10.58	17.56
MaQ	(8.5)	(5.X)	(9.5)	(6.2)	(8.9)	(7.3)	4.09	(1.2)	1.51	(5.9)
CaO	12.60	10.37	12.91	9.14	11.72	9.55	12.82	10.92	5.51	10.05
Na2O	2.06	2.09	0.93	2.04	0.13	1.78	1.44	1.51	2.29	2.97
K20	0.09	0.15	0.09	0.20	0.14	0.10	0.06	0.10	0.13	0.53
TiO2	2.35	3.73	4.08	3.55	2.87	3.79	2.89	2.33	1.66	3.63
MnO	0.23	0.25	0.39	0.30	0.30	0.30	0.30	0.25	0.24	0.29
P205	0.25	0.41	0.42	0.58	0.26	0.52	0.37	0.23	0.55	0.40
Total	(out)	(mar)	han	(((1))	(100.1)	(	99.20	(102.2)	100.19	(103.2)
LOI	(101.4)	((201))	((05.1)	(4.2)	(ICD-4J	((01.4)	10.67		7.77	
Mať	(56)	(40)	(51)	(12)	/ <b>56</b> )	(44)	347	(၈၈)	23.0	(41)
mðu.	(30)	(40)	(31/	(42)	(30)	(44)	JH.7	(00)	23.0	(41)
Nb	18	30	31	42	27	33	25	20	78	31
Zr	136	200	215	285	176	209	170	124	572	208
Y	38	50	60	<b>6</b> 5	44	53	43	30	120	53
Sr	189	257	237	268	136	261	161	228	164	233
Zn	109	149	160	154	151	154	137	114	180	142
Ni	80	50	59	37	46	50	58	58	13	42
V	399	513	577	365	482	500	486	452	<b>6</b> 5	463
Cr	181	54	58	23	112	69	130	102	51	62
Ga	16	17	19	20	18	22	19	21	24	20
Sc	48	36	51	34	46	39	46	44	22	41
Co	65	78	85	73	60	78	65	69	42	12
KD							10		1.1	
58	10	10	17	24	10	17	10		49	14
	20	18 51	47	63	12	17	10		40	1 <del>4</del> 52
Dr	30	51	4/	00	30		30		125	32
Nd	15	26	23	31	21	25	17		72	28
Sm	10	20	20	0.		2.	••.			
Eu										
Gd										
Dy										
Но										
Er										
Yb										
Lu										
Data source	e									
Majors	Р	Р	Р	Р	Р	Р	F	Р	F	P
REE	D	D	D	D	D	D	D		D	D
Pet. type	1b	1b	1b	1b	1b	1b	1b	1b	1b	1b
Alteration			A1+2		A1+2		A1+2		A2	

-			$\star$							*
Sample #	A561	A100	A182	A101	<b>A61</b>	A147	A224	A200	A326	A275
Map ref	B2-3	B2-2	B2-6	B2-7	B1-8	B2-2	B2-3	B2-6	B2-2	B2-1
Dyke type	'E'	'E'	'E'	'E'	'E'	'E'	'E'	'G'	(DB9)	(DB8a)
SiO2	(I.a. D)	(50-3)	47.51	(49-0)	(47.5)	(49.2)	(53.9)	(50.5)	(50.0)	49.06
AI2O3	(12.5)	(12.2)	13.05	(135)	(14.0)	(14.6)	(12.4)	(13.9)	(IL.G)	15.00
FeOT*	15.13	15.77	16.77	14.41	14.67	12.69	12.78	14.02	13.00	<b>11.8</b> 8
MgO	(8.1)	(6·9)	5.49	(9.1)	(10.7)	(9.4)	(5·I)	(7.9)	(9.2)	5. <b>6</b> 6
CaO	12.81	<b>9.1</b> 5	10.48	11.92	11.36	12.20	7.97	12.06	12.79	14.48
Na2O	2.11	2.43	2.37	2.07	2.21	1.81	1.49	2.30	2.13	2.07
K2O	0.11	0.19	0.11	0.11	0.13	0.08	0.08	0.25	0.16	0.08
TiO2	2.50	2.59	3.28	2.27	1.91	1.56	2.44	2.00	1.52	1.38
MnO	0.25	0.23	0.25	0.23	0.22	0.18	0.17	0.22	0.20	0.19
P2O5	0.23	0.35	0.42	0.22	0.15	0.15	0.61	0.21	0.13	0.15
Total	(102.8)	(100.2)	99.71	(102.q)	(102.8)	(101.9)	(96 .9)	(103.4)	(103.7)	99.94
LOI			3.83							4.35
Mg#	(53)	(48)	40.7	(57)	(60)	(61)	( <b>46</b> )	(54)	( <b>60</b> )	50.0
Nb	18	34	28	19	12	11	64	14	12	9
Zr	132	257	190	127	94	84	454	117	81	73
Y	37	5 <del>9</del>	49	36	29	24	100	35	27	24
Sr	250	331	222	<b>186</b>	151	185	212	175	184	205
Zn	120	130	139	105	108	86	174	101	95	76
Ni	61	41	49	79	82	87	17	66	87	94
V	445	391	486	398	402	331	131	387	333	286
Cr	91	28	92	173	138	211	3	103	233	201
Ga	17	18	20	18	19	19	23	20	19	17
Sc	47	37	44	46	42	43	29	43	44	43
Co	65	69	71	68	71	57	47	59	55	51
RD			2.4							1.7
	10	21	30	44	e		22	0	6	52
	20	59	10	20	23		33 83	0 29	21	10
Dr .	29		40	20	20		80	20	21	18
Nd	20	32	25	20	14		54	17	14	10
Sm	20	~	20	20	.4		04	.,	14	10
Eu										
Gd										
Dv										
Ho										
Er										
Yb										
Lu										
Data source										
Majors	Р	Р	F	Р	Р	Р	Р	Р	Р	F
REE	D	D	D	D	D	-	D	D	D	D
Pet. type	19	19	19	1h	10	19	15	2h	2h	- 2a
Alteration	14	14	14		19	10	A1+2	لاع	CH I	A1

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Sample #	<sup>3</sup> A520	A59	<b>A6</b>	A7	A10	A18	A62	A53	A106	A160
Map ref	B2-2	B1-8	B3-2	B3-2	B3-6	B3-8	B3-4	B3-4	B2-7	B2-7
Dyke type	NG	NG	(ILS)	(ILS)	(ILS)	(ILS)	(ILS)	(GB)	<b>(</b> S4)	(S4)
SiO2	(48.1)	(48.3)	(46.0)	(46.5)	(465)	(46.1)	(44.8)	(48.5)	(49.0)	(49.1)
Al2O3	$(11 \cdot 2)$	(15.6)	(14.1)	(14.9)	(4:7)	(14:8)	(5.0)	(156)	(12.6)	(12.8)
FeOT*	16.08	13.48	10.86	12.66	12.13	12.98	13.26	12.85	14.05	13.84
MgO	(7.8)	(11.7)	(6.0)	(9.0)	( <b>8</b> ·2)	( <del>7</del> ·2)	(7.7)	(10.0)	(9.1)	( <b>8</b> · I)
CaO	11.28	10.53	12.37	11.88	11.93	10.96	10.79	11.83	12.37	11.62
Na2O	1.66	1.57	2.27	2.13	2.12	2.53	2.38	2.04	2.14	2.23
K20	0.16	0.05	0.13	0.13	0.13	0.27	0.27	0.13	0.25	0.22
1102	3.24	1.79	1.69	1.65	1.68	2.21	2.18	1.50	2.16	2.25
MnO	0.28	0.20	0.17	0.20	0.19	0.20	0.20	0.20	0.25	0.23
P200	0.42	0.19	0.19	0.10	(~ <del>7</del>	0.20	(acc)	(1000)	0.19 (102 I)	$(\infty, \mathcal{Z})$
LOI	(100.3)	(103.4)	(43.7)	(44.1)	(4++)	(47.5)	(40.6)	(1028)	(02.1)	(100 7)
	/ 、		<i>.</i> .	<i>.</i> .		<b>A</b> .		<i>,</i> , ,		, ,
Mg#	(51)	(65)	(54)	(60)	(59)	(54)	( 55)	( <b>62</b> )	(58)	(55)
Nb	22	11	15	13	14	19	19	10	17	16
Zr	159	80	101	91	<b>9</b> 5	123	123	71	115	118
Y	39	28	28	30	27	29	30	24	31	30
Sr	249	134	205	160	180	248	246	171	209	209
Zn	107	94	77	89	80	92	101	90	109	107
	53	112	5/	100	129	159	128	124	73	58
V Cr	- 503 - 00	340	291 176	307	300	324	315	337	435	43 I 96
	10	104	18	424	15	442	327 18	10	90 16	18
Sc	38	13	37	39	39	35	36	42	41	44
Co	68		48	61	57	61	67	61	60	59
Rb				•			•			•••
Ba										
La	15	7	11	8	12	9	11	7		
Ce	43	17	28	21	20	24	36	11		
Pr										
Nd	21	10	20	17	14	13	20	10		
Sm										
Eu										
Gd										
Dy										
no Er										
Lu										
		م ر								
Data source	e									
Majors	Р	Ρ	Ρ	Р	Ρ	Ρ	Ρ	Р	Р	Р
REE	D	D	D	D	D	<b>D</b>	D	D		
Pet. type Alteration	2a	18	3a	3b	3a	3a	3a	1a	1b	1b

 ${}^{3}NG$  = not asigned to a dyke type or eruptive unit. Samples A106 and A160 are sills.

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Sample # Map ref Dyke type	<sup>4</sup> <b>A55</b> B1-8 (GA)	<b>A58</b> B1-8 (GA)	<b>A41</b> B3-4 (GB)	<b>A298</b> B2-3 (DB9)
SiO2 Al2O3 FeOT* MgO CaO Na2O K2O TiO2 MnO P2O5 Total LOI	$ \begin{array}{c} (48 \cdot 4) \\ (14 \cdot 7) \\ 14.35 \\ (11 \cdot 3) \\ 11.15 \\ 1.83 \\ 0.03 \\ 1.94 \\ 0.21 \\ 0.19 \\ (10 4 \cdot 0) \\ \end{array} $	$(\begin{array}{c} (48 \ 6) \\ (14 \cdot 3) \\ 14.23 \\ (10 \cdot 6) \\ 11.41 \\ 2.04 \\ 0.05 \\ 1.89 \\ 0.20 \\ 0.19 \\ (103 \cdot 4) \end{array}$	$\begin{array}{c} (\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!\!$	(44.3) (13.4) 12.40 (6.8) 13.20 2.08 0.17 1.63 0.22 0.13 (49.3)
Mg#	(62)	(61)	(57)	(53)
Nb Zr Y Sr Zn Ni V Cr Ga Sc Co Rb Ba La Ce Pr M S Eu Gd Ho Er Y Lu	11 89 31 148 103 88 383 141 17 41 68	12 89 30 150 102 94 387 143 18 42 68	7 64 25 161 80 109 318 247 15 40 58	13 86 26 213 91 60 332 166 18 47 52
<b>Data sour Majors REE</b>	r <b>ce</b> P	Ρ	Ρ	Ρ
Pet. type Alteration	1a 1	1a	3a	<b>2</b> b

<sup>&</sup>lt;sup>4</sup>Samples A55, A58, A41 and A298 are all sills.

## Appendix 5 Volcanic glass geochemical data set

## **Explanation**

The following appendix contains the geochemical data set for the volcanic glass samples analysed by the electron microprobe at the University of Durham. The analyses are grouped in stratigraphic order. All samples are from the Hveragerŏi Volcanic System. In Part A, the raw spot analyses are listed, and in Part B, sample averages are presented. The latter are described and discussed in Chapter 5. FeO represents total iron in the sample. All sample locations are on map C

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Group	Page					
	Part A	Part B				
Gljúfur (G)	346	351				
Tindar (T)	347	351				
Selfjall (S)	348	351				
Ingolfsfjall lava shield (ILS)	349	352				
Dykes	350	352				

#### Key to Appendix 5

Unit	= stratigraphic unit or lava shield, as described in Chapter 3.
Dyke type	= refers to one of four varieties, as defined in Figure 2.11 (p. 30). Those dykes
	that do not belong to one of these groups have titles presented in brackets. If
	this title is the name of an eruptive unit or group, it is inferred that the intrusive
	material is therefore of a similar basalt type and is therefore associated with it.
Pet type	= refers to petrological types, as defined in Chapter 4 (Table 4.3, p.87)
Spot ref.	= refers to individual spot analyses.
n (spots)	= number of spot analyses combined to give average.
Other	= If the sample has had other analytical techniques applied to it, these are
	indicated accordingly:
	WR = whole-rock analyses (refer to Appendix 4)

MIN = mineral microprobe analyses (refer to Appendix 2).

Part A	۱										
Sample #	A56	<b>A56</b>	A56	A56	A56	A57	<b>A5</b> 7	A28	A28	A28	<b>A5</b> 1
Unit	GA 2	GA 2	GA 2	GA 2	GA 2	GA 11	GA 11	GB 1	GB 1	GB 1	GB 4
Spot ref.	50.1	50.2	50.3	50.3	50.4	33.2	33.3	32.1	32.2	32.3	34.13
SiO2	47.7	47.7	47.6	47.8	47.7	47.7	48.5	48.8	49.0	49.2	48.5
Al2O3	13.5	13.3	13.3	13.4	13.6	13.7	13.8	13.7	14.3	14.7	15.5
FeO	12.9	12.9	13.2	13.3	13.0	13.2	13.5	11.9	11.8	11.6	8.6
MgO	6.8	6.9	6.7	6.6	<b>6.9</b>	6.9	6.9	8.0	7.8	8.0	9.3
CaO	11.5	11.9	11.5	11.5	11.8	11.6	11.6	12.9	12.6	12.7	14.2
Na2O	2.37	2.40	2.49	2.28	2.47	2.47	2.44	2.24	2.37	2.45	1.83
K20	0.17	0.14	0.11	0.11	0.18	0.12	0.12	0.13	0.05	0.12	0.00
ТЮ2	1.94	2.11	2.02	2.17	2.09	1.88	2.13	1.62	1.65	1.57	0.64
CrO2	0.12	0.01	0.24	0.05	0.11	0.04	0.05	0.11	0.04	0.10	0.00
MnO	0.31	0.24	0.11	0.32	0.16	0.23	0.34	0.30	0.21	0.00	0.35
P2O5	0.19	0.48	0.19	0.10	0.09	0.00	0.83	0.00	0.02	0.00	0.00
Totals	97.5	97.6	<b>9</b> 7.5	97.8	98.2	97.9	<b>99</b> .3	<b>99</b> .7	99.8	100.6	98.9
Mg #	48.5	48.8	47.4	47.1	48.6	48.2	47.8	54.5	54.1	55.1	<b>6</b> 5.9
Pet.type Other	1a	1a	1a	1a	1a	1a	1 <b>a</b>	3b WR	Зb	Зb	За

Sample #	<b>A5</b> 1	<b>A5</b> 1	<b>A5</b> 1	<b>A5</b> 1	<b>A</b> 51	<b>A5</b> 1	<b>A5</b> 1	<b>A5</b> 1	<b>A4</b> 4	<b>A44</b>	<b>A</b> 44
Unit	GB 4	GB 4	GB 4	GB 4	GB 4	GB 4	GB 4	GB 4	GB 4	GB 4	GB 4
Spot ref.	34.23	34.33	34.42	34.52	34. <b>6</b> 2	34.71	34.81	34.91	35.1	35.2	35.3
SiO2	48.4	48.9	48.5	48.5	48.5	48.2	48.5	48.7	48.9	49.0	48.7
Al2O3	15.5	15.7	15.8	15.4	15.3	15.5	15.5	15.7	15.5	15.6	15.6
FeO	8.6	8.4	8.7	8.2	8.4	8.7	8.4	8.8	8.9	9.2	8.8
MgO	9.3	9.5	9.3	9.2	9.0	9.1	9.5	9.3	9.0	9.2	9.1
CaO	14.0	13.9	14.3	13.9	13.8	14.1	13.9	13.9	14.2	14.0	13.9
Na2O	1.76	1.78	1.84	1.82	1.85	2.00	1.95	1.82	1.85	1.84	1.72
к20	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.05	0.04	0.03
ТЮ2	0.63	0.65	0.69	0.78	0.65	0.67	0.68	0.64	0.66	0.67	0.57
CrO2	0.00	0.00	0.00	0.22	0.28	0.00	0.00	0.00	0.00	0.07	0.04
MnO	0.00	0.00	0.23	0.00	0.00	0.23	0.25	0.00	0.05	0.00	0.00
P2O5	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.07	0.00
Totals	<del>98</del> .5	<del>9</del> 8.9	99.5	98.2	<b>98</b> .0	98.6	98.8	99.2	99.0	100.0	<b>98</b> .5
Mg #	65.8	66.8	65.6	<b>66</b> .7	<b>6</b> 5.7	<b>6</b> 5.0	66.9	<b>6</b> 5.4	64.4	64.0	64.8
Pet_type Other	За	За	За	За	3a	За	За	3a	3a Min	За	За

Sample #	<b>A6</b> 7	<b>A</b> 67	<b>A</b> 67	<b>A6</b> 7	A148	A148	A148	A148	A148	A148	A149
Unit	TA 1	TA 1	TA 1	TA 1	тв з	тв з	тв з	тв з	твз	тв з	тв з
Spot ref.	29.1	29.2	29.3	29.4	3.103	3.203	3.301	3.401	3.502	3.602	5.101
SIO2	48.6	48.8	49.1	49.3	49.8	49.9	50.6	49.3	48.8	49.6	48.5
AI2O3	13.4	13.7	13.5	13.4	13.7	13.6	14.1	13.2	13.1	13.7	13.7
FeO	12.6	12.6	12.8	12.6	13.0	12.9	13.0	12.7	12.6	12.6	12.5
MgO	6.1	6.1	5.9	6.2	6.1	6.0	5.9	6.0	6.1	6.1	6.0
CaO	11.1	11.1	11.2	11.0	11.2	11.0	11.1	10.8	10.9	11.0	11.4
Na2O	2.53	2.42	2.55	2.72	2.64	2.80	2.81	2.76	2.84	2.61	2.58
K20	0.44	0.48	0.52	0.51	0.51	0.49	0.57	0.50	0.39	0.52	0.41
TiO2	2.34	2.44	2.25	2.17	2.34	2.43	2.65	2.28	2.38	2.31	2.15
CrO2	0.05	0.00	0.09	0.10	0.15	0.00	0.12	0.13	0.00	0.00	0.00
MnO	0.32	0.26	0.02	0.26	0.11	0.13	0.29	0.16	0.24	0.18	0.30
P2O5	0.16	0.27	0.20	0.02	0.10	0.26	0.46	0.11	0.16	0.12	0.08
Totals	97.6	98.2	98.2	<b>98</b> .3	<b>99</b> .7	<b>9</b> 9.5	101.6	97.9	97.6	<b>98</b> .7	<b>9</b> 7.7
Mg #	46.2	46.3	45.2	46.8	45.5	45.5	44.9	45.8	46.3	46.4	46.2
Pet.type Other	1a WR	1a	1a	1a	<b>2</b> b	2b	2b	2b	2b	2b	2b

Sample #	A149	A149	A149	A149	A149	A149	A122	A122	A122	A122	A122
Unit	тв з	тв з	тв з	тв з	тв з	тв з	TB 5	TB 5	TB 5	TB 5	TB 5
Spot ref.	5.201	5.302	5.402	5. <b>50</b> 3	<b>5.60</b> 3	5.704	40.1	40.2	40.3	40.4	40.5
<b>SIO2</b>	48.3	48.6	48.7	48.2	48.3	48.0	48.8	49.3	49.9	49.9	49.6
AI2O3	13.7	13.7	13.6	13.3	13.6	13.5	14.1	14.4	14.7	14.3	14.5
F=O	12.9	13.2	12.9	12.7	12. <del>9</del>	13.5	11.5	11.6	11.7	12.0	12.1
MgO	6.0	6.2	6.2	6.0	6.1	6.0	7.4	7.6	<b>7.8</b>	7.7	7.7
CaO	11.3	11.3	11.5	11.1	11.2	11.0	12.8	12.8	12.9	12.9	13.2
Na2O	2.49	2.57	2.60	2.58	2.62	2.48	2.30	2.35	2.48	2.32	2.34
к20	0.48	0.46	0.56	0.48	0.39	0.49	0.24	0.11	0.24	0.17	0.15
TiO2	2.35	2.31	2.13	2.19	2.39	2.24	1.60	1.73	1.71	1.68	1.71
CrO2	0.00	0.02	0.00	0.00	0.45	0.14	0.10	0.12	0.11	0.06	0.18
MnO	0.30	0.25	0.15	0.36	0.17	0.20	0.71	0.26	0.35	0.18	0.12
P2O5	0.17	0.21	0.44	0.24	0.12	0.35	0.00	0.00	0.18	0.12	0. <del>9</del> 4
Totels	97.9	<del>98</del> .8	<del>98</del> .7	97.2	97.9	97.8	<b>98.9</b>	100.2	102.1	101.3	101.8
Mg #	45.2	45.6	46.0	45. <del>9</del>	<b>45.9</b>	44.4	53.5	53.8	54.1	53.3	53.2
Pet,type	2b	2b	2ь	2b	26	2ь	3a	3a	3a	3a	За
Other							Min				

Sample #	A122	A97a	A97a	A97a	A154	A154	A154	A154	A158	A158	A158
Unit	TB 5	TC 2	TC 2	TC 2	S 5	S 5	S 5	S 5	S 5	S 5	S 5
Spot ref.	40.6	25.1	25.2	25.3	12.1	12.2	12.3	12.4	16.1	16.2	16.3
SIO2	49.2	48.8	48.2	49.1	48.8	48.8	<b>48</b> .9	48.9	49.5	48.5	48.7
Al2O3	14.0	12.9	13.1	13.6	13.9	13.6	13.9	13.6	14.1	13.8	13.9
FeO	11.9	14.4	14.2	14.7	12.9	12.5	12.0	12.6	12.2	12.2	12.2
MgO	7.7	5.7	5.8	6.0	6.3	6.3	6.4	6.2	6.6	6.5	6.4
CaO	13.1	10.7	10.7	10.6	11.2	11.2	11.1	10.9	11.1	10.9	10.9
Na2O	2.43	2.82	2.62	2.83	2.66	2.54	2.70	2.49	2.79	2.75	2.60
к20	0.14	0.42	0.42	0.36	0.48	0.38	0.42	0.53	0.42	0.40	0.40
ТЮ2	1.79	2.88	<b>2.96</b>	2.96	2.21	2.26	2.42	2.25	2.31	2.12	2.19
CrO2	0.04	0.32	0.11	0.15	0.18	0.07	0.75	0.14	0.00	0.00	0.08
MnO	0.26	0.22	0.05	0.40	0.19	0.21	0.20	0.13	0.31	0.26	0.26
P2O5	0.13	0.19	0.19	0.25	0.15	0.20	0.15	0.26	ND	ND	ND
Totals	100.6	<b>99</b> .1	98.4	100.9	99.0	<b>98</b> .0	98.3	98.1	<b>99</b> .3	97.4	97.7
Mg #	53.6	41.5	42.0	42.4	46.5	47.2	48.5	46.8	49.0	48.6	48.4
Pet.type Other	3a	1b	1b	1b	2b	2b	2b	2b	2b	2b	2b

Sample #	A158	A153	A153	A153	A153	A153	A155	A155	A155	A155	A155
Unit	S 5	S 5	S 5	S 5	S 5	S 5	S 5	S 5	S 5	S 5	S 5
Spot ref.	16.4	2.102	2.21	2.303	2.404	2.505	8.11	8.21	8.32	8.42	8.74
SiO2	49.7	50.0	48.7	49.5	49.6	49.9	48.6	49.7	48.7	49.0	48.6
A12O3	14.2	13.8	13.6	13.8	14.0	14.0	13.3	13.8	13.7	13.4	13.8
FeO	12.4	12.8	12.0	12.3	12.4	12.3	12.4	12.1	11.9	12.2	12.1
MgO	6.6	6.4	6.3	6.3	6.4	6.4	6.1	6.1	6.0	<b>6</b> .0	6.0
CaO	11.0	11.0	11.0	11.2	11.0	11.3	11.0	11.2	11.2	11.1	10.9
Na20	2.75	2.71	2.55	2.76	2.59	2.71	2.89	2.87	2.59	2.74	2.68
к20	0.50	0.45	0.46	0.53	0.45	0.43	0.50	0.46	0.41	0.38	0.44
ТЮ2	2.25	2.28	2.08	2.16	2.38	2.45	2.34	2.25	2.25	2.45	2.33
CrO2	0.24	0.11	0.10	0.03	0.00	0.07	0.00	0.00	0.00	0.00	0.00
MnO	0.19	0.21	0.10	0.23	0.18	0.23	0.29	0.00	0.34	0.00	0.34
P2O5	ND	0.44	0.12	0.22	0.10	0.19	0.00	0.00	0.00	0.00	0.00
Totals	<b>99</b> .7	100.1	97.1	<b>99</b> .0	<b>99</b> .1	100.0	97.3	98.6	97.0	97.2	97.2
Mg #	48.6	47.2	48.2	47.8	47.8	48.2	46.5	47.4	47.4	46.9	46.8
Pet.type	2b	2b	2b	2b	2b	2b	2b	2b	<b>2</b> b	2b	2b
Other		Min									

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Sample #	A155	A155	A157	A157	A157	A157	A9	<b>A9</b>	AB	A22	A22
Unit	S 5	S 5	S 8	S 8	S 8	S 8	ILS	ILS	ILS	ILS	ILS
Spot ref.	8.95	8.105	13.1	13.3	13.4	13.5	38.1	38.2	39.1	37.1	37.2
Si02	48.9	49.8	47.4	47.0	48.2	47.5	48.4	48.2	47.1	47.5	47.7
Al2O3	13.6	13.8	11.8	11.7	11.6	11.9	14.1	13.6	12.5	12.9	13.4
FeO	12.0	12.0	16.7	16.8	16.3	17.0	12.8	13.0	14.3	14.0	13.7
MgO	6.2	6.3	4.9	4.8	4.7	5.1	7.1	7.4	6.2	6.8	6.8
CaO	11.3	11.0	9.9	9.9	9.7	9.9	12.0	12.1	11.5	11.2	11.4
Na2O	2.72	2.83	3.05	2.90	2. <del>99</del>	3.09	2.36	2.53	2.45	2.45	2.77
K20	0.37	0.44	0.81	0.86	0.81	0.75	0.14	0.14	0.28	0.22	0.25
ТЮ2	2.23	2.28	4.54	4.39	4.45	4.48	2.28	2.23	3.19	3.04	3.29
CrO2	0.00	0.00	0.16	0.07	0.15	0.02	0.00	0.51	0.00	0.10	0.06
MnO	0.21	0.24	0.34	0.28	0.14	0.23	0.27	0.1 <del>9</del>	0.36	0.23	0.29
P2O5	0.00	0.00	0.48	0.56	0.76	0.65	0.11	0.17	0.15	0.09	0.05
Totals	<b>9</b> 7.5	<b>98</b> .7	100.1	<b>9</b> 9.2	<del>99</del> .8	100.6	<del>9</del> 9.6	99.6	<b>98</b> .1	98.6	<b>99</b> .7
Mg #	47.9	48.5	34.5	33.6	34.2	34.8	49.8	50.3	43.6	46.4	46.9
Pet,type Other	2b	2b	2c	2c	2c	2c	3a Min	За	За	За	3a

Sample #	A22	A16	A16	A16	A16	A16	A16	A39	A39	A39	A38
Unit	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS	ILS
Spot ref.	37.3	1.103	1.202	1.304	1.404	1. <b>50</b> 5	1.601	30.1	30.2	30.3	31.1
SiO2	47.8	48.1	47.5	47.2	47.9	47.5	47.1	47.2	47.1	47.2	48.0
Al2O3	13.4	12.3	12.0	12.3	12.5	12.5	12.3	12.8	12.6	12.7	12.6
FeO	13.6	14.6	14.8	14.7	14.5	14.9	14.8	14.2	13.9	13.9	14.4
MgO	6.9	6.3	6.1	6.2	6.3	6.1	6.1	6.4	6.3	6.2	6.4
CaO	11.4	11.4	11.2	11.4	11.5	11.6	11.2	11.3	11.1	11.3	11.4
Na20	2.47	2.64	2.43	2.64	2.64	2.44	2.39	2.62	2.44	2.50	2.46
к20	0.20	0.32	0.22	0.28	0.37	0.25	0.28	0.31	0.20	0.18	0.29
TIO2	3.06	3.30	3.44	3.63	3.53	3.54	3.59	3.26	3.43	3.24	3.42
CrO2	0.13	0.01	0.11	0.17	0.02	0.19	0.15	0.02	0.00	0.00	0.10
MnO	0.34	0.30	0.36	0.27	0.40	0.24	0.25	0.25	0.37	0.27	0.22
P2O5	0.07	0.30	0.27	0.10	0.20	0.19	0.17	0.1 <del>9</del>	0.05	0.06	0.15
Totals	<b>9</b> 9.4	<b>99</b> .5	98.4	98.8	<b>99.8</b>	<b>99</b> .5	98.4	<b>98</b> .5	97.6	97.6	<del>99</del> .4
Mg #	47.3	43.3	42.6	42.9	43.7	42.3	42.6	44.6	44.8	44.3	44.1
Pet_type Other	38	За	<b>3a</b>	3a	3a	За	3a	За	3a	3a	<b>3a</b>

Sample #	A38	A38	<b>A95</b>	<b>A95</b>	<b>A95</b>	A95	<b>A95</b>	<b>A9</b> 4	A94	A94	A94
Unit	ILS	ILS	Е	Е	Е	E	Е	с	с	с	с
Spot ref.	31.2	31.3	46.2	46.3	46.4	46.5	46.6	43.2	43.3	43.4	43.5
SI02	47.4	47.7	59.2	58.6	59.0	58.7	59.2	60.6	60.8	60.7	60.9
Al2O3	12.4	12.9	12.7	12.4	12.6	12.5	12.7	13.0	13.0	13.1	12.9
FeO	14.6	14.4	9.1	9.2	9.3	9.0	9.4	9.5	<b>9</b> .5	10.1	9.4
MgO	6.6	6.5	1.6	1.6	1.7	1.7	1.6	1.7	1.8	1.7	1.6
CaO	11.6	11.5	4.9	5.1	4.9	4.9	4.8	5.1	5.1	5.1	5.1
Na2O	2.51	2.54	3.98	3.91	3.76	4.01	3.73	4.17	3.88	3.85	3.92
K20	0.23	0.37	1.54	1.34	1.42	1.45	1.46	1.51	1.52	1.56	1.52
TIO2	3.18	3.53	1.50	1.54	1.27	1.25	1.38	1.44	1.33	1.42	1.49
CrO2	0.11	0.00	0.00	0.00	0.00	0.00	0.03	0.00	0.00	0.00	0.05
MnO	0.31	0.36	0.00	0.26	0.25	0.00	0.26	0.00	0.28	0.39	0.23
P2O5	0.05	0.04	0.17	0.95	0.25	0.19	0.31	0.31	0.22	0.31	0.23
Totals	<del>9</del> 9.1	99.8	94.9	93.9	94.4	93.8	94.8	97.6	<b>9</b> 7.5	<b>98</b> .3	97.3
Mg #	44.4	44.6	24.4	23.7	24.2	25.0	23.7	24.5	25.1	23.4	23.2
Pet.type Other	За	3a	1b	1b	1b	1b	1b	1Ь	1b	1b	1b

Sample #	<b>A94</b>	A53	A53	A53
Unit	с	(GB)	(GB)	(GB)
Spot ref.	43.6	14.1	14.2	14.3
SiO2	60.8	48.0	47.6	47.4
Al2O3	13.0	14.3	14.4	14.2
FeO	9.4	12.2	11.8	11.8
MgO	1.9	7.4	7.3	7.6
CaO	4.9	12.2	12.4	12.3
Na20	4.21	2.50	2.19	2.37
K20	1.44	0.15	0.17	0.14
TiO2	1.40	1.75	1.76	1. <b>69</b>
CrO2	0.00	0.11	0.14	0.84
MnO	0.00	0.00	0.11	0.27
P2O5	0.13	0.08	0.00	0.00
Totals	<b>97</b> .5	98.8	97.9	97.8
Mg #	26.1	52.0	52.4	53.3
Pet.type	1b	1a	1a	1a
Other		WR		

Sample #	A56	A57	A28	A51	A44	A67	A148	A149	A122	A97a	A154	A158	A153	A155
Unit	GA 2	GA 11	GB 1	GB 4	GB 4	<b>TA 1</b>	TB 3	TB 3	TB 5	TC 2	S 5	S 5	S 5	S 5
SiO2	47.7	48.1	49.0	48.5	48.9	<b>48.9</b>	49.7	48.3	49.4	48.7	48.9	49.1	49.5	49.0
A12O3	13.4	13.7	14.2	15.5	15.6	13.5	13.6	13.6	14.3	13.2	13.8	14.0	13.8	13.6
FeO	13.1	13.4	11.7	8.5	9.0	12.7	12.8	12.9	11.8	14.4	12.5	12.3	12.4	12.1
MgO	6.8	6.9	7.9	9.3	9.1	6.1	6.0	6.1	7.6	5.9	6.3	6.5	6.4	6.1
CaO	11.7	11.6	12.7	14.0	14.0	11.1	11.0	11.3	12.9	10.6	11.1	11.0	11.1	11.1
<b>Na2O</b>	2.40	2.45	2.36	1.85	1.80	2.56	2.74	2.56	2.37	2.76	2.60	2.72	2.66	2.76
K20	0.14	0.12	0.10	0.00	0.04	0.49	0.50	0.47	0.18	0.40	0.45	0.43	0.46	0.43
TiO2	2.07	2.00	1.61	0.67	0.64	2.30	2.40	2.25	1.70	2.93	2.28	2.22	2.27	2.30
CrO2	0.11	0.05	0.08	0.06	0.04	0.06	0.07	0.09	0.10	0.1 <del>9</del>	0.29	0.08	0.06	0.00
MnO	0.23	0.28	0.17	0.12	0.02	0.21	0.19	0.25	0.31	0.22	0.18	0.26	0.19	0.20
P2O5	0.21	0.42	0.01	0.00	0.02	0.16	0.20	0.23	0.23	0.21	0.19	ND	0.21	0.00
Totals	97.7	98.6	100.0	98.7	99.2	98.1	99.2	98.0	100.8	99.5	98.4	98.5	99.1	97.7
Mg #	48.1	48.0	54.6	66.0	64.4	46.1	45.7	45.6	53. <del>6</del>	41.9	47.3	48.7	47.8	47.4
n (spots)	5	2	3	9	3	4	6	7	6	3	4	4	5	7
Pet. type Other	1a	1a	Зb	3a	3a	1a	2b	2b	3a Min	1b	2b	2b	2b Min	2b

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Sample #	A157	A9	<b>A8</b>	A22	A16	A39	A38	A95	A94	A53
Unit	S 8	ILS	ILS	ILS	ILS	ILS	ILS	E	С	(GB)
SIO2	47.5	48.3	47.1	47.7	47.6	47.2	47.7	58.9	60.8	47.7
AI2O3	11.7	13.8	12.5	13.2	12.3	12.7	12.6	12.6	13.0	14.3
FeO	16.7	12.9	14.3	13.8	14.7	14.0	14.5	9.2	9.6	12.0
MgO	4.9	7.3	6.2	6.8	6.2	6.3	6.5	1.6	1.7	7.4
CaO	9.8	12.0	11.5	11.3	11.4	11.2	11.5	4.9	5.1	12.3
Na2O	3.01	2.45	2.45	2.56	2.53	2.52	2.50	3.88	4.00	2.35
K20	0.81	0.14	0.28	0.22	0.29	0.23	0.30	1.44	1.51	0.15
<b>TiO2</b>	4.46	2.25	3.19	3.13	3.50	3.31	3.37	1.39	1.42	1.73
CrO2	0.10	0.26	0.00	0.10	0.11	0.01	0.07	0.01	0.01	0.36
MnO	0.22	0.23	0.36	0.28	0.30	0.30	0.30	0.15	0.18	0.13
P2O5	0.61	0.14	0.15	0.07	0.21	0.10	0.08	0.38	0.24	0.03
Totals	99.9	99.6	98.1	99.2	99.1	97.9	99.4	94.4	97.6	98.2
Mg #	34.3	50.1	43.6	46.9	42.9	44.6	44.4	24.2	24.5	52. <b>6</b>
n (spots)	5	2	1	3	6	3	3	5	5	3
Pet. type Other	2c	3a Min	3a	3a	3a	3a	3a	1b	1b	1a WR

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# Appendix 6 Primary Melt Compositions

This appendix contains the primary melt compositions that have been used in the modelling in Chapter 5 (Section 5.6). Models 1 and 2 use compositions from Nui & Batiza (1991) and Model 2 using melts derived using the parameterisation of McKenzie & Bickle (1988).

<u>Model 1</u>	Melt 1	<u>Meit 2</u>	<u>Meit 3</u>	<u>Melt 4</u>
	20 Kbars	16 kbars	10 kbars	6 kbars
	28% PM	26% PM	20% PM	16% PM
Solidus T('C)	1314.6	1290.2	1255.2	1210.6
Pressure f(kb)	4.26	3.7	4.2	3.6
SIO2	50.48	50.93	51.32	51.82
TiO2	0.61	0.65	0.85	1.03
AI2O3	14.49	15.17	17.03	18.83
FeO	8.24	7.48	6.64	6.06
MgO	12.83	11.95	10.48	9.59
CaO	12.07	12.45	11.6	11.06
Na2O	1.41	1.52	1.98	2.48
K2O	0.11	0.12	0.15	0.18
Total	100.24	100.26	100.04	100.57
CaO/AI2O3	0.83	0.82	0.68	0.6
Mg #	75.51	75.99	75.77	75.76
Model 2	Meit 1	Melt 2	Melt 3	Meit 4
<u></u>	20 Khare	16 khara	10 kbars	6 kbara
	16% PM	16% PM	16% PM	16% PM
Solidus T(°C)	1380.5	1340	12 72	1210.6
Pressure f(kb)	15.25	12	7	3.6
SIO2	48.34	49.29	50.77	51.82
TiO2	1.03	1.03	1.03	1.03
AI2O3	16.74	17.19	17.88	18.83
FeO	9.61	8.23	6.78	6.06
MgO	11.88	11.11	10.13	9.59
CaO	10.62	10.74	10.93	11.06
Na2O	2.48	2.48	2.48	2.48
K2O	0.18	0.18	0.18	0.18
Total	100.88	100.24	100.18	100.57
CaO/AI2O3	0.63	0.63	0.61	0.6
Mg #	71	72.78	74.75	75.76
Model 3	Melt 1	Meit 2	Melt 3	
	1280°C	1330°C	1380°C	
Solidus T(°C)	1202	1255	1311	
Pressure f(kb)	0.1	4.1	8.6	
Depth (km)	0.3	12.7	26.6	
SIO2	54.99	53.65	52.14	
TIO2	0.17	0.16	0.16	
AI2O3	10.26	10.18	10.22	
FeO	7.79	7.38	7.9	
MgO	16.12	17.56	18.93	
CaO	9.16	9.15	9.17	
Na2O	0.85	0.84	0.84	
K2O	0.22	0.22	0.22	
CaO/AI2O3	0.8957	0.8988	0.8973	
Mg #	78.60066478	80.85553451	80.96419944	

Appendix 6

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This copy belongs to WALKER, C.L. 1992, Ph.D. thesis, University of Dorham.

# 13 North Atlantic ocean crust and Iceland

# CHERRY WALKER

# 13.1 Introduction

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This chapter summarizes geological, geophysical and geochemical observations from the North Atlantic basin, particularly 23°N (MARK area), 37°N (FAMOUS area), 40°N (the Azores), 45°N, 57–63°N (the Reykjanes Ridge), all relative to Iceland. Emphasis will be placed on the geochemistry of basalts from these areas, especially the nature of the source region from which the basalts originated and, to a lesser extent, what they can tell us about dynamic processes in the crust and mantle involved in their genesis. The geochemical data will also be considered in the light of other lines of evidence concerning the nature of the mantle under this ocean basin. Coverage of the South Atlantic Ocean is not within the scope of this chapter, but details can be found in Schilling *et al.* (1984), Le Roex *et al.* (1985a, b), Chaffey *et al.* (1989), Sun and McDonough (1989), ODP Leg 108 and references cited therein.

There is a general consensus about the presence of mantle plumes of various sizes under hot elevated areas in the Atlantic Ocean, and the fact that these are largely responsible for the variations in basalt geochemistry, as well as physiographic and geophysical parameters. The exact nature of the material constituting a plume mantle source, the origin of that material, and from which part of the earth's interior it may have risen, are still strongly debated such that each ocean basin may have to be treated as a separate dynamic mantle model. However, the source of any plume is geochemically unique and distinct from the globally extensive source region that feeds the mid-ocean ridge system. In the area under discussion, Iceland and the Azores are two such hot-spots fed by mantle plumes. Much research has been undertaken in this area, particularly by the Deep Sea Drilling Project (DSDP) and the subsequent Ocean Drilling Program (ODP), and this has produced vast amounts of data on numerous aspects of the Atlantic Ocean. This chapter reviews a large part of this, concentrating on the nature of the mantle domain, and it is proposed that a mixing model (Morgan, 1971; Vogt, 1971; Schilling, 1973a, b) is applicable to these hot-spots and their surrounding regions. This differs from the previous models by the complexity of the geochemistry, due to differing 'mixtures' of the various components available in each hot-spot source region and other local heterogeneities that may exist in the MORB



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that the central bathymetric high present in each ridge segment is due to thermal uplift and a local increase in the magma supply rate from an underlying magma chamber. This could also be achieved by localizing the magma production area in the mantle (Figure 13.2b). Mantle upwelling regions vary in size, and may control the size of the crustal segment above. Vogt and Johnson (1973) did not observe any segmentation of this sort on the Reykjanes Ridge, although it does occur in the vicinity of the Azores plume at 37°N (Searle, 1979) (Figure 13.1a). Segmentation on Iceland occurs on the scale of volcanic zones (Figure 13.3a) and/or on a smaller scale, represented by the volcanic systems (Figure 13.4a).

In concordance with mantle segmentation controlling crustal segmentation as discussed above, the distribution of the plate boundary should be initiated by asthenospheric mantle processes. Similarly, Vogt and Johnson (1973) postulated that the TFZ could dam up the plume asthenopheric flow but, in light of new evidence (Lin *et al.*, 1990) suggesting that transforms are caused by flow patterns in the asthenosphere, this seems unlikely. Large transforms (e.g. the Charlie Gibbs Fracture Zone) may be inherited from old continental structures at the time of continental break up, whereas small offsets and discontinuities may be controlled by mantle processes.

# 13.4.2 Lithospheric thickness

The average seismic P-wave velocity of the asthenosphere is  $8.0-8.2 \text{ km s}^{-1}$  under normal ocean basins and extends down to approximately 250 km. Bott (1965) postulated that there is anomalously low velocity mantle (7.2-7.4 km s<sup>-1</sup>) under Iceland down to considerably greater depths, relative to normal mid-ocean ridges. Seismic tomography studies reveal that at 400 km depth (under the Azores and Iceland) this anomalous mantle of the plume heads is detected but below this the plume necks are not detected (Woodhouse and Dziewonski, 1984; Creager and Jordan, 1984, 1986; Grand, 1987). The normal asthenosphere may mix with that of the plume. Francis (1973) noted a decrease in the seismicity from 59.5°N on the Reykjanes Ridge towards Iceland, which was later interpreted in terms of the low velocity mantle from the plume extending this far south (Francis, 1975).

The lithosphere is the outer brittle layer that overlies the more plastic asthenosphere, and strictly includes all the solid material from the top of the crust down through the upper mantle to the top of the plastically deforming asthenosphere, and represents a solid rigid plate that may move independently. This boundary cannot be detected convincingly by refraction profiles alone, which only record changes in the seismic P-wave velocities, such that gravity surveys must be employed. At oceanic spreading centres the lithosphere thickens away from the axial zone, from 2 km at the ridge-axis (Searle, 1984) to 130 km thick at ages > 100 Ma (Parsons and Sclater, 1977).

In Iceland the lithosphere-asthenosphere boundary has been interpreted as the boundary between seismic layers 3 and 4 (Palmason, 1971). More recent work confirms the observation that low velocity values occur below the Icelandic crust (7.4 km s<sup>-1</sup> relative to 8.3 km s<sup>-1</sup> on the Reykjanes Ridge) (RRISP Working Group, 1980). It appears that the upper mantle constituent of the lithosphere beneath Iceland, that would be expected to have P-wave seismic velocities of 8.3 km s<sup>-1</sup>, is either mostly absent (Figure 13.6b), or there are some other peculiarities reducing the velocities to  $7.4 \,\mathrm{km \, s^{-1}}$ . In view of the cumulative evidence that suggests that this low velocity mantle extends down to depths greater than 250 km (Bott, 1965; Francis, 1969; Long and Mitchell, 1970; Woodhouse and Dziewonski, 1984; Creager and Jordan, 1984, 1986; Grand, 1987), it seems likely that the first interpretation is more accurate, as it satisfies the majority of the evidence. Underplating, which adds the cooling upper mantle to the base of the crustal lithosphere at ocean spreading centres, does not seem to occur under Iceland (Eysteinsson and Hermance, 1985).

Asthenospheric flow is frequently envisaged as radial from a plume (Morgan, 1971) and as lateral rolls from the MOR (Lin *et al.*, 1990). The radial flow of plume mantle may be represented by the distribution of the active volcanic and spreading zones and, in addition, physiographical and geochemical observations (Morgan, 1971; Vogt, 1971; Schilling, 1973a; Dewey and Burke, 1974; Einarsson *et al.*, 1977; Wyss, 1980), features which are exhibited by Iceland and the surrounding region.

Vogt (1971) and Schilling (1973a) suggested that variations in the topography along the Reykjanes Ridge and other regions near hot-spots were indicative of fluctuations in the rate of the asthenospheric mantle flow longitudinally underneath the ridge axis. Schilling (1973b) commented on the fact that the flow rate of the plume must be substantially greater than the flow beneath MORs, and suggests that this is why these hot-spot regions have anomalous crustal thicknesses. The degree of overflow varies from plume to plume, and temporally about a single plume. Vogt (1972) assessed the topography of the aseismic ridge between the Faeroes and Greenland and showed that between 50 and 60 Ma the discharge rate was very high and that it subsequently decreased to a minimum in the mid-Tertiary, before increasing again in the Late Tertiary.

Recent views of mantle processes, however, suggest that the upwelling plume mantle is 150–200°C hotter than that upwelling beneath the MOR, and an increase of 100°C in the melting region will produce double the volume of magma (White and McKenzie, 1989b). The consequence of this is that the variations in the topography produced by crustal thickness variation have to be a direct response of the mantle potential temperature, and not fluctuations in the discharge rate by the plume. This accentuates the importance of the morphology in interpreting mantle dynamics that are directly relevant to geochemical models.

(a)

#### 13.4.3 Existence of axial magma chambers

Considerable research has been directed towards discovering the existence and nature of magma chambers at ocean ridges (Iver, 1984; Browning, 1984; Orcutt et al., 1984; Detrick et al., 1987; Macdonald, 1989; Foulger and Toomey, 1989; Kent et al., 1990; and for a recent concise review, Langmuir, 1990), although it is not within the scope of this chapter to review such work in any detail. In summary, however, very small crustal magma chambers have been indentified at the fast spreading EPR ( $< 2 \, \text{km}$  depth), whereas none have been found at the MAR, although they have been imaged in Iceland (at about 3 km depth). If small chambers (< 200 m thick) are present in the Atlantic, then they would be extremely difficult to image seismically. Contrasting views exist over the number of chambers per crustal segment in the EPR (either one, as suggested by Ballard et al., 1981; Whitehead et al., 1984; Crane, 1985; White, 1989; or more than one, Thompson et al., 1989). and the existence of ponded magma at the base of the crust as well as or instead of a crustal chamber (Cann, 1970; White, 1989; Dick, 1989). The most recent efforts in the MAR (MARK area) have failed to image a crustal magma chamber (Detrick et al., 1990). In Iceland, magnetotelluric investigations have suggested there is a large ponded magma reservoir under the neovolcanic zones (15% partial melt in this zone of the mantle), that increases in size towards the centre of the island (Eysteinsson and Hermance, 1985). It seems then that the crustal thickness, spreading rate, and the requirement of sustained mantle temperatures may control the existence of a crustal chamber; the latter may be the most important.

The presence of a magma reservoir is significant to basalt petrogenesis because of the differentiation and mixing processes that may occur within it and which can mask the original source geochemical signature. However, compatible elements can be used to identify differentiation processes and the extent to which they have operated on the magmas. In contrast, the incompatible elements are relatively unaffected by these processes and, together with radiogenic isotopes, can be used to characterize the source and partial melting processes occurring in the mantle.

## 13.5 Petrographic series

On the basis of modal mineralogy, CIPW norms (Yoder and Tilley, 1962), total alkali contents (Macdonald and Katsura, 1964) and TiO, abundances (Chayes, 1965), basin oceanic basalts can be divided into different magmatic series. In the Atlantic and Iceland these are the tholeiitic, transitional alkali and alkali basalt series (Jakobsson, 1979a). Their characteristic geochemical features are summarized in Figure 13.8. The tholeiitic series is by far the most



Figure 13.8 (a) Alkali:silica diagram for Postglacial basalts from the Western Volcanic Zone and the Eastern Volcanic Zone discriminating between the different series (Jakobsson 1979a). (b) A: frequency distribution of postglacial rocks from the Western Volcanic Zone with respect to the normative content (cation %) of either olivine or quartz tholeiites. B: frequency distribution of postglacial rocks from the Eastern Volcanic Zone (alkali and tholeiite series) and the Western

Volcanic Zone (tholeiite series) with respect to the MgO content (Jakobsson 1979b).

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abundant in the ocean basins, followed by the transitional alkali and the alkali series, with basaltic rocks being the most voluminous.

Although magmatic series contain more evolved members than the dominant basalt, there is a distinct absence of acidic rocks in the North Atlantic. Oceanic andesites have been reported from aseismic ridges (Faeroes-Greenland), and acid and intermediate rocks form 9% of the erupted product on Iceland associated with the central volcanoes (Saemundsson, 1979). The bulk of the evolved rocks in Iceland are produced in the transitional and alkalic series. On the Reykjanes Ridge, evolved basalts and andesites occur in DSDP hole 409 in 2 Ma crust (Wood *et al.*, 1979a). There is controversy over the origin of these more evolved rocks, especially the acidic members (Moorbath and Walker, 1965; O'Nions and Gronvold, 1973; Sigvaldason, 1974; O'Hara, 1975; Sigurdsson, 1977; Macdonald *et al.*, 1987; Storey *et al.*, 1989; Thompson *et al.*, 1989), although assimilation fractional crystallization (AFC) or fractional crystallization are the two main processes invoked

Table 13.1	Characteristic	petrography	within	the	tholeiitic	series.	Compiled	from	various
authors (see	e text)								

Basalt groups	Mineralogical assemblages"	Characteristic petrography	Volcanic form on Iceland	Comment
Picrite	Abundant Mg-rich ol phenocrysts in a Mg-rich ol + plag + cpx groundmass	Lack of plagioclase phenocrysts. Olivines may often be in glomerophyric clusters	Small tava shields	There is often some dispute over the origin of these magmas, and many believe that they are of a cumulate nature
Olivine tholeiite	Plag + Mg-rich ol phenocrysts, microphenocrysts of plag needles in a plag + ol + cpx groundmass	Olivines and plagioclase often form glomerophyric clusters	Large to small lava shields, but are also found erupted from fissure swarms and central volcanoes	In the hand specimen these are very distinctive, and are the lava type of the large shield volcanoes in Iceland
Tholeiite	Plag phenocrysts ± cpx. Mg-rich ol rarely present > 3° Ground- mass of plag + cpx	Often more than one generation of feldspar phenocryst	Dominantly on the fissure swarms, but also erupted from the central volcanoes	The mineralogy and texture may be extremely variable
Quartz tholeiite	Plag phenocrysts ± e cpx. Mg-rich ol never present. Groundmass of plag + cpx	Phenocrysts very abundant and up to three generations may be present	Central volcanoes and mature fissure swarms	These are probably over-represented in Iceland

<sup>a</sup>(ol) olivine; (plag) plagioclase: (cpx) clinopyroxene

(Chapter 5). In general, the evolved compositions are more characteristic of the Atlantic oceanic islands (e.g. the Azores, Canaries, Tristan da Cunha) than the oceanic ridge (Chapter 9).

### 13.5.1 Tholeiitic basalt series

The dominant basalts erupted in the North Atlantic and Iceland fall into two petrological groups (Table 13.1 and Figure 13.8)—the olivine tholeiites and the tholeiites—both of which are found intercalated in a core section from the FAMOUS area (37°N, MAR, DSDP Leg 37) (Bougault and Hekinian, 1974; Blanchard *et al.*, 1976), and have also been reported from 45°N (Aumento, 1967), 53°N (Hekinian and Aumento, 1973) and 25°N in Cretaceous crust (Rice *et al.*, 1980).

In Iceland, these two groups are identified as the dominant eruptive products. The olivine tholeiites are erupted mainly in the off-axis shield volcanoes, but are also observed in the axial region, whereas the tholeiites are confined to the latter within the central volcanoes and fissure swarms (Figure 13.5). Within the present day rift zone, the olivine tholeiites are confined to the WVZ and the NVZ, and may become more abundant towards the centre of Iceland in the Langjokull region (Sigurdsson *et al.*, 1978; Schilling *et al.*, 1978; Meyer *et al.*, 1985). Within a single volcanic system, Jakobsson *et al.* (1978) suggested a temporal cyclic relationship between the two magma types. The large off-axis shield volcanoes are much rarer in the Tertiary lava pile, and it has been suggested that these large voluminous eruptions are due to the decompression associated with isostatic rebound towards the end of and after deglaciation.

Which of these two petrological groups is the dominant erupted product in both Iceland and the MAR is unclear. On Iceland, the occurrence of the olivine tholeiites in the axial region, as well as off-axis, suggests that they could be dominant overall in late glacial and post-glacial times. There are contrasting reports of the dominant basalt type in the MAR, that could reflect a sampling bias, and the lack of refined correlation between bathymetry, location and relative ages of the samples. If such shield volcanoes occur in the Atlantic as off-axis volcanoes, then the same may apply. Off-axis volcanism has been reported by Leg 49, on the Reykjanes Ridge (Luyendyk *et al.*, 1979), but this may be too close to Iceland to be typical. Another occurrence is at  $37^{\circ}N$  within the median valley, at comparable distances observed in Iceland of 15 km (Figure 13.1) (Needham and Francheteau, 1974), but this region may also be atypical.

Many workers consider these two types of basalt to be derived from each other by simple fractional crystallization processes, but there is frequently a chemical hiatus between them (Bougault and Hekinian, 1974; Blanchard *et al.*, 1976; Jakobsson *et al.*, 1978), suggesting that they are not from the same batch of magma and have different fractional crystallization histories.

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The parental magmas for the tholeiites are very rarely, if at all, sampled. The geochemical and field evidence on Iceland suggests a tectonic control for the distribution differences in these two magmas, with the olivine tholeiites being derived directly from the mantle, with very little modification en route. Either way these two tholeiitic types are petrographically representative of typical mid-ocean ridge basalts.

# 13.5.2 Alkali basalt series

The mineralogy and petrology of the alkali and transitional alkali series are much more varied. The series characteristically contain undersaturated feldspathoid minerals and can also be recognized by their alkali pyroxene compositions. Nevertheless, the North Atlantic basalts rarely possess feldspathoid minerals with the normative nepheline 'hidden' in the pyroxene. Bass (1971) and Ridley *et al.* (1973) pointed out that slow spreading ridges have a greater abundance of alkali rocks. In the Atlantic Ocean they have been reported from: (1) Reykjanes Ridge, in the same drill hole as tholeiitic lavas (Wood *et al.*, 1979a): (2) 45<sup>2</sup>N (Aumento, 1968; Wood *et al.*, 1979b; Tarney *et al.*, 1979); (3) the FAMOUS region (37°N) (Aumento, 1967) and (4) from various oceanic islands, such as the Azores (40°N) (Assuncao *et al.*, 1989) and the Cape Verdes (14.6–17°N) (Davies *et al.*, 1989) etc. In Iceland, the lateral zones (EVZ, SVZ, OVZ and the extinct SKVZ) are the only places that basalts with this alkali affinity have been erupted (Figure 13.3).

# 13.5.3 Relationship and origin of the different series

There has been considerable dispute over the origin of the alkali basalt series. To help resolve this problem the spatial and temporal relationships between the two series may provide some additional constraints. Iceland provides a perfect opportunity to examine such relationships. There are 29 volcanic systems currently active in Iceland; only four of these are alkaline, seven are transitional and the remainder are tholeiitic (Figure 13.8) (Jakobsson, 1979a). In view of the current interpretation of the plate boundary configuration (section 13.3.1; Saemundsson, 1974a; Einarsson, 1990), these alkali and transitional alkali rocks occur in the propagating tip of the EVZ, and in a zone that may represent either a 'leaky' transform zone or a diminishing volcanic zone (SVZ), (Sigurdsson, 1970). Alternatively, they can collectively be seen as an expression of the radial flow from the plume. The alkali volcanic systems are much less productive than the tholeiitic systems.

There is strong evidence to support mixing of the tholeiitic and alkalic series at crustal levels (McGarvie, 1984; Blake, 1984). Recently, McGarvie *et al.* (1990) have shown that the tholeiitic magmas of a volcanic system in the north of the EVZ travel up to 100 km to mix with the transitional alkali series (rhyolites and basalts) in the south and that this process of mixing commenced and became more frequent in post-glacial times (supporting the idea of a propagating rift). The alkali volcanic systems have noticeably fewer primitive equivalents to the tholeiitic picrites outcropping in the post-glacial eruptives of the WVZ and lack the off-axis shield volcanoes, as well as displaying a very high percentage of acid rocks (Jakobsson, 1979a).

The tholeiitic volcanic systems in Iceland display well pronounced tectonic fabrics akin to those in the Atlantic, but these are reduced in the alkali systems. It should be pointed out that the magmas erupting above the area of the hot-spot, at Vatnajokull, show no alkali characteristics. There is no consistent correlation between the erupted volumes of these two magma types in a single eruption, as has previously been suggested (Sigvaldason, 1974), but it varies regionally and temporally within and between single volcanic systems (Jakobsson, 1979a; Sigurdsson *et al.*, 1978).

Within the Tertiary volcanics (16-3.1 Ma) currently exposed, at least 44 volcanic systems have been identified, all of which are tholeiitic (Walker, 1963). This may be due to the obvious potential sampling bias of the limited exposure of the Tertiary rocks in Iceland. There can be a gradual change in the type of series being erupted within a single volcanic zone with time, which may be either towards alkalinity or the reverse. In the Plio-Pleistocene and Upper Pleistocene (3.1-0.7 Ma), which was the time of onset of glaciation in Iceland, the EVZ and the SKVZ became active, and the SVZ began to erupt more alkali rocks. Both the SVZ (Sigurdsson, 1970) and the SKVZ (Sigurdsson *et al.*, 1978) demonstrate a trend towards alkalinity during approximately a 2–3 Ma period (Figure 13.7e and f). The southern part of the EVZ has always erupted alkali and transitional alkali rocks (Jakobsson, 1979a) since the Pleistocene.

Sigvaldason (1974) and O'Nions et al. (1973) have shown that both alkali and tholeiitic magmas can be derived from one homogeneous source, as indicated by the Sr isotope data from the alkaline Westerman Islands and tholeiites from the rest of the neovolcanic zone. This suggests that other processes are responsible for the petrographic differences. However, other workers (e.g. Hart et al., 1973) have argued that remelting a metasomaticallyenriched mantle could produce alkali basalts, in a similar manner to the process proposed for Hawaii (Frey and Roden, 1986). This theory would require virtually all the lower oceanic lithosphere to be metasomatized if all the alkali basalts from the Atlantic oceanic islands are produced in this way. This seems to be drawing on a process that is far more complicated than necessary. The Cape Verdes hot-spot presents a problem to this theory as it has been penetrating the same oceanic lithospheric section for 125 Ma (Courtney and White, 1986), being situated near the African pole of rotation. This theory can satisfy neither the geophysical evidence for these areas nor the evidence in Iceland. There is no upper mantle part of the lithosphere

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under the EVZ in Iceland (Figure 13.3), suggesting that metasomatic enrichment is not the cause of the eruption of alkali basalts in this area. An alternative model whereby the alkali basalts in Iceland are produced by melting the thicker lower crustal layers of the flank zones, which contain alkali amphibole (Oskarsson *et al.*, 1985; Steinthorsson *et al.*, 1985), can be discounted on isotopic evidence as the observed ratios would not have had time to develop in the crust following remelting (Meyer *et al.*, 1985).

If it is considered that the alkali magmas are derived from the same mantle source as the tholeiites by a melting mechanism alone, then the percentage of partial melt must be very small and the area undergoing melting must be much larger than that which produces a tholeiitic magma batch of the same volume (Gast, 1968). In this instance MOR tholeiites are believed to be derived by 20-30% partial melting whereas alkali magmas represent only 4-6% melting. Similarly, irrespective of the depth of generation of primitive MORB (Chapter 6), alkali basalts, if from a similar homogeneous source, must have originated from even greater depths than the tholeiites, or equilibrated with the mantle at higher pressures.

Can inferences concerning source homogeneity and degree or depth of melting of alkali and tholeiitic melts be related to the distribution relationships observed in Iceland? The EVZ shows a gradual increase in alkalinity towards the south, which corresponds to an observed increase in crustal thickness in this direction (Palmason, 1971) and increased distance from the melting column of the plume that must be rising beneath Iceland. The reason for the presence of the alkali rocks in this region could, therefore, be attributed to smaller degrees of partial melting, at greater depths in the mantle, reaching this region relative to the rest of the neovolcanic zone in Iceland. This is in accordance with a propagating tip penetrating an older and thicker crust.

The SVZ erupted tholeiitic magmas throughout the Tertiary until 2.5 Ma. There has been a gradual change in the amount and chemistry of the magma produced in this region towards a smaller volume of more alkali magma (Sigurdsson, 1970). Sigurdsson et al. (1978) demonstrated that the SKVZ also erupted magmas that became progressively more alkalic with time within this geological period. Palmason (1971) reports that the depth to the asthenosphere in this region increases northwards from 8 km at Langajokull to 16 km under the SKVZ. The reduced proximity of the melting regions and heat source (centre of the plume) with time in these volcanic zones would be compatible with a ridge jump initiated by the plume as the Mid-Atlantic plate boundary drifted westwards. This gradual reduction in the heat flow to the melting source regions would cause a decrease in the percentage of partial melt produced, and the thickening crust clogging up the volcanic zone would possibly lead to extinction, as in the case of the SKVZ. The SVZ, although further away, is orientated in an east-west direction and could represent a weak radial flow line of the plume (Einarsson et al., 1927). The OVZ may be a similar flow direction, running sub-parallel to the EVZ 335

(Figure 13.3). The most recent geophysical evidence from the Atlantic (Lin *et al.*, 1990) may support the idea of mantle flow being the cause of ridge segmentation, that is, the cause of the distribution of the volcanic activity. This would suggest that the different series in the EVZ and the SVZ may be derived from the same source region, although geochemical evidence indicates that this may not be the case for the SVZ (see later).

### 13.5.4 Clinopyroxene-phyric basalts

Clinopyroxene phenocrysts are anomalously abundant in basalts from the vicinity of  $34-45^{\circ}$ N and  $60-68^{\circ}$ N, peaking on the Azores and Iceland. This has been related in various ways to the presence of the plumes in these areas (Cann, 1970; Moore and Schilling, 1973; Bougault and Hekinian, 1974). There are exceptions to this general observation: for example, at  $45^{\circ}$ N on the MAR, which has many of the geochemical characteristics of hot-spots but no clinopyroxene anomaly (Schilling *et al.*, 1983). The relative abundance of clinopyroxene between hot-spot areas varies, as in the lavas from the Azores relative to those from Iceland. Schilling *et al.* (1983) interpret this in terms of these regions having distinct fractional crystallization histories. The pyroxene-bearing basalts from the FAMOUS area are more evolved than the tholeiitic suites from the same dredge haul, suggesting they formed by late fractionation in a magma chamber. However, Bougault and Hekinian (1974) postulated that 'pyroxene basalts' are derived from a different parental magma than the two usual tholeiitic groups.

A smaller percentage of partial melting of a uniform source would be expected to produce abundant clinopyroxene in the magmas. This is the case in the highly undersaturated alkali basalts, but the alkali olivine basalts and the olivine tholeiites are equally saturated with respect to clinopyroxene, but rarely exhibit it as a phenocrystic phase, that is, Mg-rich magmas usually do not possess clinopyroxene phenocrysts, whereas Mg-poor evolved magmas may. The occurrence of clinopyroxene as a phenocryst phase is rare in typical MORB, but it can occur (Rice *et al.*, 1980) and exhibits two forms, the equant type in evolved rocks with low MgO concentrations and the anhedral ('resorbed') type that has a disputed origin.

What is important to establish is the order of crystallization and not just the presence of clinopyroxene. It is assumed that the equant type phenocrysts have formed in the usual way by fractionation of olivine + plagioclase + clinopyroxene, possibly in a magma reservoir. The resorbed type phenocrysts are thought to crystallize before most of the olivine and all the plagioclase, and to be derived from 'somewhere else'. It is the latter that is a matter of dispute; the clinopyroxene could be derived by mixing of a new, relatively unevolved, magma batch with phenocrysts of a previously more evolved batch. This undoubtedly occurs in MORB genesis, as is displayed by

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plagioclase phenocryst generations (Blanchard et al., 1976), but analysis of the compositions of some of the clinopyroxenes involved reveals that they have Cr- and Mg-rich undifferentiated compositions. Experimental evidence suggests that with an increase in pressure, the order of fractional crystallization will alter to clinopyroxene + olivine. The relict phenocrysts could have crystallized on the walls of dykes at depths of approximately 20 km before being carried up to the surface as xenocrysts. This raises the question of whether the so-called 'clinopyroxene anomaly' towards Iceland is due to increased crustal thickness, enabling differentiation to occur either in the magma chamber to produce the equant type phenocrysts (Luyendyk, Cann et al., 1979), and/or at deeper levels in the crust for resorbed type phenocrysts. With the information on the crustal structure that is already known, answers to the above questions can be derived. In the off-axis plumes such as the Azores and the Cape Verdes, the presence of the sub-oceanic lithosphere could provide a place for the high pressure xenocrystic clinopyroxene to crystallize.

13.6. Geochemical variation

# 13.6.1 Basaltic chemical types and the plume model

Morgan (1971, 1972) and Vogt (1971) introduced the idea of hot convective mantle plumes rising beneath oceanic islands (Chapter 9). Initially MORB tholeiites were thought to be remarkably uniform and, against this reference point, a progressive increase in light rare earth element (REE) enrichment, large ion lithophile elements (LIL), halogens and the radiogenic isotopes Pb and Sr, coupled with a decrease in the Nd isotopes, was observed towards the plume regions (in particular Schilling, 1973a, b, 1975; but also by Tatsumoto et al., 1965; Gass, 1970; Peterman and Hedge, 1971; Hart et al., 1973; Sun et al., 1975; White 1976; White and Schilling, 1978; Schilling et al., 1983). It was proposed that this chemical gradient was due to the presence of two distinct source regions, a depleted low velocity zone (DLVZ) source region that is globally present under the ocean ridge systems, and a primordial hot mantle plume (PHMP), which rises up from deep in the mantle. The two sources mix in varying proportions beneath the ridge axis, but progressively less mixing takes place with increasing distance from any one plume. The plumes in the Atlantic were thought to define mantle domains and attempts were made to map these using the major element composition of glasses (Melson and O'Hearn, 1979; Dmitriev et al., 1979; Sigurdsson, 1981). incompatible elements and isotopes (Schilling et al., 1983). The latter study confirmed Sigurdsson's (1981) observations, but noted that although the domain boundaries are not sharp, differentially enriched plumes and 'normal'

segments along the MAR between the plumes could be distinguished. Fracture zones may act as natural barriers to plume activity, such as the Hayes Fracture Zone which cuts off the Azores plume domain (Bougault and Cande, 1985; White and Schilling, 1978; Bougault and Treuil, 1980).

A chemical terminology for basalts was established based on what was essentially the Atlantic model for slow spreading mid-ocean ridges. The latter can be described in terms of the degree of elevation of the median valley relative to the norm, which shows an approximate positive correlation with the shallowness of the ridge in the Atlantic. Elevated (E-), transitional (T-) and normal (N-) ridge segments are a physiographical terminology generally applied to the MAR (and other mid-ocean ridges). It was then discovered that a good correlation exists between the degree of elevation and shallowness of the segments on the one hand, and the chemistry of the basalts erupted in the vicinity of a plume on the other (Schilling, 1973a, b). 'Elevated' was loosely redefined as a genetic term as 'plume' (P-) (see Schilling, 1975, for detailed definitions; subsequently used by Schilling *et al.*, 1983; Ito *et al.*, 1987).

The relative enrichment of the MAR basalts was estimated in terms of the degree of enrichment of the light REE. This can be demonstrated by the La/Sm concentration ratio, or related functions, such as the ratio of the enrichment factor relative to a chondritic composition, that is, [La/Sm]<sub>FF</sub> (Schilling, 1973a). Unless otherwise stated in this chapter, the relative degree of elemental enrichment and depletion is related to chondritic values. Normal ridge segment basalts (N-MORB) were defined as possessing an [La/Sm]<sub>FF</sub> of less than one, transitional ridge segment basalts (T-MORB) of about 1. and plume or elevated ridge segments (P-MORB or E-MORB) greater than one. The terminology was applied to other ocean basins regardless of the spreading rate and linked composition with ridge physiography. Later Schilling et al. (1983), redefined N-MORB as [La/Sm]<sub>EF</sub> < 0.7, the transitional segments as having a sharp gradient of enrichment in the [La/Sm]<sub>EF</sub>, and the 'plume' elevated segment displaying a maximum in [La/Sm]<sub>FF</sub> with arbitrary cut-off values for individual plumes. The terminology is weakened. however, by the frequent fluctuation of the degree of enrichment of basalts from within a single transitional or enriched segment drill hole (Wood et al., 1976b; Tarney et al., 1979). Also, Le Roex et al. (1985b) observed transitional enriched basalts in the very slow spreading segments of the American-Antarctic ridge far from the Bouvet plume. This emphasizes the pitfalls of applying genetic significance to the terminology of the basalts.

As the preceding section has outlined, much of the variation within E-MORB and T-MORB areas of the MAR was originally accounted for by interaction between enriched plume sources and depleted MORB sources. However, when data from the Atlantic increased sufficiently in the 1970s to highlight the variability of enrichment (DSDP Legs 37 and 49), there was a fashion of opposition to the mantle plume hypothesis. Various alternatives were proposed including: (1) a veined source (Dick, 1977; Hanson, 1977; Wood,

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	Basalt type, locality and (sample number)									
	N-MORB	T-MORB MAR 63"N Rcykj. Ridge (49-407-45-3) <sup>6</sup>	E-MORB MAR 63 N Reykj. Ridge (49-409-24-2) <sup>6</sup>	E-MORB MAR 63°N Reykj. Ridge (49-407-47-1) <sup>b</sup>	Icelandic bas		alts			
Analysis	MAR 37°N S. HFZ (82-556-2-1)"				Tholeiite (jak 97a-2) <sup>e</sup>	Transitional atkali basalt (jak 97a-6)	Hawaiite (jak 97a-10) <sup>r</sup>	Average N-MORB <sup>4</sup>	Average E-MORB <sup>d</sup>	
Major oxides	; (wt%):									
SiO <sub>2</sub>	50.56	48.69	49.86		49.3	47.08	48.46			
TiO <sub>2</sub>	1.04	2.15	1.78		2.28	4.63	3.00		_	
Al <sub>2</sub> Ō <sub>3</sub>	15.96	15.23	14.32		13.51	12.71	16.40	_		
Fe <sub>2</sub> O <sub>3</sub> *	7.83	11.92	12.75		14.57	15.47	13.37			
MnO	0.15	0.13	0.18		0.24	0.22	0.26	_	_	
MgO	7.89	8.10	7.70		6.34	5.06	3.82	_	_	
CaO	12.74	10.60	11.39		10.71	9.91	7.71		_	
Na <sub>2</sub> O	2.44	2.65	2.44		2.46	3.08	5.11			
K <sub>2</sub> Ō	0.16	0.19	0.23	—	0.24	0.72	1.19			
$P_2O_5$	0.11	0.23	0.19		0.21	0.57	0.62	—		
race elemen	its (ppm):				``					
Ba	30	**	70	75	· _/	_	_	6.3	57	
Со		71	48	40	'	_				
Cr	326	313	317	122		_		290	46	
Hſ	1.8	3.69	3.1	3.2				2.05	2.03	
Ķ.			1494	3155		_	_	600	2100	
ŇЬ	1.6	_	11.5	17.7				2.33	8.3	

Table 13.2 Geochemical data for different MORB types from the Mid-Atlantic Ridge (MAR) relative to examples of Icelandic basalts. Average N-MORB and E-MORB from Sun and McDonough (1989)

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NG	140	146	106	48.4		—	<u> </u>	138	32
Dh	140		_	261				0.3	0.6
ГU РЬ	2	_	4.4	6.3	_		—	0.56	5.04
RU S.	-	45.6	453	17	_	_	_	44	39
50 Sr	112	45.0	103	173				90	155
31 To	0.09	0.90	0.68	0.87	_		_	0.132	0.47
1 a Th	0.07	0.50	0.59	0.67		_	_	0.12	0.60
10 T:	0.14	12769	10431	11750				7600	6000
11 V	267	383	377	357		_	_		
v v	207		41.8	35.9			_	28	22
- Zr	62	140	116	130		—	—	74	73
ire earth e	lements (ppm):								
La	1.84	8.4	6.1	8.4			·	2.5	6.3
Ce	6.45	24	18.6	23.1				7.5	15
Nd	6.88	17	13.9	16.1				7.3 .	9.0
Sm	2.59	5.2	4.71	5.02				2.63	2.60
Eu	0.98	1.9	—			—		1.02	0.91
Gd	3.66							3.68	2.97
Tb	0.71	0.91	0.97	0.93				0.67	0.53
ſm	0.43	0.6					••	0.456	0.356
Yb	2.77	3.22	3.94	3.21				3.05	2.37
1.0	0.45	0.52	_			—	_	0.455	0.354

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<sup>a</sup>Weaver *et al.* (1982) <sup>b</sup>DSDP Leg 49 <sup>c</sup>Jakobssen (1979a) <sup>d</sup>Sun and McDonough (1989)

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1979; Wood et al., 1979a, b; Tarney et al., 1979, 1980; Dick et al., 1984; Le Roex et al., 1983); (2) a streaky source (Zindler et al., 1982, 1984; Fitton and James, 1986); (3) a 'marble cake' mantle (Allegre and Turcotte, 1986), and (4) fluid or gaseous phases inducing metasomatism (Green, 1972; Frey and Green, 1974; Lloyd and Bailey, 1975; Frey et al., 1978; Schilling et al., 1980; Menzies, 1983; Schilling et al., 1983; Dick et al., 1984; Stolz and Davis, 1988).

The plume model (e.g. Schilling *et al.*, 1983), however, apparently satisfies the following features: (1) thermally-related geoid bulge (Morgan, 1971); (2) geophysical evidence suggesting that melting occurs down to 250–400 km depth beneath Iceland (Bott, 1965, 1988; Tryggvason, 1964; Francis, 1969; Long and Mitchel, 1970: Hermance and Grillot, 1970; Woodhouse and Dziewonski, 1984; (3) crustal thickness increase (Palmason, 1971); (4) flow fabrics and/or temperature variation (Vogt, 1971; McKenzie, 1984); (5) most of the isotopic geochemical variation, and (6) the temporal persistency of two source regions (for 60 Ma in Iceland). The idea of MOR magmatism existing as a passive response to extensional spreading, and large plumes (like Iceland) having a different origin of a more forceful nature were being widely, but theoretically, discussed in the early 1970s (Morgan, 1971; Vogt, 1971; Schilling, 1973a, b; Schilling and Noe-Nygaard, 1974). These ideas formed the basis to similar and currently accepted views of plume activity (e.g. White and McKenzie, 1989a, b).

# 13.6.2 Normal ridge segment: N-MORB tholeiites

It is useful to use N-type MORB as a reference frame, as this is globally the most abundant basalt type within oceans. The chemical characteristics of the different MORB types are listed in Table 13.2.

N-MORB typically displays a depleted nature, relative to chondritic values, as demonstrated by the low abundances of incompatible trace elements relative to the compatible elements. The [La/Sm]<sub>EF</sub> is low (<1). The radiogenic isotope ratios, <sup>87</sup>Sr/<sup>86</sup>Sr (0.70234–0.70245) and <sup>206</sup>Pb/<sup>204</sup>Pb (18.53–18.74) are also depleted, whereas <sup>143</sup>Nd/<sup>144</sup>Nd ratios are enriched (0.51318–0.51322) (Park and Staudgel, 1990). As demonstrated globally there is usually a good negative correlation between <sup>143</sup>Nd/<sup>144</sup>Nd and the <sup>87</sup>Sr/<sup>86</sup>Sr ratios for MAR basalts, for example, from 11°N and 6°S (Hart, 1976; O'Nions *et al.*, 1977). By definition, N-MORB should only show minor deviations in composition spatially or temporally, as is generally observed, suggesting that the source region for these basalts is remarkably uniform on the scale sampled by the erupted magmas.

The MARK area in the Atlantic (23°N) is one of the most thoroughly studied areas in terms of overall geology (Purdey *et al.*, 1979; Detrick *et al.*, 1984 for the bathymetry; Mayer *et al.*, 1985 for a general account). From 22 to 24°N (Kane Fracture Zone) the ridge is divided into three segments separated by

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non-transform discontinuities. ODP Leg 106/109 sunk holes in two of these segments on zero-age crust (Detrick *et al.*, 1990; Donato *et al.*, 1990) and found only very minor variations in the degree of depletion between the three sites. DSDP Legs 51, 52, and 53 sampled Cretaceous crust (108 Ma) at 25°N, on the same flow line as Leg 106/109, and retrieved N-MORB, suggesting that the MORB source has remained uniform for this length of time to the present. However, Cretaceous crust from the Caribbean displays some anomalous features and has been compared to thickened segments of oceanic crust such as oceanic plateaux (Floyd, 1989).

# 13.6.3 Transitional ridge segments: T-MORB tholeiites

These basalts are enriched in LIL elements,  ${}^{87}$ Sr/ ${}^{86}$ Sr and  ${}^{206}$ Pb/ ${}^{204}$ Pb ratios and depleted in  ${}^{143}$ Nd/ ${}^{144}$ Nd ratios relative to N-MORB, and the change should occur with a steep gradient such as seen along the Reykjanes Ridge (Figure 13.9). Fluctuations and variations in the same location from light REE



Figure 13.9 Chemical gradients observed along the Reykjanes Ridge, which formed the basis for Schilling's model. Ridge basalts were dredged by R.V. Trident during cruises TR41 and TR101 in 1967 and 1971. Note the broken scales south of 60°N. La and Sm enrichment factors relative to chondrite meteorite concentrations from Sun et al. (1979).

depleted to light REE enriched seems to be more typical of basalts in the transitional regions. Joron *et al.* (1980) demonstrate that there are comparable La/Ta ratios for the Reykjanes Ridge, 37°N and 45°N and classifies them all as T-MORB accordingly, although Schilling *et al.* (1983), considered 45°N as an E-MORB segment.

13.6.3.1 Reykjanes Ridge (MAR,  $61^{\circ}N$  to  $63^{\circ}N$ ). One of the two best known examples of T-MORB occurrences is the Reykjanes Ridge (Schilling 1973a,b; Hart *et al.*, 1973; Hart and Schilling, 1973; Sun *et al.*, 1975) (Figure 13.9). This area shows a gradual change from N-MORB, south of  $61^{\circ}N$ , to E-MORB at  $63^{\circ}N$ , 400 km south of the Iceland coast (Figure 13.9). Segmentation occurs on the Reykjanes Ridge only in the form of *en echelon* fissure swarms, very similar to those seen on Iceland. The median valley becomes less pronounced and shallower towards Iceland.

The major element compositions of the tholeiites show only slight variations (Sigurdsson, 1981), whereas the incompatible and the highly incompatible elements become progressively more enriched towards Iceland (Schilling, 1973b). Chondrite-normalized REE patterns change from light REE depleted, flat, and then slightly light REE enriched towards and on Iceland. This gradient occurs over 400 km of the Reykjanes Ridge and correlates well with other elements. Sr isotope work carried out on the same batch of samples analysed by Schilling (1973b) shows that the truly transitional area is nearer 200 km long, and that the gradient is more of a scatter (Figure 13.9) (Hart *et al.*, 1973). Northwards from 200 km south of Iceland (62.5°N), the  ${}^{87}$ Sr/ ${}^{86}$ Sr ratios are relatively uniform and high (0.70304), and from 400 km south of Iceland (61°N) the lower uniform ratios averaged at 0.70273 are equivalent to N-MORB (Hart, 1971). There may also be a positive correlation between  ${}^{87}$ Sr/ ${}^{86}$ Sr ratios and depth (Hart *et al.*, 1973; Schilling and Noe-Nygaard, 1974; Flower *et al.*, 1975).

Lead isotope data on quartz and olivine normative tholeiites from the same area provide supporting evidence for this gradient and show a closer correlation with the LIL elements (Figure 13.9) (Sun *et al.*, 1975). The <sup>206</sup>Pb/<sup>204</sup>Pb data give high ratios over Iceland (18.70) and low ratios south of 61°N (18.30) to the Charlie Gibbs Fracture Zone (53°N), with a progressive decrease in the transitional zone on the Reykjanes Ridge. The gradient mimics the [La/Sm]<sub>EF</sub> smooth convex upwards curve, but not the <sup>87</sup>Sr/<sup>86</sup>Sr steep gradient. The data of O'Nions and Pankhurst (1974) for Sr and Pb from their own samples were concordant with the smooth progression reported by Schilling (1973b).

The  $^{143}Nd/^{144}Nd$  isotope ratios (O'Nions *et al.*, 1977) in basalts north of 63°N are indistinguishable from Icelandic samples. The usual negative correlation between  $^{143}Nd/^{144}Nd$  and  $^{87}Sr/^{86}Sr$  ratios (the mantle array) breaks down (O'Nions *et al.*, 1977), and so is not consistently present in transitional segments. The consequence of this observation on the

terminology applied by Schilling (1975) is that T-MORB starts further north with regard to the Sr data relative to the Nd and Pb data.

DSDP Leg 49 (Wood et al., 1979b; Tarney et al., 1979) tested the temporal relationship of the Reykjanes Ridge enrichment by drilling three holes (407, 408 and 409) perpendicular to the axis (Figure 13.10). Chondrite-normalized REE patterns are highly variable from light REE depleted to light REE



Figure 13.10 A map to demonstrate the relative locations of the DSDP and ODP drill holes (collected from DSDP and ODP volumes mentioned in the text), and to summarize the degree of depletion found in the basalts recovered from each drill hole, based on [La/Sm]<sub>EF</sub> or similar ratios.

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in the degree of enrichment (Wood *et al.*, 1979b) (Figure 13.10). The consistently high trace element abundances are interpreted as a result of either high degrees of fractional crystallization or low degrees of partial melting (Tarney *et al.*, 1979), the latter being compatible with the presence of alkali basalts. The 45°N region can be distinguished from the Reykjanes Ridge on the incompatible element characteristics, but to a lesser extent from the FAMOUS area. This region of the MAR has high <sup>87</sup>Sr/<sup>86</sup>Sr ratios (White and Schilling, 1978), and the Pb isotopic data from Hole 410 (Mattinson, 1979) exhibit very high <sup>206</sup>Pb/<sup>204</sup>Pb ratios that are within the range of Pb ratios from oceanic islands in the South Atlantic.

There has been some dispute about the relationship of this segment of ridge relative to the Azores hot-spot, as it does not display any of the morphological features suggestive of a hot plume influence, although much of the geochemistry suggests that it is E-MORB. There is evidence for a plate boundary relocation in this region based on the structure of the area and nearby seamounts, which may have masked the original features or prevented their development (Searle and Laughton, 1977; Searle and Whitmarsh, 1978).

13.6.4.2 Iceland. Icelandic basalts should by definition (Schilling, 1975) demonstrate a consistent enrichment in LIL elements relative to MAR segments. The mainland Iceland olivine tholeiites and tholeiites are characterized by high Fe and Ti, low K/Rb ratios and LIL element abundances, and light REE enrichment (Shimokawa and Masuda, 1972; Brooks *et al.*, 1974). There is some scatter in the concentrations of incompatible elements such as K, Ti, Rb and Zr over Iceland relative to the MAR (Sigvaldason. 1974), such that the enrichments are seen more as a general increase in concentrations. The absolute variation in incompatible elements is probably representative of the effects of fractional crystallization.

Earlier research (Shimokawa and Masuda, 1972; Schilling, 1973a; O'Nions and Gronvold, 1973; Hart *et al.*, 1973) found that a uniformity existed in the degree of light REE enrichment and isotopic ratios which supported the classification of E-MORB proper. However, as the data set increased from both the neovolcanic zone and older rocks on Iceland, the true variability of this enrichment came to light (Figure 13.11) (Sun and Jahn, 1975; O'Nions *et al.*, 1976). Tholeiitic basalts with N-MORB characters have been sampled from the NVZ post-glacial basalts and both LIL depleted and LIL enriched characteristics have been reported from within the products of a single fissure swarm (Sigvaldason *et al.*, 1976). The same is true for  $^{87}$ Sr/ $^{86}$ Sr ratios, which were thought to have a uniform average of 0.70315 for glacial and post-glacial basalts (O'Nions *et al.*, 1973) (except for the alkali basalts of the SVZ).

Subsequent studies revealed a range from 0.70291 to 0.70341 (O'Nions *et al.*, 1976), excluding the SVZ, which is similar to the range observed on the Reykjanes Ridge. In general there is a positive correlation between the  ${}^{87}$ Sr/ ${}^{86}$ Sr ratios and the degree of LIL element enrichment in these basalts

(Figure 13.9b). In the Tertiary, the LIL element enrichment and the  ${}^{87}Sr/{}^{86}Sr$  ratios are distinctly higher and a progressive decrease in the latter from 0.7036 (Tertiary, 15 Ma) to 0.70315 (Pleistocene, 2 Ma) is observed (O'Nions and Pankhurst, 1973). The degree of decrease in the gradient increases sharply at 4 Ma (Figure 13.12a). The  ${}^{143}Nd/{}^{144}Nd$  isotope ratios for Tertiary basalts are significantly lower than those for the post-glacial basalts (Figure 13.12b) (O'Nions *et al.*, 1977), and therefore the correlation with the Sr isotope ratios exists as it does for the post-glacial basalts. However, the Tertiary basalts are less radiogenic in Pb (Sun and Jahn, 1975; Welke *et al.*, 1968). The usual correlation between the Sr, Nd and Pb systems is not apparent in the Tertiary samples, with the Pb data deviating from the usual trend.

The geochemistry shows a sharp change either side of the Tjornes Fracture Zone (Jakobsson 1979b; Schilling et al., 1983). O'Nions and Pankhurst (1974)



Figure 13.12 (a) Sr ratios versus age. Circles are basic rocks, squares are acid rocks (O'Nions and Pankhurst 1973). (b) Sr and Nd isotopic ratios, showing that the transitional alkali basalts of the EVZ, the Tertiary tholeiitic basalts and the Snæfellsness alkali basalts have the higher Sr and lower Nd ratios. From Meyer et al. (1985).

report low <sup>87</sup>Sr/<sup>86</sup>Sr ratios similar to N-MORB (based on one sample from the Kolbeinsey ridge) and relative to the variation just described within the neovolcanic zone on Iceland. However, to suggest that the geochemical anomaly does not exist on this ridge segment is a little premature. A similar discontinuity was reported by Sigurdsson *et al.*, (1978) between the Skagi and the Langjokull volcanic zones (65°10'N) on the basis of only La/Sm ratios. It may be that these are merely sampling discontinuities or, in the latter case, a direct response to differing degrees of partial melting.

# 13.6.5 Alkali basalts from Iceland and other Atlantic oceanic islands

The alkali basalts exposed on Iceland are all younger than 3.5 Ma. O'Nions *et al.* (1973) observed a higher degree of light REE enrichment and higher total abundances of REE in the alkali olivine basalts of the Westmann Islands relative to mainland Iceland. They interpret this in terms of smaller degrees of partial melting at the propagating tip of the EVZ compared to the tholeiites of the rest of the neovolcanic zone. The  ${}^{87}$ Sr/ ${}^{86}$ Sr ratios from the same region are similar to the rest of the tholeiites on Iceland (Hart *et al.*, 1973), suggesting that the  ${}^{87}$ Sr/ ${}^{86}$ Sr ratios are representative of the source, whereas light REE enrichment is not (Galer and O'Nions, 1986).

Further research suggests that Sr isotopic values and the REE enrichment are much more varied than supposed by O'Nions *et al.* (1976) for the rest of Iceland. For example, the  ${}^{87}$ Sr/ ${}^{86}$ Sr ratios (average 0.70341) and the degree of L1L element enrichment for the Snaefellsnes Peninsula alkali basalts is higher than anywhere else within the neovolcanic zones on Iceland, and are equivalent to those found on Jan Mayen (north of Iceland) (O'Nions and Pankhurst, 1974: O'Nions *et al.*, 1976). The degree of L1L element enrichment is higher than that in the Tertiary, whereas the  ${}^{87}$ Sr/ ${}^{86}$ Sr values are comparable, and it has been suggested that these alkali basalts have a similar source to the Tertiary lavas. Comparable values have been reported from Torfajokull (EVZ), where assimilation fractional crystallization is believed to have occurred (O'Nions and Gronvold, 1973), and this will affect the Sr and Nd ratios (Hermond *et al.*, 1988).

The <sup>1+3</sup>Nd/<sup>144</sup>Nd isotope ratios for two post-glacial alkali basalts from the Snaefellsnes Peninsula are comparable to the contemporary tholeiites (O'Nions *et al.*, 1977; O'Nions and Pankhurst, 1973). This indicates that the isotope composition of the source region in Iceland may have changed on a temporal scale and may even be spatially heterogeneous, if crustal contamination has not occurred. The lack of correlation between the Pb and the Sr and Nd isotope ratios was used to argue against contamination of the Tertiary lavas. However, Pb may behave as a mobile element during hydrothermal alteration (Dickin, 1981), which may explain the lack of

correlation and support crustal influences on the isotopic ratios of the Icelandic basalts.

Oceanic island basalts and Icelandic alkali basalts yield a much wider range and generally higher values of Sr and Nd isotope ratios than MORB (Chapter 9), and although the ratios within a single island group may show little variation (O'Nions *et al.*, 1977), large differences occur between different hotspots. St. Helena and Ascension have low  ${}^{87}$ Sr/ ${}^{86}$ Sr ratios, similar to that of N-MORB (Tatsumoto *et al.*, 1965). A  ${}^{207}$ Pb/ ${}^{204}$ Pb- ${}^{206}$ Pb/ ${}^{204}$ Pb plot demonstrates that the ratios for many oceanic islands near a MOR segment are extremely variable, but they all seem to show linear trends towards N-MORB values, suggesting that mixing is a regular phenomenon (Sun *et al.*, 1975). Almost every oceanic island source requires a unique combination of Rb, Sr, U and Pb in the source to account for the observed isotopic ratios and abundances.

The variation between the Azores and Iceland hot-spots is discussed by Schilling et al. (1983). These regions both possess elevated morphology, unusual tectonic characteristics and geochemical signatures. The radial extent to which basalts with N-MORB affinities are affected varies from plume to plume. The mantle beneath the islands is enriched in LIL elements, H<sub>2</sub>O and halogens. This enrichment is more pronounced in the Azores than Iceland, yet the elevation is greater in Iceland. Enrichment in Fe, Mg and Ca, and depletion in Na and SiO<sub>2</sub> for a given Mg-value, is more characteristic of Iceland. The isotopic evidence shows that the source region of the Azores (Chapter 9) has had a distinct evolution and different periods of isolation relative to Iceland (Sun and McDonough, 1989). The combined evidence indicates that there are three distinct mantle sources present in the North Atlantic, the DLVZ, the Iceland plume and the Azores plume, and that others exist to supply plume-dominated basaltic magmas to oceanic islands in the South Atlantic (Weaver et al., 1987; Chaffey et al., 1989; Davies et al., 1989; see Chapter 9).

# 13.7 Comparison of the North Atlantic and Iceland

Both regions display episodic activity, both tectonically and volcanically, within a dominantly extensional environment, with extensive fissures and normal faulting occurring even directly over the plume head in Iceland. The basic layered crustal structure is oceanic in nature and oceanic analogues can be called on to describe the present plate boundary configuration (Figure 13.3) (Fornari *et al.*, 1989). The *en echelon* arrangements of the volcanic systems in parts of Iceland have their counterparts on the Reykjanes Ridge. The EVZ can conveniently be interpreted in terms of a propagating tip of the volcanic zone. In the EPR, propagating rifts dominantly propagate away from plumes.

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The other zones (SVZ, SKVZ and OVZ) are difficult to find modern oceanic counterparts for, and thus collectively they are more likely candidates for the manifestations of radial flow of the mantle. The transform zones in normal oceanic basins rarely have classic textbook form. Ridge jumps occur in a normal oceanic environment and are not just a response to the plume, although they are more likely to occur in such an unstable environment. MOR magmatism is of a passive nature and occurs in response to extension (Schilling, 1973b; McKenzie *et al.*, 1990). Mantle plumes are the result of self-maintained mantle convection and not a passive response to plate motion and so Iceland may not be the ideal place to study such passive processes.

When using the term 'volcano' in Iceland, Jakobsson (1979b) rightly suggests that this should apply to a volcanic system as a whole. Each eruption in that system can be thought of as a monogenetic crater eruption. This notion draws the scales of volcanoes in the MAR (Brown and Karson, 1988), the EPR and Iceland closer together, and fits the idea of a large volcano in each segment having distinctive morphological, volcanic-tectonic character as well as a distinctive chemistry. Lateral magma flow occurs on Iceland within a system (and possibly between systems) for distances of 70 km (Bjornsson, 1977; Sigurdsson and Sparks, 1978; Saemundsson, 1978; Steithorsson, 1978). In addition, it is suggested that mixing of magmas occurs in a basal crustal magma reservoir (Eysteinsson and Hermance, 1985). Lateral magma movement within single oceanic segments (Lin *et al.*, 1990) probably also occurs in MAR segments in addition to magma mixing.

It seems that Icelandic rocks are more differentiated than those of the rest of the Atlantic, but, as indicated earlier, these processes can be identified and their effects subtracted from the overall geochemical characteristics, bringing the Atlantic and Iceland closer together on a comparative scale. The existence and abundance of the central volcanoes and the acid and intermediate rocks in Iceland may be attributed to thicker crust, which either enables the magmas to differentiate further in crustal chambers or to assimilate crustal material to produce acid rocks. Alternatively, these features may be a result of the sustained high geothermal gradient provided by the plume. There are calderas reported from the summits of volcanoes on normal segments of the MAR (Detrick *et al.*, 1990) comparable to calderas such as Krafla (NVZ). The alkali rocks are due to the different melting regime at the plume and are not typical of the oceanic environment, although they are also found in other areas ( $45^{\circ}N$ ) of the MAR.

Towards Iceland, along the Reykjanes Ridge, there is an observable gradation towards higher mantle temperatures that has produced thicker crust, and may account for some of the changes in the chemistry of the erupted lavas. It seems that the effects caused by the hot-spot can be identified, and as long as the plume associated processes are acknowledged, it is acceptable to use Iceland as a place to study processes at MORs.

# 13.8 Concluding statements

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- 1. The structure, composition and nature of convection in the mantle are absolutely crucial to the accuracy of any geochemical model. The sub-Atlantic mantle may be interpreted as having three scales of convection: (1) large-scale, whole mantle that emerges at the surface in plumes such as Iceland; (2) smaller scale that occurs passively at the mid-ocean ridge; and (3) minor internal 'mixing' that occurs in the mantle to account for the isotopic variability of Atlantic MORB. The small-scale convection occurring at the mid-ocean ridge does not completely obliterate radial asthenospheric flow from adjacent plumes or vice versa. The Atlantic Ocean provides a natural laboratory within which to observe the interaction between deep mantle plumes and the mid-ocean ridge tectonics.
- 2. Atlantic Ocean ridge basalts fall into three petrological series covering the full range of tholeiites to alkali basalts. The alkali basalts are predominantly produced by a smaller degree of partial melting from a similarly depleted source as the associated tholeiites, which is distinct from the more enriched plume source of Atlantic oceanic islands. The Mid-Atlantic Ridge and Icelandic tholeiites show various degrees of isotopic, LIL and light REE depletion and enrichment, and are often chemically characterized as normal (N-) MORB, transitional (T-) MORB and enriched (E-) MORB or plume (P-) MORB. Although the original classification related to the consistent correlation with the degree of elevation of the median valley, or the shallowness of the ridge, this is no longer the case. It is stressed that such genetic terms as 'plume' are misleading and the descriptive term 'enriched' is preferable in relating to different MAR segments. It is also important to appreciate that the three MORB types naturally grade into each other and, in view of the problems in defining the chemical limits to each type, it might be more appropriate to use only N-MORB (depleted features) and E-MORB (variably enriched features).
- 3. The degree of LIL and light REE enrichment in Icelandic basalts and the Reykjanes Ridge varies considerably, both spatially and temporally. The isotopic data are more consistent and show a positive gradient towards Iceland from the south, and some uniformity within the Icelandic volcanic zones. Temporal variation exists in Iceland from the Tertiary lavas to the present and must reflect a progressive change in the composition of the source region, or represent a degree of efficiency of mixing the depleted MORB source with the plume source. Anomalous isotopic ratios occur in the SVZ and Tertiary crustal assimilation has been suggested as a means of producing these features. The volcanic zones and segments in Iceland may represent finger-like expressions of radial flow that are preferentially

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channelled along lines of weakness. The correlation observed between the ridges and troughs on the Reykjanes Ridge with the high and low <sup>87</sup>Sr/<sup>86</sup>Sr ratios, respectively, agrees with the idea of the plume asthenospheric source producing higher volumes of magma and thus providing the higher isotopic ratios in the magma mixture. This suggests that mixing occurs between the two mantle sources below Iceland.

4. The presence of plumes and their interaction with the active ridge are a feature of the Atlantic Ocean crust. The nature of components identified in the plume source are varied and each hot-spot should be investigated separately. There is a general consensus that subducted oceanic lithosphere, stored and isolated in the mantle over long periods of time, is involved and has been identified as a contributor to the source regions of the Cape Verdes and Azores hot-spots. Melting of enriched sub-oceanic lithosphere may also be a contributing factor, although this source does not apply to Iceland. Differences between the Azores and Iceland hotspots, such as the anomaly in the residual gravity/residual elevation could be achieved by an increase in the temperature of the mantle beneath Iceland, and/or an increase in the percentage of partial melting present. The latter would be compatible with the presence of the different magma series of these two islands. The high LIL element enrichment in the Azores relative to Iceland supports the notion of sub-oceanic lithospheric enrichment under the Azores, and in a similar manner at 37°N and 45°N MAR.

# PART IV SOURCES

14 Stable and noble gas isotopes RICHARD EXLEY

## 14.1 Introduction

There are four main reasons why stable isotopes provide important information about the petrogenesis and source regions of oceanic basalts.

- Hydrogen, carbon. nitrogen, oxygen and sulphur are essential components of the volatiles which drive all magmatism. Direct mantle nodule samples provide sparse evidence on the host phases of these elements. The study of stable isotopes in oceanic basalts provides a window on these elements in the mantle.
- Large fractionations of stable isotope ratios exist between the crust, mantle, hydrosphere, atmosphere and the biosphere. Stable isotopes thus provide a tool for studying the dynamics of exchange between these reservoirs.
- By linking stable isotopes with, in particular, noble gases and radiogenic isotopes, the earth's degassing history may be studied.
- Very large stable isotopic effects are observed in meteorites. These provide models of unprocessed planetary material which indicate the possible isotopic heterogeneities in the deep earth. Again, the study of mantle stable isotopes via the vehicle of oceanic basalts may provide a window into deep earth history.

# 14.2 Stable isotopes

Natural variations in stable isotope ratios are usually caused by fractionation effects. These are either equilibrium fractionations, caused by the effect of different atomic masses on the energy levels of chemical bonds, or kinetic fractionations, caused by the effect of the different isotopic masses in rate-controlled processes such as diffusion. The theory of these effects was described in the classic paper by Urey (1947). Even for elements with large mass differences between their stable isotopes, the ratios vary only by a

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# EN ECHELON AXIAL VOLCANIC RIDGES AT THE REYKJANES RIDGE : A LIFE CYCLE OF VOLCANISM AND TECTONICS.

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### Abstract:

New deep-towed sidescan records and multibeam bathymetric data along the Revkianes Ridge allow us to assess the varying thermal effect of the Icelandic hotspot. The increasing predominance of faulting speculated to be derived from mantle upwelling and outflow over simple extensional neotectonics is traced from northeast to southwest. Numerous en echelon volcanic ridges built along the plate boundary are seen in various stages of growth and decay, and a "cycle" of alternating construction and dismemberment is proposed. Unfaulted, short, narrow, lensoid ridges, occasionally in small groups are interpreted as immature and developing axial volcanic ridges. Broad convex-flanked ridges appear to be mature sites of volcanic construction where the magma supply has been matained. Narrow, "bow"form ridges, occasionally with rectilinear flanks represent axial volcanic ridges which have ceased construction and are in initial stages of disaggregation. Isolated polygonal blocks, generally seen off-axis represent the most advanced stage of dismantling of the surface crustal accretionary features. While this cyclicity is speculated to be related to periodic magmatic and tectonic focusing and defocusing along the axis due to the restricted magma supply at this slow spreading ridge, it is recognised that this needs modelling. Although the oblique spreading of the Reykjanes Ridge allows one to clearly distinguish between structures attributable to plate spreading from those related to small scale accretionary processes, we propose that the major processes identified in this study are directly applicable to all slowspreading sections of mid-ocean ridges.



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### 1. INTRODUCTION:

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The Revkianes Ridge comprises more than 1000 km of the Mid-Atlantic Ridge extending from the southwestern insular margin of Iceland at around 64°N to the Bight Fracture Zone at 57°N (Figure 1). It is oriented at 036°, and has a half spreading rate of 10 mm/yr in a 099° direction (Talwani et al. 1971; Vogt and Avery 1974: Minster and Jordan 1978). Conventional bathymetric surveys of the ridge have been made by a number of workers, including Johnson et al (1971), Talwani et al (1971), Vogt (1971, 1974) and Jacoby (1980). These studies show that while the gross morphology of the Revkjanes Ridge in the north is more like a fast-spreading plate boundary (having an elevated axial region), in the south it is more typical of slow-spreading ridges (in having an axial valley defined by inwardfacing faults). This systematic variation in cross-sectional profile of the ridge, and its concomitant increase in depth from subaerial exposure on Iceland in the north to around 2500 m water depth in the south, is considered to result from higher temperature in the mantle with proximity to the Icelandic hotspot (Voot 1971. 1974). More recently, this north-south contrast in axial morphology has been explained by variations in coupling between the brittle upper and the ductile lower parts of oceanic plates at spreading ridges as a function of distance from the hotspot (Chen & Morgan 1990 a and b).

These general statements of ridge aspect, however, are insufficient to explain the detailed morphology of the axial zone. In particular, Searle & Laughton (1981) describe a bathymetric segmentation defined by a series of approximately northsouth trending elevated axial ridges, 20 to 30 km long and 3 to 6 km wide, arrayed in an overlapping and en echelon pattern. This geometry encompasses characteristics of second to third orders of segmentation (as defined by Macdonald et al 1988). New geophysical data has enabled us to make more precise observations and deductions about the processes of segmentation that operate on- and immediately off-axis. We suggest that it is only at lengthy sections of obliquely spreading plate boundaries, such as the Reykjanes Ridge, that it is possible to differentiate and quantify those volcanic and tectonic processes that are associated with extensional dilation and extrusion at the axis from those derived from mantle upwelling, rollover and outflow at the margins of the spreading plates.

### 2 METHODOLOGY

2.1 Existing data:

Following numerous studies using conventional bathymetric mapping in the 1970s including Talwani (1971), Jacoby (1980) and others, geophysical surveying of the Reykjanes Ridge between 58° and 61°N by Searle & Laughton (1981) combined GLORIA long range and ship-mounted side scan sonar and bathymetry to map the ridge crest and axial valley. This led to the identification of a series of linear ridges arrayed in an en echelon fashion within the neovolcanic zone. The ridges have a typical strike of around 014°, with crests elevated by between 200 and 500m from the axis floor. Due to their en echelon configuration, parts of up to three ridges locally appear to overlap along a flow-line. Few fault scarps are observed within the axial zone on GLORIA data, although outside the central zone. large numbers of linear reflectors oriented subparallel to the ridge axis were interpreted by these early workers as major fault scarps with throws of between 200 and 500m. Searle & Laughton interpreted the en echelon ridges in the neovolcanic zone as volcanic features, and discussed the difference in orientation between the more northerly-striking fissure complexes which they considered to form their roots in the neovolcanic zone, and the off-axis faulting oriented parallel to the plate boundary. They noted that faulting which occurred in the central zone was predominantly orthogonal to the extension direction, and they inferred the distribution of axial parallel faulting to be a direct function of the variation in strength and temperature of the lithosphere away from the plate boundary.

### 2.2 New data:

The most recent survey programme to be undertaken on the Reykjanes Ridge was completed in September-October 1990 (Parson et al., in press; Figure 1), using the R/V Maurice Ewing fitted with a Hydrosweep multibeam bathymetric mapping system (Grant and Schreiber 1990). The Hydrosweep system is a widebeam swath sonar operating at 12 kHz, capable of resolving bathymetry to better than 20 metres over a continuous swath of seafloor whose width is up to twice the water depth in normal spreading conditions. The Institute of Oceanographic Sciences Deacon Laboratory (IOSDL)'s Towed Ocean Bottom Instrument (TOBI; Rouse 1991) was deployed in areas B and C (Figure1) and provided high resolution deeptow sidescan sonographs of the seafloor as well as seismic profiles and magnetic data. 5

TOBI comprises a neutrally buoyant vehicle towed on a conducting cable from a ship via a depressor weight and an umbilical. The main instrumentation consists of a double-sided 30 kHz sidescan sonar providing a resolution of better than 5 m over a total insonified swath width of 6 km, and a 7.5 kHz sub-bottom profiler capable of penetrating up to 60 m of sediment. The vehicle was operated at a height above the seafloor of between 200 and 500 m. As discussed below, we have provisionally interpreted the broad-scale variations in acoustic backscatter on the ridge to reflect relative ages of volcanic activity, and used both shadow geometry and acoustic texture to evaluate all scales of volcanic landforms and levels of tectonism. The deeptow vehicle was also fitted with a transmissometer for the detection of hydrothermal plume signatures.

During the EW9008 cruise, we surveyed 220 km of the Reykjanes Ridge axial zone using the TOBI deeptow sidescan, insonifying a total of more than 2000 sq km of seafloor (Figure 1). In addition, we recorded over 35000 sq km of reconnaissance Hydrosweep multibeam bathymetry data. These data were combined with earlier recorded Hydrosweep data (cruise EW9004, Shor et al, work in progress), providing close to 100% coverage of the axial ridge from 64°N southwestwards to 57°N. In order to constrain spatial and temporal variations in along-strike morphology over the 800 km of ridge studied, three areas were selected for detailed sidescan surveying and off-axis swath mapping (Areas A, B and C; centred on 62°, 60° and 58°, respectively; Figure 1). These areas were selected both to assess the various effects of the Iceland hotspot as well as focus attention on sites of most recent studies e.g. at 62°N (submersible studies - Holcombe, Uchupi and Ballard, unpublished manuscript); at 60°N, (shallow seismic activity -Nishimura 1989); and at 58°N. (hydrothermal activity - Kuznetsov et al. 1985). Operational difficulties at the most northerly site prevented deployment of the TOBI deeptow, but data were successfully recorded from the central and southern areas.

### 3 OBSERVATION AND INTERPRETATION

3.1 Area A - (61°30'N to 62°30'N)

Our studies in the northern Reykjanes Ridge confirm previous workers' observations that its gross morphology consists of an axial zone striking 036° which is dominated by a central ridge or high. At 62°N, the average depth of the ridge crest is approximately 900 m, and it is clearly crossed by a number of linear ridges oriented approximately 014°, which are between 20 and 45 km long and 1.5 and 6 km wide (Figure 2). We refer to these features as en echelon Axial Volcanic Ridges

(AVRs). Spacing between the ridges varies, and neighbouring AVRs are of different size and shape, often overlapping one another by up to 50% of their length. With our new swath bathymetric data we are able to make very precise analyses of their form, size and structural variations along strike.

In their most robust form, the AVRs form periclinal features, elevated above the main floor of the axial zone by more than 300 m, and locally shoaling to a water depth of less than 550 m. In plan view, they have a crude lensold form with irregularly lobate margins and taper towards irregular, blunt tips. We provisionally interpret these features as constructional volcanic ridges, whose acute linear forms suggest they are built by a combination of diffuse and point sourced fissure-controlled extrusion. Direct observations by submersible confirm their architecture to be of superposed pillow-lava complexes forming mounded flows. The smallest and narrowest AVRs are characterised by discrete linear volcanic ridgelike features, up to 5 km long and 0.5 to 1 km wide, and occasional conical seamounts. The larger, broader AVRs are commonly ornamented with abundant flat topped volcanoes or seamounts, occasionally with calderas or breached craters. Since the multibeam bathymetry is particularly successful in imaging linear scarps of greater than 40 m throw, we were able to guantify fault geometries in the axial zone at all but the smallest scales. In area A, within 10 km of the axis, the ridge has only a few axis-parallel faults, with throws of generally less than 100 m. Within the axial zone, rectilinear scarp-like features oriented subparallel to the 014° AVR trend are common and suggest a strong degree of tectonic control of the extrusives.

#### 3.2 Area B: (60°35'N - 59°40'N)

This central area has been the focus of recent reconnaissance surveying following the detection of earthquake swarm during May 1989 (Nishimura, 1989), and more recently by Cann, (pers comm; November 1990). Latest follow-up studies have included acquisition of SeaMarc 2 sidescan sonar data (Shor et al, work in progress), with which we were able to identify some of our target areas.

Our survey data in the central Reykjanes Ridge confirms that the whole ridge lies deeper, with a more pronounced axial valley than in Area A, in an average water depth of around 1250 m. A number of AVRs rise above the axial floor to about 850 m water depth, and in this area they are generally larger in all dimensions than further north. They again are oriented about an azimuth of 014°. They are characterised by a considerable variety of form, ranging from narrow to broad features with convex and concave flanks dipping at less than 30° away from irregular crests. Their margins are commonly embayed and irregularly indented.

As in Area A, they are locally truncated by sharp rectilinear scarps suggesting either syn-constructional, or post-constructional faulting, or both.

Deeptow TOBI sidescan data (Figure 3a, interpreted in Figure 3b) reveal details of small scale tectonic and volcanic features. The central axial zone, generally 6-10 km in width, is characterised by acoustically strongly reflective seafloor, dominated by an irregular, mottled or blotchy texture. In detail, this mottled texture can be recognised as comprising a series of small mounds, between 20 and 200 metres in diameter, which from shadow geometry must be less than 50 m height. These hillocks are superposed, or mounded upon each other, forming a highly distinctive hummocky texture reminiscent of a cauliflower heart (Figure 3). In some places this neovolcanic terrain occupies the entire axial valley floor; in other sites there is a smooth, uniformly highly backscattering acoustic texture that we interpret as sheet-like extrusive surfaces. Locally, the mounds comprising this "cauliflower" texture build upon each other, coalescing to form more elevated areas of ridge floor, creating a lumpier, and more centralised topography (Figure 3). We tentatively interpret these larger mounds as stacks of pillowed basaltic flows, forming knoll-like elevations estimated to be up to 500 m across and 50 to 150 m high. These features essentially embryonic or aborted seamounts, and directly analogous to pillow mounds preserved in ophiolites (e.g. Troodos, C.J. MacLeod, pers. comm.). Where these larger constructs merge, they form the AVRs referred to above, and are directly correlated to the axial ridges described by Searle and Laughton (1981). Narrow AVRs, also characterised by hummocky-volcanic terrain, often have sharp-crested spines (between 1 and 2 km long, 100 to 300 m wide, and 50 to 100 m high) and are usually surrounded by flatter, sedimented seafloor. Flattopped seamounts, 0.5 to 2.5 km in diameter and 50-150 km high, resembling shield volcano forms described from sub-glacial eruptions, are abundant in this area. Mindful of the restrictions in the resolution of our data, we here define a volcano or seamount as aroughly circularextrusive form, large enough to be traced by 2 concentric closed 25 m contours on our swath bathymetry data.

In general there is a systematic variation with size in the nature of the construction of the ridges on the AVR. Narrow AVRs are characterised by a hummocky texture, consisting of a multitude of overlapping mounds, each 20 to 100 m in diameter, and a few tens of metres high. Broader AVRs also have hummocky-textured terrain, but are usually dominated by individual, conical and flat-toppped seamounts, 0.5 - 1.5 km in diameter, and 50-150 m high. In places, some broad AVRs are characterised by clusters of parallel, linear ridges interpreted as fissure-swarms that are associated with the only AVR-parallel faults within the fault zone. More regular conical or flat-topped volcances up to 1.5 km wide are observed

distributed often randomly throughout the axial valley. Their shapes have striking similarities to those exposed on land in Iceland, where these sub-glacial table mountain forms are usually composed of the less evolved, olivine tholeiitic lavas (Jakobssen et al 1978; C. Walker, PhD thesis)

Throughout Area B, narrow (20-200m wide), linear, strongly backscattering targets between 500 m and 10 km long are observed, and are interpreted as fault scarps. Axis-parallel faults are common off-axis, where they are marked by openly curvilinear, inward-facing scarps with throws of up to 200 m (Figure 3b). Within the axial zone, axis-parallel faults are relatively few and have undetermined throws of less than 50m. AVR-parallel faults within the axial zone are generally low angle, between 10° and 30° and dip inwards towards the centres of the AVRs. We recognise, however, that the fault dips are a minimum estimate, since in calculating them we are unable to take into account the presence of any scree slope.

There are few AVRs which are completely unfaulted; almost all of them show evidence of brittle deformation to some degree. Commonly this includes a range of structures from fine, closely-spaced strands or bundles of lineations, (unresolved on our multibeam data but clearly imaged by TOBI deeptow sonographs, X on Figure 3a). These can be seen to individually correspond to throws of less than 20 m up to subparallel continuous scarps extending for several tens of kilometres. Some fractures observed on TOBI data have negligible offset on swath bathymetry data. These latter lineaments are more recognisable as the predominantly inward-facing dip-slip structures first recognised by Searle & Laughton (1981).

Even within the 100 km length of the axial zone surveyed within Area B there are distinct variations. In the northern section of Area B, in a portion of the ridge which we interpret to be recently volcanically active because of its well-developed and relatively undeformed state, it is possible to map out the largest single AVR on the ridge, centred on 60°15'N. This is 25 km in length and 6 km wide at the 900 m isobath (Figure 4). Here, the floor of the axial zone generally lies at between 900 and 1050 m water depth, but the shallowest parts of the AVR reach less than 600 m. There are a number of flattened conical and shield constructs, but the majority of the central 10 km wide zone of the axial floor in this area imaged on TOBI records is characterised by hummocky, cauliflower-textured, coalesced AVRs.

At the southernmost portion of the Area B, the AVRs are more clearly defined by the 1000 m isobath, have steeper sides and have a higher aspect ratio (>20 km long and up to 4 km wide). Here, several closely spaced AVRs have clearly overlapping flanks, and are marked by a strong alignment of volcanoes along their crests, which rise to less than 700 m water depth from the axial floor lying at

between 1000 and 1150 m (Figure 4). Numerous, mostly flat-topped volcanoes which extend up to 1.4 km in diameter at their base are scattered throughout the area, interrupting both the AVR terrain and the intervening axial floor.

Throughout Area B, volcanoes within the axial floor or forming parts of the AVRs show a large variety of shapes and states of deformation. Within the broad spectrum of constructional volcano forms recognised, there are many modified examples, including breached or turretted rims, single and composite collapsed calderas, and secondary, parasitic or compound cones. Brittle deformation of the volcanoes is readily recognised on the TOBI data as fractures, fissures and faults, and extensively deformed and fragmented volcanoes are commonplace (Figure 3a). Whilst perfectly circular plan-forms are ubiquitous both on- and off-axis, numerous examples of 'D-form' or 'bow-form' morphologies are observed where faults have down-thrown and dismembered volcanic cones (Figure 2a). Elsewhere in the central parts of the axial zone, sidescan data clearly show multiple sub-parallel faults cutting volcanoes with fault throws of less than 20 m, and while these volcances broadly retain their original form, the pervasive cross-cutting fabrics suggest they are in the earliest stages of being dismantled. The overall trend bulk/mean? of the major faulting defined by the Hydrosweep bathymetry within the central 5-10 km of the axial zone varies from a strike of 014° within the axis to 036° at the shoulders.

### 3.3 Area C: ( 57°45'N - 58°20'N)

The southernmost survey area has the only pronounced axial valley of the three areas studied. Throughout Area C, the average of the axial floor is at around 1850 to 2000 m water depth, and the approximate crestal depth of the AVRs lies just shallower than 1500 m. The most striking difference in the regional morphology of Area C from Areas A & B is in its well-developed axial-parallel median valley walls. In Area C, inward-facing scarps between 10 and 12 km apart descend up to 500 m from the valley shoulders to the axial valley floor. In cross-section, the axial graben are clearly asymmetric. Steeply dipping (at about 45° fault scarps) generally define only one wall of the valley, while the opposite face is less regular, and more shallowly inclined. Searle and co-workers (Owens, et al.1991) have found a similar asymmetry in the distribution of crustal magnetization in the area. The faults bounding the axial valley are only continuous for a maximum of 30 km, and give way along strike to inward facing fault scarps on the opposite flank of the median valley. This 'flip-flop' asymmetric character of the

axial-graben faults is not unique, and is recognised elsewhere on the Mid-Atlantic Ridge (e.g., in the TAG area - Karson and Rona, 1990; Lin et al., 1990).

AVRs in Area C have the greatest elevation, up to 500 m above an average depth of the axial zone of 1950 m. They range in character from narrow to broad features, and are dominated by a hummocky terrain (Figure 2b). The narrower AVRs have linear volcanic ridges and conical seamounts, and the broader ones are more commonly ornamented with flat-topped seamounts. Sharply defined flanks of the AVRs are common, and suggest a high degree of structural control. On TOBI data, the central portions of AVRs can be seen to be dominated by the familiar "cauliflower" texture of mounded, hummocky piles of volcanics, but the acoustic reflectivity as recorded on both near-bottom echo-sounder and seismic profiles recorded over most of the AVRs in this area is more subdued than in the north. We interpret this as indicating significant sediment blanketing and hence a likely greater age.

Volcanoes are less common, but have the range of morphologies and dimensions as described above; ie, conical to flattened, breached, turretted and collapsed (Figure 5). There is a minor number of small volcanoes (less than 500 m in diameter) which show little or no brittle deformation. Substantial areas between the AVRs within the axial zone are characterised by flat-lying seafloor of weak reflectivity, which although being locally intensely faulted, shows no evidence for neovolcanic activity. These areas we interpret as sedimented axial valley floor that is magmatically inactive. The proportions of constructional volcanic axis to this sedimented floor observed here is the highest of all three areas studied. All of the large scale faulting is inward-facing, synthetic in style, but on sidescan data it can be seen to be often associated with very tightly clustered bundles of small-offset antithetic faults which characteristically face into the main fault plane. There is also some intersection of AVR-trending fault structures with both axis-parallel faulting, and spreading-parallel structures, producing a finely-spaced reticulate pattern.

### 4 SUMMARY AND DISCUSSION

4.1 En Echelon Volcanic Ridges (AVRs)

Our mapping of the morphology and the deformation state along the Reykjanes Ridge lead us to the following observational statements:

1. The typical morphological unit of volcanic construction at the Reykjanes Ridge is a convexo-convex elongate "lens-shaped" periclinal structure, generally

source. Chemical and structural similarities between the Icelandic and Atlantic Ocean crust will be emphasized.

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The opening of the Atlantic may be related to the development of the Icelandic and other plumes that underlie the ridge and flanks of the Mid-Atlantic Ridge (MAR) (Bott, 1988; Silver et al., 1988; White and McKenzie, 1989a, b; White, 1990). However, plumes are not necessarily crucial to continental rifting, although the 1-2km dynamic thermal uplift they provide helps gravity sliding (White and McKenzie, 1989a). In the North Atlantic, the Icelandic plume was active shortly before the opening of the ocean and small-scale rifting had been occurring for tens of millions of years before the Icelandic plume was initiated at approximately 62 Ma. After a 5 Ma period of plume activity, the rifting between Greenland and the Rockall Plateau became well established (White, 1988; White and McKenzie 1989a). Plate driving forces such as slab pull, ridge push and suction forces at well established subduction zones are thought to be collectively responsible for the opening of the Atlantic (White, 1990), rather than whole mantle convection associated with plumes and hot-spots as was earlier suggested by Morgan (1971).

### 13.2 Morphology and structure of the Mid-Atlantic Ridge

The Atlantic basin is deeper, older and spreading faster in the south than in the north. For example, half-spreading rates are  $15 \text{ mm y}^{-1}$  at the Kane Fracture Zone,  $11-12 \text{ mm y}^{-1}$  at the Azores and  $10 \text{ mm y}^{-1}$  on the Reykjanes Ridge and Iceland. The mid-ocean ridge (MOR) is intermittently segmented and offset by transform regions of various magnitudes. Topographic anomalies occur in the vicinity of hot-spots and may be represented by subaerial groups of oceanic islands developed at various distances from the ridge.

High resolution bathymetric studies (Luyendyk *et al.*, 1977) and submersible dives (first executed in the FAMOUS area, 37 N: Bellaiche *et al.*, 1974; Needman and Francheteau, 1974) have enabled earth scientists to study the nature of the seafloor topographic fabrics in great detail and to appraise and discuss their significance in relation to crustal and mantle structures.

The morphology and topography of the ocean basin can be described in terms of sets of parallel ridges and troughs at various orientations to the main spreading direction, which have varying widths and lengths (Figure 13.1a). In a typical slow spreading ridge environment, such as the North Atlantic, the intensity of the fabric is greater than at a fast spreading ridge. The arrangement, intensity and relative proportions of these lineaments are determined by such features as a median valley at the ridge crest (average 30 km wide), an axial neovolcanic zone (usually < 10 km) and ridge axis discontinuities such as transform faults. The MAR is segmented by large transform faults and by smaller non-transform discontinuities (Ramberg *et al.*, 1977; Schouten and







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White, 1980; Macdonald and Fox, 1983; Lonsdale, 1983; Macdonald *et al.*, 1984; Sempere and Macdonald, 1986) developed, on average, every 55 km (Macdonald, 1986). Recent work (Lin *et al.*, 1990; Sempere *et al.*, 1990) confirms that the median valley floor of a segment of ridge is itself elevated towards the centre of that segment (Figure 13.2). It has been demonstrated on the East Pacific Rise (EPR) that these segments behave as structurally and geochemically independent units (Thompson *et al.*, 1985; Langmuir *et al.*, 1986).

The plate boundary as represented by the axial rift zone may gradually migrate relative to a globally fixed reference frame, such as a hot-spot, by the normal spreading process. This should be distinguished from the processes that involve the relocation of the plate boundary, both laterally and axially, by ridge jump and rift propagation, respectively. The former is not uncommon in the Atlantic and has been reported from 36°N (DSDP Leg 37) and 23°N (ODP Leg 106/109). To the north of Iceland, the Kolbeinsey ridge is believed to have undergone several episodes of ridge jump (Johnson *et al.*, 1971) before the last jump occurred at 36 Ma (Oskarsson *et al.*, 1985).

The detail of the axial neovolcanic zone in a symmetrically spreading ridge is usually defined as an undulating central topographic ridge of young volcanic products erupted from fissures. It has been suggested that the slower spreading ridges with lower magma supply rate typically have more than one locus of volcanic activity (i.e. volcano) per segment, relative to the fast spreading ridges that typically have one volcano per segment (Searle, personal communication). The volcanic ridges become laterally removed from the axis by spreading, where they are faulted to form steep rift mountains up to 1500 m or more high (Bougault and Hekinian, 1974; Sempere et al., 1990) and which mark the margins of the median valley typical of a slow spreading ridge. The topography on the immediate flanks of the median valley is thus very pronounced with successive parallel ridges and scarps, with several hundred metres of relief, which becomes less pronounced as a function of distance from the ridge axis, due to sediment burial and subsidence as the lithosphere cools (Parsons and Sclater, 1977), until the slope levels off at the foot of the ridge rise (Figure 13.1b).

These characteristic geomorphological features of the ridge are disturbed in the vicinity of a hot-spot. The depth to the ridge shallows, and the median valley and segmentation become less pronounced. This is demonstrated south of Iceland, on the Reykjanes Ridge (Talwani *et al.*, 1971), which is adjacent to the Iceland hot-spot, the largest plume in the North Atlantic. Vogt (1971) observed a topographic fabric oblique to the axial valley and the magnetic stripes (time-transgressive ridges) south of Iceland and near the Azores, but not near Jan Mayen (Vogt, 1974). A hot-spot often leaves a trail in the form of an aseismic ridge or a chain of islands (Chapter 9), depending on whether it coincides with a mid-ocean ridge segment. The aseismic ridge between the Faeroes and Greenland records the history of the Iceland hot-spot from 60 Ma. Assessment of the topography reveals a subdued rate of eruption from the plume between 20 and 30 Ma (Vogt and Avery, 1974).

Seismic refraction profiles through the oceanic crust have shown that there are principally three seismically distinct layers of material (White *et al.*, 1990). The crust is defined here to be from the surface down to the petrological Moho, represented by the base of the cumulates. The layers are distinguished by their different seismic P-wave velocities, which are determined by a combination of pressure, temperature, porosity, density and mineral composition and alignment. From this work the average crustal thickness of the oceanic lithosphere, away from interference of the ridge-axes, hot-spots, seamounts, or transform faults, is 6 km (White 1989). The thickness at the ridge-axis is variable. The crustal thickness along a single ridge segment is less in the vicinity of the ridge axis discontinuities, and greater at the centre of each segment. For example, in the MAR. 24 to  $30^{\circ}$ N, the crust is thinned



Figure 13.2 (a) Plan view of different styles of ridge segmentation between fast and slow spreading ridges. (b) Schematic cross-section of ridge segmentation (adapted from Lin et al., 1990).

by as much as  $50\%_0$  at the segment boundary and it has been suggested that the longer the segment, the thicker the crust (Lin *et al.*, 1990) (Figure 13.2).

Earthquake epicentres have been reported at depths of 6 km in the Atlantic region (Lillwall *et al.*, 1978). However, fault spacing often suggests that the crust is 2 km thick and so not completely formed, but that it subsequently thickens rapidly within a short distance (about 15 km) from the ridge axis. This could indicate that either the crust is of variable thickness at the ridge axis, being thicker at slower spreading centres relative to fast ridges, or that the sub-ridge asthenosphere is capable of brittle fracture, suggested by the presence of basaltic dykes observed in ophiolite sections (Browning, 1984).

The Atlantic oceanic crust increases in thickness towards Iceland along the Reykjanes Ridge, and in the vicinity of other plateaux, such as the Azores Islands (Searle, 1976a), Cape Verdes Islands (Courtney and White, 1986) and other seamounts in the North Atlantic. This increase in crustal thickness is representative of an increased rate of magma production, which correlates with mantle potential temperature (White and McKenzie, 1986b). There is confusion over the relationship between the magma flux and the spreading rate. White (1989) states that due to the constant average thickness of oceanic crust (6-7 km), the magma flux is independent of the spreading rate and that asthenospheric processes exert a dominant control over the volume and distribution of the magma in the ocean basins. However, other workers (Dick, (1989) frequently describe low magma fluxes on slow spreading ridges. This parameter must be variable in some instances with episodic volcanic activity and due to the changes observed at plumes and within a single segment. On a typical spreading segment, in either the EPR or the MAR, there is a comparable average crustal thickness of 6-7 km, although local deviations occur. This average thickness will have taken longer to produce in the MAR relative to the EPR, and so it follows that the Atlantic must have an overall lower magma supply flux for a given period of time. Nevertheless, for any given unit volume of mantle, regardless of the spreading rate and time, the actual percentage of melt produced from that volume will be comparable at about 30%. It follows that the spreading rate should not significantly affect the bulk chemistry of the magmas.

Faulting may or may not have significant effects on the magmatic processes operating at the axial zone, and so will be included here, very briefly, for completeness. In oceanic environments the faulting is dominantly extensional (Taponnier and Francheteau, 1978). In general terms, inward dipping bundles of normal faults are observed on both fast and slow spreading ridges. They are formed at approximately 2 km intervals (Searle, 1984), although the spacing is greater on slow spreading ridges and the fault scarps are longer. The movement on these faults ceases at a distance of 5-30 km from the axial zone (Taponnier and Francheteau, 1978). The style of extension varies as a function of spreading rate and local magma flux. Extensional faulting in the Atlantic, and on other slow spreading ridges, at a time when the magma flux

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is very low, is believed to mimic that on the continents (Sempere et al., 1990). Detachment faults similar to those seen in the Basin and Range Province (western USA) are observed in the MARK area of the Atlantic (23°N) (Brown and Karson, 1988). White et al. (1990) have documented many different styles of faulting, of which three major types have been observed to penetrate to the Moho (Chapter 3: Figure 3.6). Serpentinization occurs at the base of transform faults which penetrate the whole crustal sections, and this lowers the seismic Moho (layer 4 boundary) by as much as 2-3 km (White et al., 1990). If any of these faults were activated in the vicinity of the active axial zone, then they could be a means of serpentinizing the lower crust and so producing apparently less variation in the seismic crustal thickness away from the axis. This is interesting because the crustal thickness at the axial zone is variable, yet at some distance from the axis it is reported to be more consistent and raises the question as to whether the seismic Moho is largely representative of a serpentinized upper mantle (Hess, 1962; Clague and Straley, 1977; Francis, 1981).

# 13.3 Morphology and structure of Iceland

The regional topography and morphology of Iceland is dominated by a central high (2000 m above sea level) in the vicinity of the Vatnajokull region (Figure 13.3). The Mid-Atlantic plate boundary is exposed on land, where it presents an excellent opportunity to study the oceanic crust. Before describing the details of the ridge crest in Iceland, it is necessary to introduce the basic geological characteristics of the island as these will be referred to throughout the chapter and are vital to the geochemical interpretation.

# 13.3.1 Present plate boundary configuration

At 64°N, the Mid-Atlantic plate boundary is expressed on land as a complicated arrangement of active volcanic and seismic zones (Figure 13.3a). These will be briefly described below, but for more detailed descriptions, see Ward (1971), Palmason and Saemundsson (1974), Saemundsson (1974, 1978, 1979) and Einarsson (1990). Throughout the literature the various volcanic zones have often changed their names (for example. Brooks *et al.*, 1974; McGarvie *et al.*, 1990); this chapter will follow the most widely used version, summarized in Figure 13.3a.

The main actively spreading volcanic zones are divided into the Western, Eastern and Northern neovolcanic zones. The Western Volcanic Zone (WVZ) is the on-land continuation of the Reykjanes Ridge. It extends inland, becoming progressively wider and trending dominantly north-east for half the distance across Iceland. It then bends due east at Langjokull to join the Eastern and Northern Volcanic Zone junction under Vatnajokull. (Figure 13.3). The

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Figure 13.3 (a) Location map of Iceland, showing the distribution of the volcanic and fracture zones, and locations mentioned in the text. Bathymetry contours are drawn every 400 m (adapted from Einarsson 1990). (b) Map of Iceland showing the distribution of the different magma series (after Jakobsson 1979 b).

Northern Volcanic Zone (NVZ) has a more northerly trend, and extends north of the Vatnajokull glacier to the north coast of Iceland. The Eastern Volcanic Zone (EVZ) is the widest (72 km) and trends south-west parallel to the WVZ from Vatnajokull, as far as the island of Surtsey. A comparatively small volcanic zone protrudes from the south-east corner of Vatnajokull, and has been referred to as the Oraefajokull Volcanic Zone (OVZ) (Steinthorsson 319

et al., 1985). There is a possible fifth active volcanic zone, trending from east to west, on the Snaefellsnes Peninsula, west Iceland (SVZ). It has a seaward expression, in the form of a topographic ridge (Jokulbanki Ridge), that swings round from an east-west orientation to run parallel to the Reykjanes Ridge for at least 150 km (Brooks et al., 1974). In much of the recent literature, the EVZ, OVZ and SVZ are referred to as lateral or flank zones (terminology of Saemundsson, 1979), as many volcano-tectonic properties differ from the rest of the neovolcanic zone, which is more representative of the typical mid-ocean ridge environment.

There are two zones of high seismicity which represent transform zones (Einarsson, 1991). In the north-east of Iceland the neovolcanic zone is sinistrally offset by the 75 km wide, 100 km long belt, known as the Tiornes Fracture Zone (TFZ) which connects the NVZ with the Kolbeinsey (or Iceland-Jan Mayen) ridge. In detail, there are at least three parallel north-west-south-east trending seismic belts roughly 30-40 km apart (Saemundsson, 1974; Einarsson, 1991). The dominant structural orientation within each belt is north-south and is displayed by a series of horsts and graben. In the south of Iceland there is an east-west trending, sinistral zone of high seismicity, known as the South Iceland Seismic Zone (SISZ), that is 10-15 km wide and 100 km long (Einarsson et al., 1981). It joins the WVZ and EVZ without laterally offsetting either. The majority of the region is covered by an alluvial plain which hinders ground level examination of the region. Nevertheless, in the east, the topography is deflected from the normal north-east trend to an east-west orientation, and in the west, from the normal north-east trend to a north-south direction (Saemundsson, 1967; Walker, to be published).

Segmentation in Iceland, on a smaller scale, occurs in the form of volcanic systems (see later) (Saemundsson 1978). These are frequently arranged in an *en echelon* fashion in response to non-orthogonal spreading. It has been suggested that these behave as structurally and geochemically independent units, similar to segments on the EPR (Whitehead *et al.*, 1984) and possibly the MAR (Sempere *et al.*, 1990), although there is evidence to show that this is not always the case in the flank zones (McGarvie 1984; Blake 1984; McGarvie *et al.*, 1990).

# 13.3.2 The neovolcanic zone

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The axis of the ridge crest, often referred to as the neovolcanic or axial zone in Iceland, has been described in detail by many workers (Saemundsson, 1978, 1979; Jakobsson *et al.*, 1978; Jakobsson, 1979a, b). The topography of the neovolcanic zone can be described as a slightly, but variably, elevated region, relative to the distal parts of Iceland. Geomorphologically, it is made up of a number of individual units or volcanic systems (Walker, 1963, 1974; Saemundsson, 1978) (Figures 13.4a and 13.5). Each system is composed of a set of parallel eruptive ridges or fissures, constituting a fissure swarm that





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Figure 13.5 (a) Plan view of the volcanic system, as defined by Jakobsson (1978), showing the spatial relationship between the central volcano (CV), faults and fissures (F + F) and the off-axis lava-shield or shield volcano (SV). (b) Schematic cross section of the topography across a volcanic system (as observed in SW Iceland, e.g. CV = Hengill and SV = Skjaldbreiur or Ingolfsfjall). The spreading axis is indicated by arrows.

varies in length from 17 to 105 km, in width from 5 to 30 km (Jakobsson, 1979a), and which has a life span between 300 and 500 ka. These swarms may, with time, develop a central volcano (Saemundsson, 1978). The latter is characterized by being topographically elevated relative to the fissure swarm due to an increase in the magma extrusion at one site (Figure 13.5b). The system may also contain earlier slightly off-axis (5–10 km) shield volcanoes or lava shields of various sizes. These can be recognized by the unusually large accumulation of compositionally monotonous compound lavas and very low angles of slope on the flanks, typical of Hawaiian shield volcanoes or Skjaldbreidur (WVZ). It is important to appreciate that the volcanic morphology is greatly dependent on whether the eruption occurs under ice or subaerially (Thorarinsson, 1974).

It seems that there is often a correlation in Iceland between the composition of the lavas and the morphology of the volcano from which they were erupted (Jakobsson *et al.*, 1978, 1979a, b). Some of the variation between the different volcanic systems in Iceland can be attributed to their different stages of development (Jakobsson 1979a), and/or as a function of crustal thickness in that region. Away from the neovolcanic zone, in the Tertiary Icelandic crust, Walker (1963) identified similar volcanic systems, but the off-axis shield volcanoes appear to be less abundant (see Section 13.5.1).



Figure 13.4 (a) Map of Iceland showing the distribution of the 29 volcanic systems (stippled) (after Jakobsson 1979 b). (b) Map of Iceland showing the relative ages of the rocks. Wide horizontal = Tertiary (> 3.1 Ma); close horizontal = Plio-Pleistocene (0.7-3.1 Ma); clear = Upper Pleistocene and post-glacial (< 0.7 Ma); stippled = solid drift (Sæmundsson 1979). (c) Distribution of recent seismicity. Epicentres are taken from earthquake lists (PDE) of the US Geological Survey for the period of 1963-1987. Only epicentres determined with 10 or more stations are included. Focal mechanisms are shown schematically on the lower hemisphere stereographic projection of the focal sphere. Compressive quadrants are black (after Einarsson 1990).

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The thickness of the crust in Iceland (Palmason, 1971; RRISP Working Group, 1980) varies from 8 to 16 km (compared with the range of 3-8 km for normal oceanic crust). The boundary between layers 3 and 4 is interpreted as the base of the crust (Figure 13.6). It is shallower beneath the central volcanoes of the neovolcanic zone and the extinct central volcanoes in the Tertiary and Skagi zones, suggesting that crustal and mantle lithospheric underplating occurs later under Iceland (Figure 13.6b). No crustal thinning



Figure 13.6 (a) Map of Iceland showing the neovolcanic zone (stippled) and the dips of the older lava successions (adapted from Pálmason and Sæmundsson, 1974). Note that the dip is dominantly towards the neovolcanic zone, but also two synforms are present and maybe interpreted as relict sites of the plate boundary. The traverse shown in 13.6b is indicated. (b) Seismic section of the south end of the EVZ and the east flank of the Reykjanes Ridge, showing a generalized crustal and upper mantle section. Letters indicate positions of large shots. Numbers of P-wave velocities are indicated in km/s. Crustal layers are continuous across the transition from the Reykjanes Ridge to Iceland, whereas there is a drastic change in the upper mantle structure close to the shelf slope, where a well developed oceanic lithosphere abuts on anomalous upper-mantle material (RRISP working group, 1980).

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is observed in the vicinity of the transforms or the southward propagating tip of the EVZ and there is no observable change in thickness in the area of the hot-spot crest (RRISP Working Group, 1980). Areas where there is a noticeable increase in the thickness include northern and north-western Iceland, and the south-eastern corner under the proposed propagating EVZ (Palmason, 1971; Flovenz, 1980; RRISP Working Group, 1980).

Faulting in Iceland in the neovolcanic zone is dominated by normal extensional faulting, and is characterized by a horizontal minimum compressive stress in the north-west direction. The maximum stress varies between volcanic zones and between the latter and transform regions (Einarsson, 1990). The faulting varies in style and intensity over the neovolcanic zones, and seems to correlate with the morphological type of volcano. The volcanic systems of the WVZ and the NVZ show parallel fissure swarms (both eruptive and non-eruptive), well defined extensional normal faulting and shallow axial grabens (2 km wide). Superimposed on this regional stress in the more evolved volcanic systems, is a radial component induced by the presence of a central volcano (Ode, 1957). The off-axis shield volcanoes lack extensive normal faulting. The other main volcanic zones, or flank zones (EVZ and SVZ), lack the pronounced structural features described above, especially in the more distal regions of each zone (south and west, respectively) (Jakobsson 1979b). The structure of the Tertiary volcanic systems in eastern Iceland has been interpreted in terms of volcanic systems similar to those of the WVZ and the NVZ of the neovolcanic zone (Walker, 1963; Helgason, 1984, 1989). The lava piles on Iceland have been mapped and found to dip towards the present neovolcanic zones. There are two flexural monoclines in the Skagi region and in north-east Iceland (Figure 13.6a).

### 13.3.3 Evolution of the Icelandic plate boundary

The Iceland plateau began its evolution around 36 Ma (Oskarsson et al., 1985) when the magma plume flux rate relative to the spreading rate increased (Vogt and Avery, 1974). From this period to the present, the location of the neovolcanic zone has never reached tectonic stability, as the relative position of the plume and the Mid-Atlantic plate boundary changed by a series of ridge jumps and ridge propagation episodes (Oskarsson et al., 1985). The movement was initiated by the westward drift of the plate boundary relative to the stationary Icelandic plume. Although there are differing interpretations of the evolution of the plate boundary (Ward, 1971; Palmason, 1973; Saemundsson, 1974; Sigurdsson et al., 1985), the following summary provides an outline of the origin of the currently active and recently extinct zones (Figure 13.7).

At 36 Ma, the plume was situated to the west of the plate boundary, and the Kolbeinsey ridge propagated northwards from the plume, causing



Figure 13.7 (a-f) Evolution of the plate boundary through time (36 Ma to present) (information extracted from Sæmundsson, 1974: Oskarsson *et al.*, 1985; Jancin *et al.*, 1985): KR = Kolbeinsey Ridge: RR = Reykjanes Ridge: AR = Aegir Ridge: SKVZ = Skaggi Volcanic Zone: SVZ = Snæfellsnes Volcanic Zone: NVZ = Northern Volcanic Zone: EVZ = Eastern Volcanic Zone: RP = Reykjanes Peninsula; L = Langjökull; Patterned area indicates the location of the centre of the head of the Icelandic plume.

extinction of a more easterly ridge segment (Figure 13.7a) (Vink, 1984). The northern end of the Reykjanes Ridge was represented by the Jokulbanki Ridge that bent into the Snaefellsnes peninsula. By 27 Ma this was connected to the Kolbeinsey Ridge via the Skagi Volcanic Zone (SKVZ) (Figure 13.7b). This configuration is thought to have remained stable for 17 Ma, during which time the plate boundary moved westwards relative to the plume. At 9 Ma, instability caused the Jokulbanki-Snaefellsnes Ridge to jump eastwards, producing the WVZ (Figure 13.7c). The NVZ and the TFZ may have been activated prior to this (Oskarsson et al., 1985), or not until 7 Ma when the SKVZ became extinct (Jancin et al., 1985) (Figure 13.7d). At 3.5 Ma the NVZ began to propagate southwards (Jakobsson, 1979b; Einarsson and Eriksson, 1982) at a rate of 3.5-5 cm y<sup>-1</sup> producing the EVZ and the transient SISZ (Figure 13.7e) (Einarsson et al., 1981). At 2.5 Ma (Figure 13.7f) the extinct SKVZ was reactivated for a short period (1.8 Ma) (Everts et al., 1972), before the present day configuration was established. Since 0.7 Ma, the neovolcanic zones have changed very little, and have remained approximately in their current situation, with only minor oscillations of the order of 10 km.

The ridge jump has not always been from the west to the east, and temporally small oscillations have occurred, such as in the SKVZ (Sigurdsson *et al.*, 1978), and in north-east Iceland (Helgason, 1984, 1985, 1989), as indicated by the dips of the lavas (Figure 13.6a).

The Snaefellsnes peninsula (SVZ) has been interpreted as a transform region (Sigurdsson, 1970; Schafer, 1972; Saemundsson, 1974b; Sigvaldason, 1974; Sigurdsson *et al.*, 1978), based on east-west orientated volcanic systems (Piper, 1973) and off-axis seismicity between it and the Langjokull region, central Iceland, as reported by Einarsson (1990) (Figure 13.7c). The interpretation given here of the evolution of the neovolcanic zone suggests that it is analogous to the present day Reykjanes Peninsula, but is almost extinct. Unequal spreading rates have been suggested as a cause for the initiation of the SVZ 'fracture zone' as well as the SKVZ (Sigurdsson, 1970; Walker, 1975; Burke *et al.*, 1973). This has been challenged (Searle, 1976b) in that it is believed that it is a true volcanic zone undergoing extinction. It is not always possible to find direct modern oceanographic analogues for the plate configuration on Iceland, and this has led many workers to suggest that they may represent radial overflow of the plume (Einarsson *et al.*, 1977).

## 13.4 Mantle structure under the Atlantic and Iceland

# 13.4.1 Asthenospheric mantle flow

All thermal convection systems are driven by density differences which result from lateral temperature variations, and it has been proposed that the earth's mantle is convecting vigorously (McKenzie *et al.*, 1990). One of the foremost questions about the large-scale structure of the mantle is the exact nature of this vigorous convection: on what scale is the convection operating, and how many convective systems exist within the mantle?

The 'actual temperature' of the mantle increases with increasing depth. The 'potential temperature' is the actual temperature corrected to accommodate the effects of change in pressure and subsequent adiabatic decompression as one unit volume moves within the mantle via convection. As a block of mantle rises from (for example) the 1400°C horizontal isotherm, the particles expand and its real temperature decreases. Melting will occur when this temperature intersects the pressure and temperature sensitive solidus (McKenzie and Bickle, 1988, and references cited therein). This passive upwelling process is believed to occur beneath the mid-ocean ridges as the plates are moved apart by extension (McKenzie, 1984). More recent investigations reveal that the upwelling is not continuous along the ridge axis, and that it follows the pattern of ridge segmentation as envisaged by recent gravity work (Lin *et al.*, 1990).

Ballard et al. (1981), Whitehead et al. (1984) and Crane (1985) suggested

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Figure 13.11 REE patterns for basalts from different volcanic regions in Iceland. Enrichment factor = (REE concentration in sample)/(REE concentration in chondrites) (Meyer et al., 1985). Data for the Vestmannaeyjar are from O'Nions et al. (1973), and for the Snæfellsnes from O'Nions et al. (1977).

enriched (Wood *et al.*, 1979). The older basalts in all three holes (35-40 Ma) are more enriched in light REE relative to the younger samples near the ridge. They also report alkali basalts from the same hole as tholeiites. On closer inspection of the data, the distal hole (407) has the higher K<sub>2</sub>O and TiO<sub>2</sub> abundances and the basalts with alkali affinity are the samples with the light REE enrichment. It seems then that if smaller degrees of partial melting produced these alkalic basalts, then it follows that they should have the more enriched light REE patterns, and this is in accordance with a period of low output of the plume (Vogt, 1971; Schilling, 1975). Pb isotopic data support the notion that the Icelandic plume geochemical anomaly was present at 20 Ma, and possibly even at 28 Ma (Mattinsson, 1979).

13.6.3.2  $FAMOUS area (MAR, 37^{\circ}N)$ . The other well studied location is the FAMOUS area (37°N; DSDP Legs 37, 49 and 82) (Figure 13.10). The area is composed of 40–50 km long segments that produce both tholeiitic and alkali basalts. According to Schilling (1973a,b) the basalts are light REE enriched and this feature increases along the ridge axis towards the Azores in the north and temporally towards the present day axis. The REE patterns

from 2 Ma crust (Leg 49, hole 411) are variable, with either light REE enriched concave upwards patterns, or depleted light REE contents (Tarney *et al.*, 1979) (Figure 13.11b).

DSDP Leg 37 was designed to test the chemical variability perpendicular to the ridge axis in close proximity to the Azores hot-spot. A distal hole drilled in 16:5 Ma crust (Leg 37, hole 335) has light REE depleted patterns  $([La/Sm]_{eF} < 1)$  and depleted abundances of LIL elements (Puchelt *et al.*, 1977; Schilling et al., 1977), suggesting N-MORB characteristics. Although O'Nions and Pankhurst (1977) found various degrees of light REE enrichment in hole 335 (16.5 Ma), N-MORB patterns are still dominant. More proximal holes (334 and 332) show dominantly light REE enrichment patterns (O'Nions and Pankhurst, 1976; Puchelt et al., 1977; Schilling et al., 1977). The <sup>87</sup>Sr/<sup>86</sup>Sr ratios range from 0.70287 to 0.70316 (O'Nions and Pankhurst, 1976) with the higher values from the older 13 Ma crust, and this supports the LIL element enrichment observed by O'Nions and Pankhurst (1977), but is clearly in contrast to Schillings' predicted enrichment trend from old to young crust. The <sup>143</sup>Nd/<sup>144</sup>Nd values for the same proximal (Hole 332) samples (O'Nions et al., 1977) are low and range from 0.51309 to 0.51315, but appear to follow the systematic variation displayed with the Sr isotopes in the mantle array. Pb isotopic data are heterogeneous spatially and temporally (Mattinson, 1979).

DSDP/IPOD Leg 82 (Bougault and Cande, 1985) returned to this area (Figure 13.10 and 13.11) and drilled 35 Ma crust (Hole 556, anomaly 13). They recovered N-MORB with slightly enriched <sup>206</sup>Pb/<sup>204</sup>Pb ratios. Closer to the Azores (hole 557, 18 Ma crust) enriched basalts were recovered. Further south, hole 558 (35 Ma old crust) was drilled on the same flow line as hole 335 of Leg 37 (16.5 Ma crust). The 35 Ma crust, which contains both depleted and enriched basalts within the same hole, is similar in this respect to Hole 413, Leg 49. Continuing further south on the same anomaly (35 Ma), the compositions are enriched, both isotopically and in the light REE and other LIL elements. On the same flow line, but closer to the axis (Figure 13.10), Hole 561 showed a variation in the degree of element enrichment. South of 33°N and the Haves Fracture Zone, three more holes were sunk on the same anomaly (Holes 562 to 564), and consistently produced N-MORB. This suggests that the Hayes Fracture Zone acts as a natural barrier to the chemical effect of the Azores plume (Bougault and Cande, 1985; White and Schilling, 1978; Bougault and Treuil, 1980).

# 13.6.4 Enriched ridge segments: E-MORB tholeiites

13.6.4.1  $45^{\circ}N$  region, MAR. This area has a well developed rift valley and flank morphology akin to typical MOR (Figure 13.1b). Schilling *et al.* (1983) reported high [La/Sm]<sub>EF</sub> ratios of 2.33 and so defined it as E-MORB. Site 410, DSDP Leg 49, was drilled in 10 Ma crust and revealed considerable variations

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symmetrical, typically between 3 and 15 km wide and from 5 to 35 km in length. We refer to this unit as an en echelon Axial Volcanic Ridge, or AVR. We have examined 42 examples of AVRs, and recognise four different types. These are illustrated schematically as an axial profile in Figure 7. Morphologically, they can be summarised as follows: 1) thin ridges defined by sharply-aligned "fissure-type" volcanoes and locally hummocky terrain, 2) broader ridges with conical-seamounts, 3) whale-back shaped periclinal AVR's with abundant flat-topped seamounts, cut by AVR-parallel fracture-swarms, and 4) broad ridges that are covered by a thin blanket of sediment and dissected by both AVR-parallel and axis-parallel faults. These differences we interpret as various stages in an evolutionary life cycle of AVR development, including volcanically constructive periods (exemplified by types (1) and (2) above), to tectonically destructive periods (as seen in types (3) and (4)) (Figure 7).

2. The range of AVR aspect ratios and the symmetry of the AVR shape is a function of the stage of development of the construct. (a) Broad, convex AVRs are mature structures within which the supply of magma is maintained, and the extrusive component is largely controlled by linear fissures suggested to lie at the root of the AVR. (b) Short, narrow, lensoid ridges which occasionally occur in groups or closely overlapping arrays are interpreted as immature, or developing AVRs. These are occasionally bounded by rectilinear flanks, which we interpret to indicate strong tectonic control on the volcanic activity building the AVR. (c) Narrow, "D"-, or "bow"-form ridges, with sharply truncated flanks, which in detail seem to be cut by numerous fault structures on all scales are considered to represent advanced stages of AVR dismemberment; (d) Isolated and elevated elongate polygonal blocks, typically occurring off-axis along the southern Reykjanes Ridge, are interpreted as relict AVR fragments which mark a late stage of ridge disaggregation.

3. Faults cutting AVRs identified on TOBI sonographs are often not resolved on Hydrosweep records, and are thus likely to be of the order of 20 m throw or less. We have observed so few unfaulted AVRs on sidescan records that we believe that the tectonic dismemberment of the AVRs is a process that is initiated very soon after, or during, the final stages of AVR growth. In the northern and central portions of the Reykjanes Ridge that we have studied, faulting is so pervasive and the disaggregation process is effected so rapidly, that outside the neovolcanic zone there are few recognisable fragments of AVRs preserved on any of our records.

4. AVRs are initiated at sections of the axial zone which are relatively cool, and thus respond in more brittle fashion to extension at azimuth related to plate divergence. Resultant fissure eruptions build the AVR, warm the segment and thus

transfer the site of potential brittle failure elsewhere. A cyclicity of magmatic and amagmatic crustal extension is thus inferred.

The cyclicity in EVR formation and destruction also provides the necessary process by which the potentially unstable pattern of overlapping segments can survive. Searle & Laughton (1981) noted the multiple overlap of parts of adjacent AVRs, and stated that if the present pattern of spreading elements had persisted for some time, then all of the AVRs could not have been spreading simultaneously. Unless the AVR's continuously reform in some cyclical fashion, they will deform one another where they overlap, either by decapitation or by distributed shear linked by short-offset transform faults. With our more detailed data, however, it is clear that degraded AVRs can be seen off-axis .We propose that the cyclicity may be the

expression of a limited magma supply to the axial-zone and the resultant focusing of magmatic pathways in the brittle-crust. Alternatively, even a robust magma flow to the surface could be focused by Rayleigh-Taylor instability (Whitehead et al. 1984) at the top of the partially molten-zone, and the part of the axial-zone that is fed with magma will thus accommodate plate-separation by crustal generation accompanied by extensional faulting and fracturing. Adjacent, less plentifully supplied areas will allow plate-separation by amagmatic, tectonic extension. Since the horizontal component of melt flow in the crust is the result of differential stress gradients, the build-up of tensional stress in the tectonic portion of the axial-zone will result in magma focussing into these areas and away from the adjacent magmatic portions of the axial-zone. It is this alternating style of magma focussing and defocussing that appears to control the construction and subsequent destruction of the AVR's. AVRs in the northern part of the study area appear generally to be in 'growth' stages compared to those further south. This implies greater magmatic flux and supply in the north relative to the south. This helps to confirm the generation of greater melt volumes closer to the hotspot.

### 4.2 Axial morphology - the effect of the hotspot

Our new data allow us to assess the variations in the processes of ridge volcanism and tectonism on two scales: firstly, those affecting the gross morphology of the Reykjanes Ridge, in particular as a function of its distance from the Icelandic hotspot; and secondly, those bearing upon the constructional components of the ridge, ie, the AVRs - their growth and destruction. The change in the cross-sectional morphology of the axial zone, from a central axial high in the north, to a central axial graben in the south, has been long suggested to be a function of the decreasing

thermal effect with distance from the Icelandic hot-spot (Vogt 1971, Searle & Laughton 1981). We can deduce that at least two processes are involved here:

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1) The higher temperature of the mantle, with increasing proximity to the hot-spot, yields greater degrees of partial melling resulting in greater magma productivity and a greater magmatic crustal thickness. Greater magma supply, however, could result from higher solid-state flow rate with the same degree of partial melling (cf. Ceuleneer et al. Trans Roy. Soc. in press).. This is confirmed by seismic refraction work presented by Talwani et al, 1971). This greater crustal thickness, when in isostatic equilibrium, produces a greater elevation to the seafloor than thinner crust to the southwest.

2) the higher mantle temperature/lower mantle density results in a shallower depth and greater width to the brittle-ductile transition in the north compared with the south. The latter effect is particularly important in controlling the style and amount of axial-parallel faulting. The shallow brittle-ductile transition means that a greater degree of crustal extension could be accommodated by ductile mass transfer in the north as compared to the south (Chen & Morgan, 1990). Furthermore, the higher mantle temperature near the hot-spot results in lower mantle viscosities, which allows a more rapid isostatic equilibration of the upwelling mantle beneath the axial zone forming the central axial high topography, which in turn also determines the spacing of axial parallel faults. In contrast, the cooler, more viscous mantle in the south results in a greater delay in isostatic equilibration beneath the axial zone, resulting in larger throw axial valley faults and the development of an axial graben.

Two principal mean orientations of faults are recognised: those oriented around 014°, approximately perpendicular to the spreading direction, and those around 036°, approximately parallel to the axis, or plate boundary. We are able to identify a systematic variation in the distribution of these faults along strike (Murton, work in progress; - although some authors suggest otherwise - Spencer & McAllister 1991; McAllister & Spencer, 1991; Spencer et al 1991. In Area A, Hydrosweep data indicate almost exclusively faults which are related to the spreading direction, and which range from 012° to 016°. These faults are locally seen to control the boundaries of the AVRs, although the resolution of the Hydrosweep data is insufficient to identify neotectonics associated with the disaggregation of the constructs as described above from TOBI data. East and west of Area A, reconnaissance GLORIA data suggests that continuous axial parallel faulting is observed only outside a central 15 - 20 km axial zone (Searle & Laughton, 1981). In Area B, while only subordinate axial parallel faulting is observed in the axial zone, within 5-10 km of the axis it becomes the dominant tectonic fabric. Fault throws increase away from the axis until reaching a maximum of 300 m, after which they are constant. Faulting sub-orthogonal to the spreading direction is dominant within the axial zone, both bounding (and presumably controlling) the extrusives building the AVRs, as well as fracturing and fragmenting the ridges and volcanoes as they cool and become more rigid. Axial parallel faults generally crosscut, but rarely offset, the spreading-orthogonal fabric, suggesting normal faulting. In Area C, faults subparallel to the axis are common throughout, and are dominant over examples of spreading-orthogonal fabrics.

We interpret this consistent variation as a result of the declining thermal effect of the hotspot on the Reykjanes Ridge at increasing distances from Iceland. The fabrics oriented perpendicular to the spreading direction are present throughout the ridge, to some degree and are interpreted as extensional effects responding to the oblique spreading across the plate boundary. It is noted that the density of faulting within our individual study areas varies along the ridge, and we speculate that the relative presence or absence of the faulting signifies respectively cooler (more brittle) segments, alternating with warmer (less brittle) segments. We have referred above to the causal mechanisms of this medium scale thermal variation in our discussion of evolution of the AVRs. The axis-parallel faulting occurs in the upwelling zone of the plate boundary in response to the uplift and lateral outflow of mantle away from the axis. Where this zone of upwelling and the thermal effects associated with outflow is broad, (such as is the case at fast spreading axes, or those parts of slower-spreading ridges adjacent to hotspots), the effects of plate cooling and subsidence are seen further away from the axis than in cooler, slower-spreading ridges. We have seen that as we move progressively away from effects of the hotspot. the axial parallel faulting becomes more prevalent closer to the axis.

Using the Hydrosweep data we can construct an accurate along-axis bathymetric profile over the 800 km of ridge, and we identify systematic bathymetric segmentation on at least three scales (Figure 7). Firstly, the average depth of the axis, which we refer to as the axial baseline, increases from zero at its subaerial expression at the Reykjanes Peninsular to around 2400 m at 57°N. Secondly, we identify a series of long wavelength bathymetric variations above this baseline, of a frequency of between 30 and 100 km and an amplitude of around 200 m. These perturbations we refer to as "swells". Thirdly we recognise the bathymetric variations associated with the AVRs, which vary about the baseline defined by the swells on which the AVRs are built. These as we have described above vary in spacing between 30 and 5 km and in amplitude from 200 to 500 m. We suggest below that these systematic variations are related to the varying magma supply to the ridge.

As others have suggested from studies of ridge thermodynamics from other mid-ocean ridges, (Whitehead, et al. 1984), we assume that a non-uniform variation in temperature exists along the Reykjanes Ridge. Such variations would produce associated variations in degree of melting and crustal thickness and, if they propagated down the ridge, could explain the diachronous southwesterly-convergent, V-shaped ridges and gravity anomalies observed by Vogt (1971) and Strange (1991). This non-uniform variation may be systematic, and that is one of the parameters which we sought to address during this cruise. Variations between relatively "warm" and "cool" portions of the axis would produce local variations in the response of the plate boundary to the extensional forces associated with the spreading. On this scale, cooler parts of the ridge would tend to behave in a more brittle fashion, and are more likely to be characterised by faulting developed orthogonal to the extension, or spreading direction, ie parallel to  $\sigma_3$ . In contrast, relatively warmer sections of the ridge would behave in a relatively less brittle, fashion; where faulting being controlled largely by the cooling effects at the plate margin as it moved laterally from the axis, and therefore be aligned sub-parallel to the ridge. Ridge parallel faulting is only observed locally within the neovolcanic/ neotectonic zone, and is first recognised consistently in the northeastern Reykjanes Ridge at around 15km from the axis (IOSDL; unpublished GLORIA data). Towards the southwest, however, major, inward-facing faults are observed at the flanks of the median valley.

Finally, we speculate that the bathymetric variations of the AVRs as well as the swells and troughs, which we have interpreted as related to varying effects of magmatic and tectonic activity, are not unique to the Reykjanes Ridge (Nelson, 1981; Lonsdale 1985). However, only on the Reykjanes Ridge can we unequivocally differentiate these processes and their effects due to the oblique nature of the plate separation south west of Iceland. This new model of crustal accretion and segmentation can be readily applied to those preliminary observations made by Sempere et al. (1990) on the ridge-axis discontinuities separating segments of intermediate length (10 - 100 km) along the Mid-Atlantic Ridge between 24°N and 30°N.

### Speculations:

At the stage of analysis of the present data set, with little in the way of groundtruth data to support our interpretations of the morphology and textural information, models explaining the volcanic and tectonic processes at slowspreading plate boundaries must be general. We would like, however, to broaden the discussion of the role of AVRs in mid-ocean ridge constructional volcanism with the following points:

### We speculate that:

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 the variations in the different distribution of axial-parallel and spreading-orthogonal fault geometries is an effect of the distance from the Icelandic hotspot, and the consequent variation in mantle temperature.

2) AVRs are initiated at sections of the axial zone which are relatively cool, and thus respond in more brittle fashion to extension at azimuth related to plate divergence. Resultant fissure eruptions build the AVR, warm the segment and thus transfer the site of potential brittle failure elsewhere. A alternation of magmatic and amagmatic crustal extension is thus produced.

3) The AVR cyclicity, as a result of tectonic focussing of magma supply, is due to a limited and episodic magma flux at a slow spreading ridge. Possible asthenospheric flow along the axis away from the hotspot accentuates the alongstrike thermal gradient and the increased spacing between warm segments of the ridge.

4) This cyclicity is probably a characteristic of construction processes at all M.O.R.s.

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### Figure Captions

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1. Bathymetric chart of the study area, illustrating sites of detailed swath bathymetric surveys (Areas A, B and C) and TOBI survey. 099° spreading direction (B and C arrowed). Half-spreading rate is 10 mm/yr.

2. Hydrosweep swath multibeam data for part of Area A, NE Reykjanes Ridge. Contour interval is 25 m. Plot illustrates shaded, elongated N-S AVR approximately 2.5 km wide at the 800 m isobath and in excess of 10 km in length. Small lobate volcanic constructs ornament the margins.

3a. TOBI deep-tow sidescan sonograph of part of the axial floor in Area B. Light tones correspond to areas of high backscatter, locating freshly erupted volcanics, and/or rough, inclined surfaces. Mottled textures are interpreted as hummocky, pillowed extrusives locally buried by sub-circular volcances, both conical and shield-like. Local sedimented areas are imaged as uniform, low backscatter fields. A range of axis-parallel (035°) and AVR-parallel (014°) faulting is present.

3b. Line interpretation of 3a. Open stipple locates cauliflower texture, grey tones are shield volcanoes and unornamented area is partially sedimented. Random stipple conical volcano at 60° 07.5'N, 29° 17.5'W has partially breached caldera, and rises to shallower than 650 m from valley floor at 950 m.

4. Oblique, three-dimensional view of a mature AVR in central Area B derived from gridded Hydrosweep data. The AVR axis is oriented at 014°, subparallel to evolving, subordinate ridges. Dominant AVR arrowed.

5a. TOBI deep-tow sidescan sonograph montage of the axial valley at the southeastern end of Area C. The highly reflective, and therefore presumably "young" AVR morphology is emphasised by the acoustic shadows defining "haystack"-like mounds with a small number of shield volcanoes. The largest of the flat-topped volcanoes (at 57°50'N, 32°39"W) exhibits a pair of collapse pits or small scale calderas. Within these pits, secondary cones have formed. More uniform, brightly scattered zones with irregular borders flank the ridge, and are interpreted as sheeted lava flows. At the margins of the median valley, a number of inward-facing N-S trending faults cut the darker sedimented valley floor.

5b. Line interpretation of Figure 5a. Key as for 3b, except random dash ornament locates sheet flows.

6. Schematic cartoon of an axial volcanic ridge at an oblique spreading segment in various stages of its life cycle from young (1), adolescent (2), mature (3) and old (4). Note the change from fissure volcances [F] to conical [C] and shield [S] and back to conical types. Also note the initial AVR-parallel faulting, cut by later ridge-parallel faulting as the AVR disintegrates.

7. Simplified bathymetric section derived from ship centre-beam axial profiles. Large scale bathymetric partitioning of the axis is indicated by the kink in 'baseline' depth curve (dotted) at around 60°N. Dashed trace outlines the swell topography defined by nadirs in the axial topographic profile (continuous line). High frequency topographic variations locate the axial volcanic ridges discussed in text.

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Legend to Maps A1 and A2

The eruptive units in this legend are arranged in stratigraphic groups, as defined in Chapter 3, which are presented here in approximate order of age: i.e. the Gljúfur Group is older than the Selfjall Group. However, the exact relative ages between these units are provided in Figure 3.2 (p. 43).



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## MAP B2

This map belongs to WALKER, C.L. 1992. Ph.D. Thesis, University of Durham.

Hveragerdi Village	Bjarnafell <sup>o</sup> cranivostatis <sup>o</sup> cranivostatis <sup></sup>
Map C Volcanic glass samples This map belongs to WALKER, C.L., 1992: Ph.D.Thesis, University of Durham.	Ingolfstjall R R R R R R R R R R R R R



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