Carbon systematics of the Icelandic crust and mantle



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For Grandad Bert

*

nil carborundun desperandum

*

"I have passed through fire and deep water, since we parted. I have forgotten much that I thought I knew, and learned again much that I had forgotten." — *J.R.R.Tolkien*

Declaration

I hereby declare that except where specific reference is made to the work of others, the contents of this dissertation are original and have not been submitted in whole or in part towards any other degree or qualification at this, or any other university. This dissertation is my own work and contains nothing that is the outcome of work done in collaboration with others, except as specified in the text.

My supervisor, John Maclennan, has provided comments on drafts of Chapters 2–6. John Maclennan, Oliver Shorttle, Glenn Gaetani, Véronique Le Roux, and Frieder Klein are co-authors on a submitted manuscript on which Chapter 6 is based, and provided comments on drafts of this manuscript. David Neave has provided comments on drafts of Chapters 3–4.

The total length does not exceed the 275 numbered page limit for the Degree Committee of Earth Sciences, including 225 pages of text, appendices, illustrations, captions and bibliography.

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Abstract

In recent decades there has been an increased interest in the carbon content of Earth's geochemical reservoirs due to the impact of atmospheric carbon on the habitability of our planet. Earth's interior likely hosts a greater mass of carbon than that of the oceans, atmosphere and crust combined, which has buffered the carbon content of the atmosphere over geological time. Yet only a few direct measurements of carbon from the upper mantle, and none from the lower mantle, have been made.

Undegassed basalts erupted at mid-ocean ridges have previously been used to estimate the carbon content of the upper mantle. However, due to the low solubility of carbon within silicate melt, these undegassed basalt suites are rare. The majority of basalts have lost their mantle carbon information en route to eruption through the crust. Various crustal processes act to modify the geochemistry of melts before eruption, therefore it is important to be able to characterise the effect of these processes to better interpret the volatile signals preserved in erupted products.

Pressure, and therefore depth, is a key parameter controlling volatile solubility and can be estimated using a variety of igneous barometers. This thesis presents results from crystallisation experiments conducted on basaltic glass from the Miðfell eruption, Iceland. The experiments provide new data that has been used to test a variety of barometers and crystallisation models used by igneous petrologists, and could aid future barometer recalibration. A key part of this work was the development of an experimental method for stabilising 5 kbar conditions in a piston cylinder apparatus. The experiments have shown that clinopyroxeneliquid barometry is more reliable than multi-reaction barometry. However, knowledge of equilibrium clinopyroxene compositions is crucial for accurately determining pressure using the clinopyroxene-liquid barometer. More experiments conducted at mid-crustal pressures are required for a full recalibration of these barometers.

The results of testing igneous barometers and crystallisation models have been applied to two suites of olivine-hosted melt inclusions from the Kistufell and Miðfell eruptions to help determine the melt evolution history of these basalts. These eruptions were targeted due to previously measured noble gas isotopic ratios that suggest a primordial mantle component present in their melting regions, and therefore evoking the possibility that they could hold information about deep mantle carbon. Barometry suggests that Miðfell phases equilibrated, and therefore crystallised, at mid-crustal pressures (5–7 kbar), which could allow for the entrapment of undegassed melt inclusions within olivine.

The two melt inclusion suites were found to differ in trace element variability, with the observation that the more trace element enriched eruption, Kistufell, had lower relative trace element variability than the more depleted eruption, Miðfell. Several processes, both in the crust and the mantle, are likely responsible for the level of trace element enrichment and variability, including extent of mantle melting, source heterogeneity, and melt transport.

The depleted nature of the Miðfell melt inclusions has allowed them to preserve some of the highest CO₂/Ba and CO₂/Nb ratios ever recorded in basaltic glass, with ratios over five times greater than undegassed mid-ocean ridge basalt values. This carbon enrichment is not due to any crustal melt modification process, but rather pertaining to lower mantle carbon-rich lithologies that have been tapped by the Icelandic mantle plume. The carbon reservoir beneath Miðfell is estimated to contain 744 \pm 188 ppm carbon, 15 times greater than the depleted upper mantle. This value matches estimates of bulk mantle carbon from planetary mass balance calculations and provides evidence for carbon-rich domains within the Earth.

Table of contents

Li	List of figures xvii					
Li	List of tables xx					
1	Intro	troduction				
	1.1	Aims a	und objectives	1		
	1.2	Thesis	structure	3		
	1.3	Geolog	gical setting of Iceland	4		
		1.3.1	Plume-ridge interaction	4		
		1.3.2	Rift zones and volcanic systems	4		
	1.4	Magma	atic processes	6		
		1.4.1	Melt formation and mixing within the mantle	6		
		1.4.2	Melt evolution and modification in the crust	8		
		1.4.3	Melt inclusion entrapment and interpretation	9		
		1.4.4	Constraints from igneous barometry	10		
	1.5	Deep E	Earth carbon reservoirs	11		
2	Sam	ples and	d methods	13		
	2.1	Introdu	uction	13		
	2.2	The Ki	stufell eruption	14		
		2.2.1	Eruption overview	14		
		2.2.2	Sample collection	16		
	2.3	The M	iðfell eruption	16		
		2.3.1	Eruption overview	16		
		2.3.2	Sample collection	18		
	2.4	Analyt	ical methods	18		
		2.4.1	Sample preparation	18		
		2.4.2	Secondary ion mass spectroscopy	20		

		2.4.3	Electron probe microanalysis
		2.4.4	Uncertainty analysis
		2.4.5	Scanning electron microscopy
		2.4.6	Vapour bubble-melt inclusion CO ₂ reconstructions
	2.5	Data c	ollation
		2.5.1	Database of Icelandic melt compositions
		2.5.2	Database of global melt inclusion compositions
3	Exp	eriment	al methods 33
	3.1	Introdu	action
	3.2	Startin	g material
		3.2.1	Preparation
	3.3	Experi	ments at 0.001 kbar
		3.3.1	Sample preparation
		3.3.2	Experimental set-up 36
		3.3.3	Experimental procedure
	3.4	Experi	ments at 10 kbar
		3.4.1	Assemblage preparation
		3.4.2	Experimental procedure
		3.4.3	Pressure and temperature uncertainty
	3.5	Experi	ments at 5 kbar
		3.5.1	Assemblage preparation
		3.5.2	Experimental procedure development
		3.5.3	Final experimental procedure
		3.5.4	Pressure and temperature uncertainty
		3.5.5	Na-gain to the starting bulk composition
		3.5.6	Further refinement to the experimental procedure
4	Exp	eriment	al petrology: Miðfell eruption 49
	4.1	Introdu	uction
	4.2	Petrolo	9gy
		4.2.1	Run products
		4.2.2	Glass compositions
		4.2.3	Mineral compositions
		4.2.4	Liquid-mineral modal proportions
		4.2.5	Partitioning of major elements between minerals and melt 71
		4.2.6	Clinopyroxene compositional variability

	4.3	Crysta	llisation models	84
		4.3.1	Petrolog	85
		4.3.2	Plagioclase-melt equilibrium models	88
		4.3.3	MELTS	89
	4.4	Igneou	s barometry	92
		4.4.1	Clinopyroxene-liquid barometry	92
		4.4.2	Olivine-plagioclase-augite-melt (OPAM) barometry	96
		4.4.3	Multi-reaction barometry	100
		4.4.4	Summary of igneous barometers	104
	4.5	Applic	cation to natural Miðfell samples	105
		4.5.1	Thermometry	105
		4.5.2	OPAM barometry	106
		4.5.3	Clinopyroxene-liquid barometry	109
		4.5.4	Multi-reaction barometry	109
	4.6	Conclu	usions	110
5	Mali	tinclus	ion goochomical hotorogonaity	113
5	5 1	Introdu	uction	113
	5.1	Petrol	action	115
	5.2	521	Kistufell	117
		5.2.1	Miðfell	117
		5.2.3	Whole rock matrix glass, and melt inclusion geochemistry	119
	5.3	Comp	ositional heterogeneity of olivine-hosted melt inclusions	121
		5.3.1	Significance of melt inclusion compositional variability	122
		5.3.2	Principal component analysis	125
		5.3.3	Correcting major element oxides for post-entrapment crystallisation	126
	5.4	Melt n	nodification in the crust	128
		5.4.1	Diffusive re-equilibration of water (H_2O)	130
		5.4.2	Volatile degassing	132
		5.4.3	Concurrent melt mixing and crystallisation	135
		5.4.4	Plagioclase assimilation	137
		5.4.5	Gabbro assimilation	142
	5.5	Melt v	ariability in the mantle	144
		5.5.1	Major element enrichment	145
		5.5.2	Trace element variability	148
		5.5.3	Isotope signatures	151
	5.6	Comp	ositional enrichment and variability	152

	5.7	Conclu	usions
6	Dee	p mantl	e carbon 161
	6.1	Introdu	uction
	6.2	Carbo	n in the solid Earth
		6.2.1	CO_2 in MORB glasses
		6.2.2	CO ₂ in Kistufell and Miðfell melt inclusions
		6.2.3	Global melt inclusion suite
		6.2.4	CO ₂ solubility and degassing
	6.3	Miðfel	ll carbon enrichment
	6.4	Carbo	n enrichment from crustal processes
		6.4.1	Simple plagioclase addition
		6.4.2	Interaction with gabbroic material
	6.5	Carbo	n enrichment from mantle processes
		6.5.1	Carbonatite melt addition
		6.5.2	Carbon liberation from the mantle
	6.6	Carbo	n enrichment in source mantle lithologies
		6.6.1	Recycled oceanic crust
		6.6.2	Primordial mantle reservoir
	6.7	Earth's	s carbon reservoirs
		6.7.1	Carbon content of Miðfell mantle source lithology
		6.7.2	Implications of Miðfell carbon enrichment
	6.8	Compa	aring DMM and lower mantle carbon
		6.8.1	Early reservoir differentiation
		6.8.2	Trace element-carbon recycling
		6.8.3	Melt depletion processes
	6.9	Conclu	usions
7	Sum	mary	195
	7.1	Experi	mental petrology and igneous barometry
	7.2	Geoch	emical heterogeneity within olivine-hosted melt inclusions 197
	7.3	Deep r	nantle carbon
	7.4	Furthe	r work
		7.4.1	Igneous barometry
		7.4.2	Melt inclusion heterogeneity
		7.4.3	Mantle volatile reservoirs

References		201
Appendix A	Data corrections and standards	221
Appendix B	Dataset references	223
Appendix C	Extended data tables	225

List of figures

1.1	Map of Iceland	5
1.2	Magmatic system schematic	7
2.1	Kistufell sample locality map	15
2.2	Miðfell sample locality map	17
2.3	SIMS CO ₂ calibration curves	21
2.4	Secondary standard trace element yields	23
2.5	SIMS measurement comparison between Edinburgh IMF and WHOI	25
2.6	Relative error dependency on measured trace element concentration	29
2.7	Micro-Raman spectrum of a CO_2 vapour bubble	31
3.1	Deltech apparatus set-up for 0.001 kbar experiments	37
3.2	CCO oxygen fugacity buffer for experiments	38
3.3	Graphite capsule assemblage for piston-cylinder experiments	40
3.4	Piston-cylinder apparatus set-up for 5 and 10 kbar experiments	41
3.5	NaCl in 5 kbar experimental charges	48
4.1	SEM images of 0.001 kbar experiments	52
4.2	SEM and QEMSCAN images of 5 kbar experiments	54
4.3	SEM and QEMSCAN images of 10 kbar experiments	56
4.4	Experimental PT phase diagram for Miðfell basaltic glass	59
4.5	Harker plots for experimental glass major element oxides	64
4.6	Experimental glass compositions against equilibration temperature	65
4.7	Compositions of experimental silicate phases	67
4.8	Calculated modal proportions of experimental phases	70
4.9	Mineral-liquid partition coefficients	73
4.10	Cation fractions of experimental clinopyroxenes	75
4.11	Experimental clinopyroxene cation variablity	77
4.12	Experimental clinopyroxene end-member fractions	78

4.13	0.001 kbar cpx compositions, thermometry, and cpx-liquid partition coefficients	80
4.14	10 kbar cpx compositions, thermometry, and cpx-liquid partition coefficients	82
4.15	Comparison of experimental and Petrolog-predicted mineralogy	86
4.16	Plagioclase-melt equilibrium model and experimental plagioclase compositions	88
4.17	Comparison of experimental and MELTS-predicted mineralogy	91
4.18	Clinopyroxene-liquid barometry	94
4.19	Comparison of different OPAM barometer parameterisations	98
4.20	OPAM kernel density function plots	99
4.21	Multi-reaction barometry	102
4.22	Pressure estimates from natural Miðfell compositions	108
5.1	Photomicrographs of melt inclusions from Kistufell and Miðfell	116
5.2	Miðfell hand specimen and QEMSCAN image	118
5.3	Photomicrographs of Miðfell whole rock and gabbro xenoliths	120
5.4	TAS plot of Kistufell and Miðfell samples	121
5.5	Post-entrapment crystallisation correction of major element oxides	129
5.6	H ₂ O and CO ₂ systematics of Kistufell and Miðfell	131
5.7	Halogen content of Kistufell and Miðfell	133
5.8	Miðfell melt inclusion correlation matrix	135
5.9	Trace element ratio variability with host forsterite content	137
5.10	Trace element spider diagrams for Miðfell	139
5.11	Plagioclase addition major and trace element anomalies	140
5.12	Plagioclase addition trace element model	141
5.13	Gabbroic partial melt trace element mixing model	144
5.14	Harker plots for Kistufell, Miðfell, and Icelandic melt inclusions	146
5.15	Nb/Zr enrichment in Kistufell, Miðfell, and Icelandic melt inclusions	148
5.16	Principal component loadings for observed variables	149
5.17	LREE-HREE fractionation during garnet zone melting	150
5.18	Sr-Nd-Pb isotopes for Kistufell, Miðfell, and Icelandic glasses	151
5.19	Melt inclusion La/Yb variability with host olivine forsterite	153
5.20	Observed melt inclusion trace element enrichment and variability	155
5.21	Primary mantle melt variability model	158
6.1	CO ₂ -Nb and -Ba systematics of Kistufell and Miðfell melt inclusions	167
6.2	CO_2/Ba variability compared to enrichment and host olivine composition $\ .$	168
6.3	Trace element patterns for Kistufell and Miðfell melt inclusions	169
6.4	Global melt inclusion and MORB glass CO ₂ -Nb and CO ₂ -Ba systematics .	170

6.5	The effect of CO_2 degassing on CO_2 /Ba ratios	173
6.6	Undegassed glass CO_2 -Nb and CO_2 -Ba systematics $\ldots \ldots \ldots \ldots$	175
6.7	Enrichment of Miðfell CO ₂ /ITE ratios with respect to MORB ratios \ldots	177
6.8	Modelled carbonatite melt mixing with primitive Miðfell melts	180
6.9	Potential lower mantle lithologies responsible for high carbon signal	184

List of tables

2.1	Precision and accuracy for melt inclusion EPMA and SIMS measurements .	27
2.2	Precision and yields for experimental EPMA measurements	28
3.1	Experimental bulk starting composition	35
4.1	Experimental conditions, modal mineralogy, and mineral compositions	61
4.2	Multi-reaction barometry pressure estimates	103
4.3	Thermobarometry on natural Miðfell samples	107
5.1	Major, trace, and volatile element variability in Kistufell melt inclusions	123
5.2	Major, trace, and volatile element variability in Miðfell melt inclusions	124
5.3	Primary mantle melt La variability statistics	154
6.1	CO ₂ /ITE ratios of undegassed melt inclusions and MORB glasses	165
6.2	CO ₂ /ITE ratios from Miðfell olivine-hosted melt inclusions	167
A.1	Secondary standard corrections	221
A.2	Miðfell relative error equation parameters	222
B .1	Global melt inclusion dataset references	223
C.1	Kistufell melt inclusion volatile and trace elements	226
C.2	Kistufell melt inclusion major and volatile elements	230
C.3	Kistufell olivine major elements	232
C.4	Miðfell melt inclusion volatile and trace elements	234
C.5	Miðfell melt inclusion major and volatile elements	237
C.6	Miðfell olivine major elements	239
C.7	Melt inclusion SIMS CO ₂ standards	240
C.8	Melt inclusion SIMS volatile and trace standards	241
C.9	Melt inclusion EPAM standards	242

C.10 WHOI Miðfell melt inclusion volatile and trace elements $\ldots \ldots \ldots$	243
C.11 Experimental glass major elements	244
C.12 Experimental olivine major elements	248
C.13 Experimental plagioclase major elements	252
C.14 Experimental clinopyroxene major elements	254
C.15 Experimental EPMA standards	257

Chapter 1

Introduction

1.1 Aims and objectives

The vast majority of Earth's volcanic activity occurs at mid-ocean ridges (MORs). An understanding of the magmatic processes occurring along these constructive plate boundaries is central to our knowledge of oceanic crust formation, global magmatism, and fluxes between Earth's geochemical reservoirs (Klein and Langmuir, 1987; Maclennan et al., 2001b; Dasgupta and Hirschmann, 2010; Kelemen and Manning, 2015; Mason et al., 2017). The generation of basaltic melt lies at the heart of the MOR system; moving lithophile and volatile elements from the mantle into the oceanic crust, oceans and atmosphere (Langmuir et al., 1992; Shorttle et al., 2014).

Beneath MORs, passive upwelling of asthenospheric mantle and/or active convection of a mantle plume drives the movement of mantle material to lower pressures. Upon decompression, mantle components cross their respective solidi and partially melt. Melting is initiated along grain boundaries in the mantle mineral assemblage, increasing in melt volume until an interconnected network of melt forms (Zhu et al., 2011). Melt is then able to migrate through the mantle peridotite solid residue, driven by buoyancy, up to shallower depths. The separation of melt from its solid residue fractionates the two compositions, allowing for multiple melts to be formed from a single bulk composition by progressive fractional melting (Langmuir et al., 1992; Plank and Langmuir, 1992).

At shallower depths in the mantle, melts originating from depth are thought to react with the surrounding peridotite residue to form melt channels that focus and mix melt together (Kelemen et al., 1995; Katz and Weatherley, 2012; Weatherley and Katz, 2012). Once melt has crossed the MOHO and entered the crust, it can pond in magma chambers at depths of neutral buoyancy.

Heat loss from magma to surrounding rock drives a number of crustal processes that evolve and modify the geochemistry of melts from the mantle. As magma temperature falls, mineral phases arrive on the liquidus and crystallise from the melt, which evolves to more silica-rich compositions as a result (Bowen, 1915, 1928; Langmuir, 1989; Wanless and Shaw, 2012; Shorttle, 2015). Melt can become trapped in crystallising phases, preserving the composition of the parental liquid at the time of crystallisation (Roedder, 1984; Lowenstern, 1995). Dynamic convection currents can be set up within magma chambers, again driven by heat loss, which can mix melts and homogenise their geochemistry (Maclennan, 2008a). As the country rock heats up, magma can assimilate crustal material to further modify its composition (Nicholson et al., 1991; Gurenko and Sobolev, 2006; Eason and Sinton, 2009; Brounce et al., 2012).

Melt migration to shallower depths within the magmatic system causes the magma to decompress, forcing the solubility of volatiles to decrease and gas exsolution to occur (Dixon et al., 1991; Shishkina et al., 2014). A continued decrease in temperature, increase in crystallinity, and lower volatile content changes the mechanical properties of the melt as it ascends towards eruption. These mechanical properties, along with the surface eruption environment determine the type of erupted material produced. In Iceland, basaltic lava can interact with ice, water, and air to give three very different styles of volcanic eruption and associated suites of rock (Thordarson and Larsen, 2007).

The overarching aim of this thesis is to explore the magmatic processes responsible for the diverse geochemistry of erupted products found in Iceland. This theme can be split into three principal questions:

- 1. What controls the extent of geochemical variability preserved within olivine-hosted melt inclusions?
- 2. At what pressures do melts equilibrate and crystallise phases within the Icelandic crust?
- 3. What information can olivine-hosted melt inclusions yield about the carbon content of the Icelandic mantle?

The research approach for this thesis is in two parts: (i) experiments conducted on basaltic glass from Miðfell, Iceland are used to test several igneous barometers and crystallisation models, and (ii) olivine-hosted melt inclusion suites from Kistufell and Miðfell, Iceland are analysed to investigate the control of crustal and mantle processes on compositional heterogeneity, with a focus on carbon systematics.

1.2 Thesis structure

- **Chapter 2:** This chapter describes the analytical methods used to collect data from olivinehosted melt inclusions and experimental charges. It includes an explanation of the analytical uncertainties on these measurements, along with details of corrections applied to the datasets. Descriptions of the Kistufell and Miðfell eruptions are given, along with information about sample collection and collation of previously published data.
- **Chapter 3:** Experimental procedures carried out to create the charges discussed in Chapter 4 are detailed here. This description includes sample preparation, experimental apparatus, and experimental procedures, with discussion of the methodology development required to conduct piston-cylinder experiments at 5 kbar.
- **Chapter 4:** This chapter outlines the various igneous models used by petrologists to reconstruct crystallisation history and predict equilibration pressures of solid-melt assemblages. Experiments on Miðfell basaltic glass are used to test the ability of these models to predict equilibration conditions, such as temperature and pressure, and the evolution of igneous phases as melt cooling and crystallisation takes place. The tested igneous barometers are then applied to natural samples from Miðfell to provide pressure estimates for phase assemblage equilibration in the magmatic system.
- **Chapter 5:** In this chapter the compositions of olivine-hosted melt inclusions from Kistufell and Miðfell are used to provide an assessment of the geochemical heterogeneity present in the two eruptions. Discussion of this heterogeneity explores the controls on trace element variability within olivine-hosted melt inclusions. Comparison to wider Icelandic melt inclusion suites investigates the link between variability and enrichment.
- **Chapter 6:** The carbon content of Miðfell melt inclusions is discussed in detail in this chapter. It describes observed CO₂/incompatible trace element ratios, discusses processes that could have modified Miðfell carbon content, and presents a carbon estimate for an undegassed primordial reservoir in the lower mantle.
- **Chapter 7:** This chapter provides a summary of the principal findings of this thesis and identifies potential further work.

1.3 Geological setting of Iceland

1.3.1 Plume-ridge interaction

Iceland is the subaerially exposed part of a basaltic plateau situated at the junction between the Mid-Atlantic Ridge (\sim N-S) and the Greenland-Iceland-Faeroe Ridge (\sim E-W), over a mantle plume (Schilling, 1973; Vink, 1984; White and McKenzie, 1995). The island rises \sim 1000 m above sea level at its centre, with a crustal thickness up to 40 km, and is centred on a 50–200 km wide shelf (Darbyshire et al., 2000). Just under a third of the total erupted area (350000 km²) is above sea level (Thordarson and Larsen, 2007). The large melt volumes produced here are the result of interaction between a mantle plume and a spreading plate boundary, with the North American and Eurasian plates moving apart at 20 mm/yr (DeMets et al., 1994).

Iceland is situated at the centre of a large wavelength negative gravity anomaly, which matches the size of the bathymetry profile surrounding Iceland and extent of thickened oceanic crust (Jones et al., 2002). Observations of thickened crust, mantle seismic tomography, and the centre of the gravity anomaly suggest that the centre of the Icelandic plume lies under the southern extent of Vatnajökull glacier. High degrees of melting are required to produce such thick crust (White et al., 1992). A thermally buoyant mantle plume, which causes melting by active upwelling of mantle material is the best interpretation for the observations made above (Maclennan et al., 2001a; Putirka, 2008a). Further evidence for a plume beneath Iceland comes from the geochemistry of erupted products in the form of U-series disequilibria, trace element compositions, and noble gas isotopic ratios (Maclennan et al., 2001a; Peate et al., 2001; Kokfelt et al., 2003; Mukhopadhyay, 2012). Tomographic studies of mantle beneath Iceland show a region of seismic wave speed anomaly going down through the transition zone to the lower mantle, and potentially rooted above a large low shear-wave velocity province (LLSVP; Rickers et al. 2013; Jenkins et al. 2016; Yuan and Romanowicz 2017).

The Icelandic mantle plume has been active for ~ 65 Myrs, and was responsible for the opening of the North Atlantic ocean and formation of the North Atlantic Igneous Province (White and McKenzie, 1989).

1.3.2 Rift zones and volcanic systems

Plate spreading is focused along three rift zones in Iceland (Figure 1.1). To the North, the Kolbeinsey spreading ridge comes onshore to connect up with the Northern Volcanic zone (NVZ), which extends down through Iceland forming the Highlands down to the Vatnajökull



Figure 1.1 Map of Iceland showing the three rift zones (orange) where plate spreading is localised, central volcanoes (red) where volcanism is localised within the rift zones, the Kistufell eruption (purple), and the Miðfell eruption (burgundy). The largest glacier (white) is Vatnajökull, beneath which the locus of the Icelandic mantle plume is thought to be located.

glacier. Underneath the ice-cap, extension is partitioned between the longer-lived Western Volcanic Zone (WVZ) and the younger Eastern Volcanic Zone (EVZ), which is currently propagating southward with the formation of the Vestmannæyar. The WVZ continues offshore as the Reykjanes spreading ridge. Magmatism in the NVZ and WVZ is mainly characterised by tholeiitic lavas as the two rift zones are well-established (Gudmundsson, 2000). In the EVZ, basalts transition from tholeiitic compositions in the North to alkali basalts in the South at the propagating tip (Jakobsson, 1979; Thordarson and Höskuldsson, 2008).

Volcanism within the rift zones is focused onto discrete rift segments, forming fissure swarms and central volcanoes (Thordarson and Larsen, 2007; Jóhannesson and Sæmundsson, 2009). Fissure swarms are orientated perpendicular to spreading, with intrusion after intrusion accommodating extension within the rift. Central volcanoes are longer-lived, therefore they

are associated with shallow crustal magma chambers and the evolution of silicic magma compositions due to sustained crustal residence (Jónasson, 1994; Gurenko et al., 2015).

Proposed mechanisms for magma supply to eruptions in Iceland are split into two key hypotheses: lateral flow and magma reservoirs (Gudmundsson, 1987; Hartley and Thordarson, 2013). Beneath central volcanoes there is seismic tomography evidence using wave velocity ratios, V_p/V_s , and ground deformation evidence for shallow magma reservoirs (Reverso et al., 2014). Microseismicity has been used to detect melt lenses beneath the Askja caldera in central Iceland (Greenfield et al., 2016). It is less clear whether during fissure eruptions all erupted magma proceeds via a central magma chamber and flows out laterally to the propagating fissure, or if it is fed directly from beneath. During the 2014 Bárðarbunga eruption, earthquake swarms captured the lateral propagation of a dyke away from the Bárðarbunga central caldera toward the eventual fissure eruption site NE of the volcano (Ágústsdóttir et al., 2016; Hudson et al., 2017).

1.4 Magmatic processes

Observations and interpretation of the processes occurring within Iceland's magmatic systems are required to address the aims of this thesis (Figure 1.2). These processes create and modify melts, which are ultimately homogenised to give the eruptive material seen at the surface. However, a snapshot of compositional heterogeneity can be preserved within olivine-hosted melt inclusions (Sobolev and Shimizu, 1993; Gurenko and Chaussidon, 1995; Maclennan et al., 2003b; Laubier et al., 2007; Hartley et al., 2017). They pertain to melt compositions that have long since obscured the crustal processing of carrier melts, and allow the interpretation of deep crustal and mantle processes from the geochemical variability they contain (Shorttle, 2015; Jennings et al., 2017). To gain insight into the formation of Icelandic crust, it is important to disentangle the various processes that have shaped the geochemistry of erupted rocks, and highlight the extent to which they have been significant in crustal formation.

1.4.1 Melt formation and mixing within the mantle

Source heterogeneity in the mantle beneath Iceland has long been thought a necessity based on the observations of geochemical tracers, such as isotope ratios, trace element concentrations and ratios (Sobolev et al., 2007; Shorttle and Maclennan, 2011; Shorttle et al., 2014). Noble gas isotopes suggest the presence of an undegassed primordial reservoir within the Icelandic source region (Harrison et al., 1999; Dixon et al., 2000; Brandon et al., 2007; Füri et al., 2010; Mukhopadhyay, 2012).



Figure 1.2 A schematic diagram of the magmatic system beneath Iceland, highlighting the main mantle and crustal processes that contribute to melt formation and heterogeneity.

Radiogenic isotopic systems suggest that different fractionated and long-lived reservoirs exist in the Icelandic mantle (Hofmann and White, 1982; McKenzie et al., 2004; Thirlwall et al., 2004). Trace elements and ratios, such as Nb/Zr, have been used to suggest that enriched compositions are the result of enriched source components, such as pyroxenite or another recycled lithology (Peate et al., 2010; Shorttle et al., 2014).

However, there is a clear trade-off to be had between source enrichment and extent of melting, as expressed by the fractional melting equation:

$$C_i = \frac{C_0}{D} (1 - F)^{\frac{1}{D} - 1},$$

where high melt concentrations, C_i , can be achieved by either high initial source concentrations, C_0 , i.e. an enriched source component, or low extents of melting, F. Dihedral angles between melt and solid mineral phases are such that melt is interconnected, meaning that flow by Darcy's law is inevitable decoupling the melt from the solid residue (Buck and Su, 1989; Holness, 2006). Therefore, a fractional process is often used to best describe melt formation beneath a MOR. Continuous melting is a more realistic description of the mantle melting process, with < 1% melt able to stay in equilibrium with the solid residue (Langmuir et al., 1992; McKenzie, 2000; Peate and Hawkesworth, 2005). The pure fractional melting end-member would produce the most diverse range of primary mantle melts, but continuous melting can be viewed as pseudo-fractional, and therefore will also produce heterogeneous melts.

It is clear that primary mantle melt heterogeneity is due to both source variability and fractional melting. However, the migration of melt along grain boundaries, through interconnected networks in the solid residue and into melt channels has the potential to homogenise initial melt compositions, eliminating variability (Kelemen et al., 1997a; Weatherley and Katz, 2012). Therefore, the compositional variability entering the crust is but a fraction of the potential total heterogeneity created in the mantle (Rudge et al., 2013).

1.4.2 Melt evolution and modification in the crust

Melts entering the crust have a long way to travel before the heterogeneity they inherited from the mantle can be expressed at the surface. Melt evolution through crustal processes makes it unlikely that any of this heterogeneity will survive. The crustal magmatic system is a dynamic environment, driven by the loss of heat to surrounding rock, and the loss of mass in the form of crystallised products (Maclennan, 2008a; Shorttle et al., 2016).

Crustal processes include: (i) fractional crystallisation, which drives changes in melt density, and therefore the continued migration of magma through the crust. (ii) Crustal assimilation and interaction with crustal material, which imprints different geochemical signals onto those already expressed by mantle melts (Gurenko and Sobolev, 2006; Eason and Sinton, 2009). (iii) Melt mixing, whether within melt channels or the more vigorous melt mixing environment of a magma chamber, causes melts to be homogenised towards an average composition and the loss of variability (Maclennan, 2008a; Neave et al., 2014). (iv) Degassing, at the shallowest depths in the magmatic system, volatile elements become less soluble in silicate melt, causing them to exsolve from the melt and escape the system (Dixon et al., 1991; Hartley et al., 2014). These processes make it difficult to retain any of the primary mantle melt information initially supplied to the crustal magmatic system.

1.4.3 Melt inclusion entrapment and interpretation

Although most erupted lavas have lost information about the primary mantle melt variability that formed them, it is possible to identify some of the heterogeneity present deeper within the magmatic system through the measurement of olivine-hosted melt inclusions (Saal et al., 2002; Maclennan et al., 2003b; Neave et al., 2013; Schipper et al., 2016).

Olivine is the first phase to crystallise from primary mantle melts, therefore it is exposed to the greatest melt diversity of all the silicate phases crystallised at MORs. During growth, olivine can trap pockets of melt within the mineral structure, often due to skeletal growth from undercooling or non-uniform supply of elements to growth faces, though some can be the result of olivine-melt resorption (Lowenstern, 1995). Once the melt inclusion has been sealed off from the parental magma, then the composition can no longer be modified by bulk magma evolution. It is assumed that the melt inclusion and olivine are initially in equilibrium with each other (Danyushevsky et al., 2002). Therefore, olivine-hosted melt inclusions can preserve the geochemical variability present deep within the magmatic system, but not the primary mantle melt heterogeneity.

However, several processes can modify the melt inclusion after entrapment, obscuring the initial trapped melt composition. All these modification processes occur as the olivine cools and decompresses in the magmatic system: (i) post-entrapment crystallisation (PEC) forms olivine on the walls of the inclusion, changing the chemistry of the melt by enriching the concentration of incompatible elements (Danyushevsky et al., 2002). If cooling is slow enough, then the melt inclusion will not quench to a homogeneous glass, but rather a heterogeneous microcrystalline assemblage. (ii) Vapour bubble formation occurs when CO_2 exsolves from the melt inclusion as a result of cooling, decompression and PEC. These three processes either decrease CO_2 solubility or increase the dissolved concentration until CO_2 saturates in the melt, causing it to exsolve (Lowenstern, 1995; Kamenetsky et al., 2002). (iii) Olivine decrepitation is where olivine fractures due to a pressure difference between the melt inclusion and the host magma, this can lead to the loss of any exsolved vapour bubbles and dissolved CO_2 from the sudden decompression (Maclennan, 2017). (iv) Diffusive H₂O re-equilibration resets the H₂O content of melt inclusions by diffusion of H⁺ and OH⁻ through the olivine lattice (Gaetani et al., 2012; Bucholz et al., 2013).

Careful interpretation of olivine-hosted melt inclusions is required to make sure the processes detailed above do not influence the overall interpretation of melt inclusion data (Moore, 2008). PEC can be corrected using a number of models based on the assumption that a melt inclusion and its host olivine were in equilibrium at the time of entrapment (Danyushevsky and Plechov, 2011). H₂O concentrations cannot be recovered, unless assumptions are made regarding the initial H₂O/ITE ratio of the melt, e.g. H₂O/Ce. The CO₂ content of vapour bubbles can be recombined with that of the melt inclusion using X-ray tomography and Raman spectroscopy, to give the initial entrapment CO₂ content. However, olivine decrepitation causes the irretrievable loss of CO₂ information from the melt inclusion. High CO₂ content melt inclusions are most likely to decrepitate and lose vapour bubbles. Olivines that have undergone decrepitation often anneal their fractures, so it is not obvious that decrepitation has occurred until measurement of the melt inclusion (Maclennan, 2017).

1.4.4 Constraints from igneous barometry

The role of igneous barometry in the interpretation of magmatic processes is to provide equilibrium pressure estimates for igneous phase assemblages, which can include melt. These pressures are generally inferred to represent the pressure of magma chambers within the crust (Maclennan et al., 2001b). It is hoped that with better knowledge of magma chamber depths within active volcanic systems, surface observations such as ground deformation, degassing and seismicity can be linked to petrological observations of magma reservoir depth (Edmonds, 2008; Tarasewicz et al., 2014; Neave and Putirka, 2017). This link to magma chamber depth has the potential to aid volcanic eruption forecasting by predicting the style of eruption and the hazards associated with it.

Another motive for igneous barometry is to compare geophysical observations of melt lenses and melt migration seismicity to pressure estimates from erupted products (Greenfield et al., 2016; Hudson et al., 2017). This link would help match crustal seismic structure to igneous mineral assemblages by estimating the depth at which erupted assemblage crystallised. More detailed information about magma crystallisation depth would help the further development of crustal accretion models (Maclennan et al., 2001b).

Unlike metamorphic systems where pressure-temperature conditions can be estimated using numerous P- and T-sensitive reactions between mineral end-members, igneous barometry is restricted to a relatively small number of igneous phases, including melt (Powell and Holland, 2008). Therefore, common igneous barometers have to make use of the pressure sensitivity of: (i) volatile solubility in silicate melt. CO_2 is the main volatile of interest in Iceland for barometry, however H₂O has greater importance in more hydrous magmatic settings (Dixon et al., 1991; Saal et al., 2002; Shishkina et al., 2014). Melt inclusion CO_2 content is assumed to be at saturation, if it was not at the time of entrapment then estimated pressure is a minimum. (ii) Liquid-mineral equilibria. Examples include the pressure sensitivity of jadeite in clinopyroxene, which is in equilibrium with melt, and the olivine-plagioclase-augite-melt (OPAM) barometer, which requires the measurement of a glass composition that was in equilibrium with all three OPA phases (Yang et al., 1996; Neave and Putirka, 2017; Voigt et al., 2017). (iii) Solid phase equilibria. Recent developments to activity-composition (*a-x*) models have enabled multi-reaction average pressure barometry on mafic igneous compositions (Green et al., 2016; Ziberna et al., 2017).

A key component to all these barometers is the experimental compositions used for calibration. Several studies have experimented on basaltic compositions and experimental techniques have improved over the past few decades, but very few experiments have been calculated at pressures of fundamental interest to igneous petrologists; crustal pressures (Moore et al., 2008). More experimental assemblages are required at crustal pressures to improve current barometer calibrations and to aid the development of new ones (Neave and Putirka, 2017).

1.5 Deep Earth carbon reservoirs

Earth's solid interior likely hosts more carbon than the crust, atmosphere and oceans combined, making it a significant reservoir within the carbon cycle (Des Marais, 2001; Dasgupta and Hirschmann, 2010). The carbon content of the atmosphere is of key interest to mankind, due to its impact upon planet habitability. Humanity has modified atmospheric CO_2 concentration over the last two centuries through the burning of fossil fuels, which has moved carbon from crustal storage to the atmosphere. However, on geological timescales it is the interaction between solid Earth carbon reservoirs through degassing at MORs and recycling at subduction zones that have maintained a surface environment capable of hosting life (Hayes and Waldbauer, 2006; Hirschmann and Dasgupta, 2009).

The carbon content and fluxes between key surficial reservoirs are reasonably well known as they can easily be observed directly (Des Marais, 2001). Climate scientists are keen to understand how these reservoirs can be enhanced to draw down more CO_2 from the atmosphere and alleviate the enhanced greenhouse effect. However, Earth's interior is more enigmatic, with inferences about deep Earth carbon content coming from global

mass balance arguments (Marty, 2012; Hirschmann, 2016). These calculations cannot shed light on the potential heterogeneous distribution of carbon within the Earth's mantle. Solid surface reservoirs, excluding the atmosphere, are observed to be chemically diverse, including carbon distribution, therefore it can be expected that deep solid Earth reservoirs are also heterogeneous in carbon. This supposed heterogeneity is supported by other geochemical tracers and the observation that subducted material transporting carbon to the mantle is chemically diverse (Shorttle et al., 2014; Kelemen and Manning, 2015; Clift, 2017). It is logical to believe that some of this heterogeneity will be maintained into the deep mantle, though some subduction processes may act to modify this chemical signature (Foley et al., 2000; Manning, 2004; Kessel et al., 2005).

Direct observations of mantle carbon have been made at MORs by analysing undegassed MORB suites, which exist mainly in the form of olivine-hosted melt inclusions (Saal et al., 2002; Michael and Graham, 2015; Hauri et al., 2017; Le Voyer et al., 2017). Measurements of coupled lithophile and volatile elements are used to reconstruct the carbon content of the source mantle. Melting processes fractionate mantle material, therefore lithophile elements used to calculate mantle carbon must appear to have similar compatibility during mantle melting as carbon (Le Voyer et al., 2017). CO₂/Nb and CO₂/Ba ratios are thought to be invariant to the mantle melting process, though they can be fractionated at very low melt fractions (Dasgupta and Hirschmann, 2006; Dasgupta et al., 2013; Rosenthal et al., 2015). These ratios preserved within olivine-hosted melt inclusions can estimate the carbon content of the mantle source.

No direct observations of lower mantle carbon reservoirs have been made using lithophile elements, though planetary mass balance and CO₂ degassing models infer the presence of carbon-rich material in the deep Earth (Marty, 2012; Anderson and Poland, 2017). The concentration of carbon in this reservoir is dependent upon its size and the heterogeneous nature of material in the lower mantle. Likely components hosting carbon within the deep mantle are primordial material that has been isolated for billions of years, and recycled oceanic crust (Fitton et al., 1997; Chauvel and Hémond, 2000; McKenzie et al., 2004; Thirlwall et al., 2004; Peate et al., 2010; Shorttle and Maclennan, 2011; Shorttle et al., 2014). Lower mantle reservoirs are thought to be tapped by the Icelandic mantle plume due to observations from mantle seismic tomography and primordial noble gas isotopic compositions (Rickers et al., 2013; Jenkins et al., 2016; Yuan and Romanowicz, 2017). Therefore, it is possible that lower mantle carbon could be expressed in Icelandic erupted products associated with high plume geochemical signatures.

Chapter 2

Sample collection and data acquisition

2.1 Introduction

Icelandic basalts hold key information about crustal and mantle magmatic processes within the chemistry of phenocrysts, the textures of crystalline phases, and the composition of basaltic glass. Careful interpretation of these observations can allow scientists to separate the different processes responsible for the diversity of Icelandic magmas; from mantle melting, through melt transport, to degassing in the shallow crust. Large, robust datasets are crucial for disentangling these numerous melt modification processes by providing a well-characterised and statistically significant estimate of the magnitude and distribution of chemical variability within the sample of interest.

Olivine-hosted melt inclusions trapped within high forsterite (Fo) olivines are thought to be representative of the range of primitive melt compositions at depth within a magmatic system. The measurement of a large number of melt inclusions trapped within high-Fo olivines provides a more statistically reliable quantification of primitive melt heterogeneity than achieved by only measuring a few. Previous studies on the Miðfell eruption, introduced below, have made interesting observations of primitive melt heterogeneity, but have drawn their conclusions from only \sim 30 melt inclusions (Gurenko and Chaussidon, 1995; Gurenko and Sobolev, 2006). More recent melt inclusion studies have characterised eruptions using tens to hundreds of measurements, as preparation of melt inclusions for analysis has become much more standard practice (Saal et al., 2002; Maclennan, 2008a; Hartley et al., 2014; Neave et al., 2014, 2015; Schipper et al., 2016).

Chapters 5 and 6 draw upon two large melt inclusion datasets, collected from samples of the Kistufell and Miðfell eruptions, to discuss the controls on trace element heterogeneity, and the carbon content of the Icelandic mantle. This chapter outlines each eruption, and details the geological setting and sampling locations. It also describes the various methods

used to acquire major, trace and volatile element data from olivine-hosted melt inclusions, basaltic glass, olivine, and experimental charges. Details of data previously published and used for comparison with the Kistufell and Miðfell are also given to place these eruptions in a wider Icelandic and global context. This chapter also presents the analytical techniques required for measuring experimental igneous phases discussed in Chapter 4, with details of experimental procedures presented in Chapter 3.

2.2 The Kistufell eruption

2.2.1 Eruption overview

The Kistufell eruption is a monogenetic table mountain on the North West margin of the Vatnajökull glacier, and is part of the Bárðarbunga volcanic system in the Northern Volcanic Zone (64° 48' N, 017° 13' W; Figure 2.1). The region around Kistufell shows the highest eruption frequency and crustal accretion in Iceland, with crustal thickness beneath Kistufell in excess of 40 km (Darbyshire et al., 2000). In 2014 seismic activity was detected beneath Kistufell as a result of the Bárðarbunga unrest. Integrating these magmatic observations with those of large gravity anomalies and mantle seismic velocity perturbations suggests that the Vatnajökull region lies above the axis of the Icelandic mantle plume (Pilidou et al., 2005; Hooft et al., 2006).

Post-glacial lavas have obscured the base of the Kistufell eruption and glacial erosion has reduced the volume of the mountain, leaving an estimated exposed volume of $\sim 1 \text{ km}^3$. The structure of Kistufell is composed of a major sequence of subglacially erupted pillow lavas near the base, hyaloclastites and breccias above, all capped by a minor sequence of subaerially erupted lavas (Breddam, 2002). This progression is typical for subglacially erupted mountains in Iceland. Eruptions initiated under ice form pillow basalts, as the overlying ice provides a confining pressure on the meltwater and erupting basalt, preventing magmatic fragmentation. As the eruption progresses, the overlying ice melts and thins, reducing the confining pressure. Once the meltwater is able to expand to steam under low confining pressures, hyaloclastites are formed by the fracturing of lava into glassy shards. This material is often reworked by meltwater, leading to subaquatic sedimentary rock structures within the hyaloclastite. After meltwater-lava interactions stop, the eruption proceeds subaerially forming capping lavas and forms a table mountain or 'tuya' (Thordarson and Hoskuldsson, 2002; Jakobsson and Gudmundsson, 2008).

Kistufell is observed to be a fairly small table mountain relative to others in the region with the table-top lying 300 m above the base and similar to the height of the present-day ice-sheet,


Figure 2.1 A map of the NW Vatnajökull region, showing the Kistufell eruption and sampling locality (purple square) within the Bárðabunga volcanic system, NVZ. Kistufell table-top lavas were erupted subaerially during the last glacial (brown).

suggesting that it was erupted at a time when the overlying ice was relatively thin. Given the position of the capping lava above the plain and observations of ice thinning during modern eruptions under Vatnajökull, overlying ice thickness at the time of the Kistufell eruption has been estimated as 600–700 m (Breddam, 2002). During the Weichselian glaciation (0.01–0.11 Ma) ice thickness was thought to be 1000–1500 m, suggesting that Kistufell erupted towards the end of this period when the ice had thinned considerably (Bourgeois et al., 1998; Jakobsson and Gudmundsson, 2008). The top surface of Kistufell is well preserved suggesting that it has not be glaciated in the past.

The Kistufell hyaloclastites and pillow basalts contain olivine, and occasionally Cr-spinel and/or plagioclase phenocrysts (< 100 μ m), and ubiquitous olivine, Cr-spinel, and plagioclase microphenocrysts (< 100 μ m). These phases are set within a glassy or microcrystalline groundmass (Breddam, 2002). A previous study has identified the presence of highly forsteritic olivines (X_{Fo} \geq 0.90) within the Kistufell eruption, some of which contain melt inclusions < 1 vol% (Breddam, 2002). Glassy pillow rims were targeted for collection at Kistufell, as they are more likely to contain olivines with glassy melt inclusions than pillow

interiors. It is important that analysed melt inclusions are glassy, as it reduces the risk of melt inclusions being strongly modified by post-entrapment crystallisation. Other studies have re-homogenised crystalline melt inclusions by heating of the host olivine until the crystalline phases melt. However, care must be taken with this procedure to prevent reactive melting of the host olivine into the melt inclusion (Head et al., 2011; Cabral et al., 2014; Hudgins et al., 2015). Analysis of glassy melt inclusions prevents the need for this extra process of sample preparation.

Chapter 5 uses data from the Kistufell eruption to explore the controls on trace element heterogeneity of olivine-hosted melt inclusions. The geophysical observations of the Icelandic mantle plume beneath the Kistufell region and primordial noble gas isotopic signatures measured in Kistufell lavas, suggest that the influence of the mantle plume on primitive melt geochemistry could be investigated with this eruption.

2.2.2 Sample collection

Samples of glassy pillow rims were taken from a suite of basalts, basaltic glasses and gabbroic xenoliths collected by Ben Winpenny, Margaret Hartley and David Neave in August 2009. Basaltic glass from pillow rims (sample '09KS04') was collected on the East side of Kistufell at location 'A' from Breddam (2002); at the base of a gully between terminal moraine to the East and a scree slope on the East flank of Kistufell to the West. Pillows were abundant in an ~ 50 m section. Care was taken to avoid sampling of altered basaltic glass. A locality position was recorded by hand-held GPS of 64° 47.442' N, 017° 10.456' W, elevation 949 m (Fig. 2.1).

2.3 The Miðfell eruption

2.3.1 Eruption overview

The Miðfell eruption is situated on the North East shore of Þingvallavatn and is part of the Hengill volcanic system in the Western Volcanic Zone (64° 10' N, 021° 03' W; Fig. 2.2). The crustal thickness in this region is around 25 km (Darbyshire et al., 2000). It is a subglacial picritic eruption of Weichselian age (Trønnes, 1990, 0.01–0.11 Ma). The eruption forms a NE-SW trending hyaloclastite ridge (móbergshryggur), with an older section of pillow basalts exposed towards the SW end, sometimes referred to as Dagmálafell in the literature (Gurenko and Sobolev, 2006).

Olivine-hosted melt inclusions from Miðfell have previously been measured, along with glass/groundmass and phenocryst phases to characterise the petrology of the eruption



Figure 2.2 A map of the NE Þingvallavatn region, showing the Miðfell eruption and sampling locality (red square) within the Hengill volcanic system, WVZ.

(Gurenko and Chaussidon, 1995). Miðfell pillow basalts have abundant olivine phenocrysts, and some plagioclase, clinopyroxene and Cr-spinel phenocrysts (> 100 μ m). These phases are set within a glassy (pillow rim) or microcrystalline (pillow interior) groundmass that contains vesicles. Gabbroic xenoliths have also been found in the Miðfell eruption, located in a single layer within the main exposure of pillow basalts (above the pillow basalts sampled). These gabbros contain large macrocrysts of plagioclase, clinopyroxene and olivine, with some Cr-spinel and interstitial glass (Gurenko and Sobolev, 2006). The previously established primitive nature of this eruption, abundant high-forsterite olivines, and glassy pillow rims make Miðfell an ideal eruption for exploring primitive melt heterogeneity and mantle geochemical signatures.

Chapter 4 uses Miðfell basaltic glass as the starting composition for crystallisation experiments to explore igneous barometry and crystallisation models. Chapters 5 and 6 focus on the compositions of olivine-hosted melt inclusions from Miðfell to gain insight into controls on trace element heterogeneity, and the composition of the Miðfell mantle source and primordial mantle reservoirs.

2.3.2 Sample collection

Samples were collected during September 2014 and consisted of pillow rim glasses (MID2014-01) and pillow interior whole rock (MID2014-02) samples. Weathered lavas containing yellow olivine phenocrysts, mud-filled vesicles, and red-stained surfaces were avoided during sampling. Glassy material was collected from the tops of *in situ* pillow basalts. The sampling location was a partially quarried face on the SE side of Miðfell, just off the main road to the SE following a track towards the mountain. (GPS location: 64° 10.456' N, 021° 02.859' W, elevation 160 m.)

2.4 Analytical methods

2.4.1 Sample preparation

The Icelandic basalts collected from Kistufell and Miðfell were prepared for Secondary Ion Mass Spectroscopy (SIMS) to measure trace and volatile elements, Electon Probe Microanalysis (EPMA) for major element quantification, and Scanning Electron Microscopy (SEM) imaging. A second set of Miðfell olivines was processed separately at Woods Hole Oceanographic Institution (WHOI), USA for X-ray tomography to measure the size of vapour bubbles within melt inclusions, Raman spectroscopy to quantify the CO₂ concentration within the vapour bubbles, and SIMS to measure the trace and volatile elements within the melt inclusion glass. Experimental charges created at WHOI using the Miðfell basaltic glass composition were prepared for EPMA to measure major element compositions of silicate phases and basaltic glass, Quantitative Evaluation of Minerals using SCANning electron microscopy (QEMSCAN), and SEM imaging.

Icelandic basalts

Olivines were picked from hand-crushed separates of unaltered pillow glass from Kistufell and Miðfell. Care was taken to avoid crushing basalt from crystalline pillow interiors to maximise the proportion of glassy olivine-hosted melt inclusions within the crushed separates. Any pillow rim glass that appeared muddy or stained orange from weathering was excluded to prevent contaminated or altered melt inclusions and matrix glass from being analysed. Olivine crystals were selected under a picking microscope for polishing if (i) a glassy melt inclusion could be seen within the crystal, or (ii) the matrix glass coating the olivine looked unaltered and glassy. Selected olivines were individually mounted on glass slides using Crystalbond. Matrix glass-coated olivines were polished using silicon carbide papers (600, 800) to locate any melt inclusions. Olivines containing melt inclusions were then polished down through the crystal to expose the melt inclusion at the surface. A reflected light microscope was used to identify the point at which the melt inclusion was exposed and then finer papers (1200, 2500) were used on the olivine to improve the quality of the polish. Olivines containing cracks through the melt inclusion, crystalline melt inclusions, embayments, melt inclusions dominated by spinel inclusions or very large vapour bubbles were not analysed.

Polished olivines with exposed melt inclusions were removed from the Crystalbond using acetone, further washed in acetone, dried, and then mounted together in one-inch rounds of epoxy resin. A polish with 2500 silicon carbide paper was given to each mount in order to provide a smooth top surface to the epoxy, reducing the potential for surface contaminants. Microdiamond paste polishes (3 μ m and 1 μ m) were used to give each epoxy mount a smooth flat surface to their exposed melt inclusions for SIMS analysis. Mounts were ultrasonically cleaned in de-ionised water before and after use of each grade of microdiamond paste.

Sample mounts were first coated in gold at the NERC Ion Probe Facility at the University of Edinburgh for SIMS, and then coated in carbon at the University of Cambridge for EPMA. This order of analyses was used to avoid carbon contamination of the exposed melt inclusions before carbon measurements had been conducted on the ion probe.

Experimental charges

Experiments conducted at 0.001 kbar were quenched in water, forming glass beads that could be partially crushed by hand and then mounted as glassy fragments in one-inch rounds of epoxy. Experiments conducted at 5 and 10 kbar were removed from the cooled piston-cylinder apparatus (see Chapter 3 for details) and extracted from the centre of the experimental assemblage. At WHOI, each capsule was cut in half using a thin diamond disk bench saw to expose the quenched reaction product. One half of the charge was then mounted in a one-inch epoxy round. All mounted experimental charges were polished using automated bench polishing machines, stepping down through silicon carbide paper grades from 800 to 2500 grit, and finishing with 3 μ m and 1 μ m microdiamond grade, before being carbon coated at the University of Cambridge for EPMA, QEMSCAN, and SEM imaging.

2.4.2 Secondary ion mass spectroscopy

Secondary Ion Mass Spectroscopy (SIMS) is a destructive analytical technique that quantifies isotopic compositions, and hence the elemental concentrations of a sample. A beam of ions with energy ~ 10 keV is accelerated towards the sample surface, transferring energy to the sample and causing some of the atoms to leave the surface; some of these atoms may carry charge. This process, called sputtering, creates a diverse set of ion species, which are chemically and isotopically fractionated. A well collimated primary ion beam is used to limit analysis to a small area (1–10 µm diameter); the secondary ions produced are then accelerated away from the sample surface through a mass spectrometer and counted using detectors, such as electron multipliers. Measured secondary ion intensities are converted to concentrations using the ion yields of known concentration standards (Gill, 2014).

There are some effects of secondary ion production that must be corrected for: (i) molecular ion interferences as some elemental isotopes and molecules have very similar atomic masses e.g. $^{151}\text{Eu}^+ = 150.919702$ and $^{135}\text{Ba}^{16}\text{O}^+ = 150.900580$; and (ii) matrix effects, caused by the interaction of atoms during sputtering ion formation.

Chapters 5 and 6 use SIMS measurements of 129 melt inclusion and 13 matrix glass points from the Kistufell samples, and 108 melt inclusion and 20 matrix glass points from the Miðfell samples. Each sample set was analysed for trace and volatile elements on a Cameca ims-4f instrument over the course of two five-day sessions at the NERC Ion Probe Facility at the University of Edinburgh, uncertainties are presented in this section.

CO₂ analyses in olivine-hosted melt inclusions and matrix glass

Carbon was measured separately from other volatile and trace elements as a moderately high mass resolution configuration was required to ensure good separation of ${}^{12}C^+$ and ${}^{24}Mg^{2+}$ ion peaks. A liquid nitrogen cold ring was placed around the sample chamber in order to reduce volatile element background counts, and field aperture #2 was put in place to reduce measurements to a spot size of < 20 µm.

Beam conditions for the moderately high mass resolution carbon analyses were as follows: a primary O⁻ ion beam with a primary accelerating voltage of 10.7 keV; a secondary accelerating voltage of 4.5 keV minus an energy filter offset of 50 V, giving a net impact energy of 15.2 keV; a beam current of 6 nA; an image field of 150 µm for the second Kistufell session, and 25 µm for the other three sessions.

The analysis spot was rastered over a 15 μ m² area with the O⁻ beam for > 180 s to reduce the chance of surface contamination. For all sessions, isotopes measured for carbon analysis were counted over 15 cycles, with the final eight being used for concentration calculations;



Figure 2.3 Calibration curves for CO_2 -¹²C/³⁰Si count data from measured standards for each day of SIMS measurement; (a) Kistufell, and (b) Miðfell.

this was to ensure that count rates had stabilised and to further reduce the chance of surface contamination affecting the measurements. Isotopes measured in the carbon analysis session were as follows, with count times in parentheses: for Kistufell session one ${}^{24}Mg^{2+}(5)$, ${}^{12}C(10)$, ${}^{28}Si/2(2)$ and ${}^{30}Si(2)$; for the other three sessions ${}^{24}Mg^{2+}(5)$, ${}^{12}C(10)$, ${}^{40}Ca/2(2)$ and ${}^{30}Si(2)$.

A background correction of ¹²C counts per second (cps) was estimated by measurement of the BIR standard, as it contains nominally zero carbon (Jochum et al., 2005). A set of primary standards (ST1, ST2, ST3, ST4, S4_13, S5_13, and 17_2) were measured to convert melt inclusion and matrix glass count data into CO₂ concentrations using the equation below. These basaltic glass standards were from the Edinburgh IMF in-house calibration set and derived from published experimental glass compositions (Pichavant et al., 2009; Shishkina et al., 2010). CO₂ concentrations within the standards ranged from 0–2552 ppm, covering the expected range of CO₂ within the melt inclusions. Calibration curves for each carbon analysis day for both Kistufell and Miðfell are shown in Figure 2.3. The gradient of each calibration curve was used to convert sample count ratios to CO₂ concentration:

$$CO_2 (ppm) = \frac{{}^{12}C (cps) - {}^{12}C background (cps)}{{}^{30}Si (cps)} \times calib. gradient (ppm).$$

Accuracy of the CO_2 measurements was calculated using a jackknife accuracy method with the primary standard set. Each standard was in turn left out of the initial calibration of the counts data, and then its CO_2 concentration was calculated using this calibration. The omitted standard could then be used as a secondary standard to calculate a value for accuracy by dividing recovered CO_2 by known CO_2 . The accuracy values for each omitted standard were combined to give an average accuracy (Table 2.1). Each standard gave a similar value for accuracy. CO_2 precision was calculated using repeat analyses of the matrix glass, as it was assumed that the matrix glass was homogeneous and this composition had the most repeat analyses, therefore giving the most reliable estimate of precision. Accuracy and precision values suggest that confidence can be placed on the SIMS CO_2 measurements.

H₂O, F, Cl and trace element analyses in melt inclusions and matrix glass

 H_2O , F, Cl and trace elements were measured after all carbon analyses had been completed using a low mass resolution SIMS setup. Beam settings remained the same as the carbon analyses, apart from the energy filter offset, which was increased to 75 V. The same analysis pits were used as before, and a 15 μ m² pre-analysis raster for > 120 s was used to reduce the chance of surface contamination. During this time, isotope peaks were verified and centred for each of the measured species.

The following isotopes were measured over eight cycles (count times in brackets): ¹H(5), ⁷Li(5), ¹⁵F(5), ³⁰Si(2), ³⁵Cl(5), ³⁷Cl(5), ³⁹K(2), ⁴⁷Ti(2), ⁸⁸Sr(3), ⁸⁹Y(3), ⁹⁰Zr(3), ⁹³Nb(3), atomic mass 130.5(3), ¹³⁸Ba(3), ¹³⁹La(3), ¹⁴⁰Ce(3), ¹⁴¹Pr(5), ¹⁴³Nd(5), ¹⁴⁹Sm(8), ¹⁵¹Eu(8), ¹⁵⁶Gd(5), ¹⁵⁷Gd(4), ¹⁵⁹Tb(5), ¹⁶¹Dy(5), ¹⁶⁵Ho(5), ¹⁶⁷Er(8), ¹⁶⁹Tm(8), ¹⁷¹Yb(10), and ¹⁷⁵Lu(10).

Measurement of atomic mass 130.5 provided an estimate of background counts for each analysis. Count data was processed using JCION-6 software at the NERC IMF at the University of Edinburgh. This program used analyses of NIST610 as a primary standard to convert ion yield into concentrations, with corrections for molecular interferences and matrix effects (Jochum et al., 2011). Initial concentrations were calculated using an ion yield based on the assumption that the unknown glasses contained 50 wt% SiO₂, so corrections had to be applied to the trace and volatile element dataset once SiO₂ had been measured by EPMA at the University of Cambridge.

Secondary standard corrections

Measurements from each analytical session were compared with each other to ensure internal consistency by using secondary standards and repeat analyses of matrix glass. A set of known standards (ST1, ST2, ST3, ST4, S4_13, S5_13, 17_2, T1-G, BCR-2G, and GSD-1G) was used with water concentration ranging from 0.02–4.82 wt% and only two standards below 1 wt%. Water was calibrated and checked in the same way as the trace elements,



Figure 2.4 SIMS trace element yield estimates presented as recovered fractions of trace element secondary standards GSD-1G and BCR-2G (Jochum et al., 2005), and T1-G (Jochum et al., 2006). (a) Kistufell measurements of GSD-1G; (b) Miðfell measurements of GSD-1G; (c) corrected Kistufell measurements of BCR-2G with yield fractions at the bottom of the plot; (d) corrected Miðfell measurements of T1-G with yield fractions at the bottom of the plot.

using a NIST610 primary standard and then the secondary standards listed above. These standards highlighted that accuracy was poor for Kistufell session one and both Miðfell sessions. Figures 2.4a and b show the estimates of trace element accuracy for Kistufell and Miðfell, respectively, based on repeat measurements of secondary standard GSD-1G (Jochum et al., 2005). GSD-1G is a basaltic glass, close in composition to Kistufell and Midfell glasses, therefore it is an appropriate secondary standard to use in applying a correction to the dataset. Trace element concentrations for the sessions highlighted above have been corrected such that measured GSD-1G concentrations are the same as known reference values. Trace element yields have been estimated for this corrected data using other secondary standards: BCR-2G for Kistufell (Jochum et al., 2005, Figure 2.4c) and T1-G for Miðfell (Jochum et al., 2006, Figure 2.4d). Trace element yields for SIMS are shown in Table 2.1 and secondary standard corrections are detailed in Appendix A.

Vapour bubble-melt inclusion reconstructions

A separate set of 23 melt inclusions, 13 of which had their CO₂ contents reconstructed from vapour bubble and melt inclusion measurements, were analysed at the North-East National Ion Microprobe Facility at Woods Hole Oceanographic Institution (WHOI) by Glenn Gaetani. Five matrix glass measurements were also made. Volatile concentrations (CO₂, H₂O, F, S, and Cl) were measured using the Cameca ims-1280 using a 1 nA ¹³³Cs⁺ primary beam focused to a $< 10 \times 10 \,\mu\text{m}^2$ spot size and rastered over $\sim 20 \times 20 \,\mu\text{m}^2$ for > 120 s. Secondary ions were extracted with a 10 kV secondary voltage potential. Mass spectrometer entrance and exit slits were set to achieve mass resolution ~ 6000 in order to separate ${}^{16}O^{1}H^{-}$ from ${}^{17}O^{-}$ peaks and resolve all other known potential isobaric interferences. Background and surface volatiles were minimised by centring secondary ions within a field aperture to allow transmission of ions from only the innermost $10 \times 10 \,\mu\text{m}$ of the sputtering crater. Isotopes were taken over 5 cycles, with typical measurement precision < 1% (relative standard deviation). Volatile concentrations were derived from measurement of eight natural basaltic glass standards of well-characterised volatile composition and very low volatile synthetic forsterite background indicator (Hauri et al., 2002b). Trace and rare earth element analyses were made using the Cameca ims-3f at WHOI using a 1 nA O⁻ primary beam focused and rastered to a total spot size of $\sim 40 \,\mu\text{m}$ in diameter. Secondary ions (as above) were extracted with a secondary voltage of 4.5 kV, with an energy offset of 75 V and an energy window of \pm 30 eV. Mass spectrometer entrance and exit slits were fully open for a mass resolution of \sim 500. Analyses included magnet cycling, 120 s of pre-sputtering, and mass calibration followed by five cycles of isotope measurement. Electron multiplier background (in counts per second) was measured during the analytical routine by counting on the electron multiplier with the secondary magnet set for off-peak mass 6.7. Corrections for isobaric molecular oxide interferences on many rare earth element masses followed procedures established by Zinner and Crozaz (1986). Concentrations were derived from a calibration on standards of known concentrations, which included MPI-DING glasses ATHO-G, T1-G, and StHs 6/80 (Jochum et al., 2006).

To confirm that the measurements made at WHOI and Edinburgh IMF are in agreement, Miðfell matrix glasses were measured at both institutions. The analyses for comparison are plotted in Figure 2.5, which shows that the majority of trace element measurements are in agreement between the two institutions. The main differences are in measurements of Eu, Sm and to a lesser extent Nb and Ba. All the Eu measurements made at WHOI are lower than those made at Edinburgh, while Sm, Nb, and Ba measurements are higher. It can be concluded that while the two instruments are in good agreement and their analyses can be used together as one dataset, caution should be taken when comparing Eu, Sm, Nb, and Ba



Figure 2.5 Comparison of Miðfell matrix glass trace element analyses conducted at the Ion Microprobe Facility (IMF), University of Edinburgh; and Woods Hole Oceanographic Institution (WHOI).

measurements between the two institutions. If the WHOI analyses are taken to be correct then CO₂/Nb and CO₂/Ba ratios presented in Chapter 6 are overestimated, along with the carbon estimates derived from them. However, Edinburgh IMF trace element analyses have been corrected using secondary standard GSD-1G, as detailed in the previous section, to ensure they are as accurate as possible. Accuracy is ~ 85% for the majority of Miðfell trace elements (Figure 2.4; Table 2.1), which is of slight concern as it suggests underestimation of the trace element concentrations. However, these values have been calculated from only three repeats of secondary standard T1-G, so repeatability could be affecting the accuracy calculation. Precision values from repeat analyses of the matrix glass are lowest for the most incompatible elements, as they are present in the highest concentrations. The heavy rare earth elements show lower precision, but around the expected value of ~ 10%.

2.4.3 Electron probe microanalysis

Electron Probe Microanalysis (EPMA) is an non-destructive analytical technique for quantifying the chemistry of small areas of solid samples. A beam of electrons is fired at the sample, which causes some lower shell electrons to be displaced from atoms present in the sample. Outer shell electrons drop down to fill the vacancy in the lower shell, releasing a photon with energy equivalent to the energy step between valance shells, which in most cases is of X-ray frequency. Each element produces a characteristic set of X-rays, which appear as lines on an X-ray spectrum. Concentrations of each element present within the X-ray spectrum are quantified by comparing the intensity of X-rays from the sample with the intensity of X-rays from a known standard. This approach requires the use of a wavelength dispersive spectrometer to improve the resolution of the X-ray spectrum. Diffracting crystals used at the University of Cambridge are made from Lithium Fluoride (LIF), Pentaerythritol (PET), and Thallium Acid Phthalate (TAP). A matrix correction is applied for: (i) atomic number effects; (ii) attenuation of X-rays produced in the sample; and (iii) fluorescence, i.e. the secondary generation of X-rays.

Olivine-hosted melt inclusions and matrix glasses

Major element compositions for melt inclusions, matrix glass and host olivine were measured by electron probe microanalysis (EPMA) using a Cameca SX-100 electron microprobe at the University of Cambridge. Matrix glass analyses were performed with a spot size of $10 \,\mu m$, an operating potential of 15 keV and a beam current of 10 nA. A 1 µm beam with a current of 20 nA was used for measuring olivine major element concentrations. Counting times for glass analyses were as follows: 10 s (Na, K), 20 s (Si, Fe), 60 s (Mg, Al, Ca, Cr, Mn), 80 s (Ti), and 120 s (P). Counting times for olivine analyses were as follows: 20 s (Si), 30 s (Mg, Fe, Cr, Mn), 60 s (Ni), 120 s (Al, Ca, Ti). F, Cl and S were all measured with a 100 nA beam current and for 240 s each. Calibration standards were as follows: jadeite for Na, periclase for Mg, diopside for Si and Ca, corrundum for Al, K-feldspar for K, apatite for P, rutile for Ti, fayalite for Fe, chromium metal for Cr, manganese metal for Mn, fluorite for F, halite for Cl, and pyrite for S. Secondary standards from the Smithsonian reference set were analysed to check the electron probe calibration. Repeat analyses of San Carlos olivine, basaltic glasses 113716 and 111240-52, and matrix glass provided estimates of precision and major element accuracy (Jarosewich et al., 1980). Table 2.1 shows that electron probe precision for the key oxides in basaltic glass is high, with less significant components (Cr_2O_2 , MnO, and K_2O) showing much lower precision. Accuracy values for significant major oxides are > 98%.

Experimental charges

Major element compositions of glass, olivine, plagioclase and pyroxene were measured from the experimental charges created at WHOI. Calibration standards were the same as detailed for olivine-hosted melt inclusions (see above). Glass count times were as follows: 10s (Na), 20s (Si, Mg, Al, K, Ca, Fe), 60s (Cr, Mn, Ni), 90s (S, P), 150s (Ti). Olivine count times: 20s (Si, Mg, Fe), 30s (Cr, Mn), 90s (Ca, Ti), 120s (Ni), 180s (Al, P). Plagioclase count times: 10s (Na, K), 20s (Si, Al, Ca), 40s (Fe, Mn), 90s (Ti, Mg). Clinopyroxene count times: 10s (Na), 20s (Si, Ca, Fe), 30s (Mg, Al, Cr, Mn), 90s (Ti).

		Kistufell		Miðfell				
	EPMA	SI	MS	EPMA	SIMS			
Element	P (%) ^a	A (%) ^b	P (%) ^c	P (%) ^d	A (%) e	P (%) ^f		
SiO ₂	0.8	-	-	0.5	-	-		
TiO ₂	2.1	-	-	2.0	-	-		
Al_2O_3	0.7	-	-	0.3	-	-		
Cr_2O_3	49.3	-	-	78.8	-	-		
FeO _t	1.5	-	-	1.3	-	-		
MnO	20.0	-	-	16.6	-	-		
MgO	0.8	-	-	1.0	-	-		
CaO	1.0	-	-	0.6	-	-		
Na ₂ O	2.5	-	-	2.7	-			
K_2O	17.5	-	-	14.3	-	-		
P_2O_5	15.3	-	-	6.9	-	-		
CO ₂	-	101.7	17.2	-	95.8	10.7		
H_2O	-	101.5	18.7	-	103.6	20.1		
F	47.5	91.8	9.0	49.1	33.9	14.1		
Cl	39.5	84.3	67.9	18.3	128.2	77.5		
S	2.0	-	-	1.0	-	-		
Li	-	-	-	-	92.4	15.4		
Κ	-	95.6	10.1	-	83.2	4.7		
Ti	-	91.4	2.4	-	80.6	3.8		
Sr	-	95.2	4.0	-	83.2	3.3		
Y	-	96.0	2.1	-	86.7	3.2		
Zr	-	98.4	3.4	-	88.8	4.1		
Nb	-	100.5	10.7	-	82.9	11.3		
Ba	-	91.8	9.3	-	83.8	4.9		
La	-	97.1	4.7	-	85.0	9.7		
Ce	-	93.9	2.8	-	84.3	8.8		
Pr	-	92.4	9.7	-	83.8	10.7		
Nd	-	94.5	8.0	-	84.9	11.3		
Sm	-	100.0	7.3	-	88.6	13.7		
Eu	-	99.3	8.2	-	91.9	12.9		
Gd	-	111.4	10.8	-	87.6	16.7		
Tb	-	107.2	12.1	-	94.7	16.4		
Dy	-	101.9	10.7	-	95.4	10.1		
Но	-	106.6	9.0	-	86.7	13.0		
Er	-	108.0	11.4	-	104.5	12.6		
Tm	-	106.9	13.3	-	100.3	13.6		
Yb	-	98.4	8.5	-	71.5	15.3		
Lu	-	108.4	10.9	-	77.3	19.6		

Table 2.1 Representative estimates of precision and accuracy for SIMS and EPMA analyses of melt inclusions and matrix glasses from Kistufell and Miðfell.

^a Number of repeat analyses, n, of matrix glass (n = 27, n = 9 for Cl). ^b Recovery of secondary standard BCR-2G after correction by secondary standard GSD-1G for session one (n = 4); CO₂ jack-knife accuracy (n = 13); H₂O yield from H standards (Pichavant et al., 2009, n =10); Cl yield from measurement of secondary standard KL-2G (n = 1). ^c Repeat analyses of matrix glass (n = 13). ^d Glass standard repeats (n = 10). ^e Recovery of secondary standard T1-G after correction by secondary standard GSD-1G (n = 3); CO₂ jack-knife accuracy (n = 21); H₂O accuracy from H standards (n = 23). ^f Repeat analyses of matrix glass (n = 17); CO₂ average precision of standards (n = 19).

	Glass		Oliv	vine	Plagio	oclase	Clinopyroxene		
Element	A (%)	P (%)	A (%)	P (%)	A (%)	P (%)	A (%)	P (%)	
SiO ₂	99.9	0.4	99.1	0.6	99.0	0.5	100.1	0.8	
TiO ₂	101.4	0.4	-	-	104.4	55.3	101.0	1.8	
Al_2O_3	101.4	1.0	93.3	6.0	99.4	0.5	101.6	0.5	
Cr_2O_3	82.2	95.4	167.0	111.8	-	-	-	-	
FeO	101.5	1.1	99.3	1.6	104.3	6.0	100.3	2.3	
MnO	90.4	11.5	102.9	18.3	16.2	231.9	82.2	41.9	
MgO	99.0	1.8	99.9	0.8	143.7	24.5	99.9	0.5	
CaO	100.2	1.8	137.1	35.7	99.9	0.7	100.4	0.2	
NiO	100.8	149.2	101.2	2.7	-	-	-	-	
Na ₂ O	100.0	2.4	-	-	101.2	6.5	100.5	3.2	
K ₂ O	102.1	3.0	-	-	113.7	43.8	-	-	
P_2O_5	99.2	5.7	-	-	-	-	-	-	
SO ₂	108.6	26.3	-	-	-	-	-	-	

Table 2.2 Representative estimates of precision (P) and accuracy (A) for glass, olivine, plagioclase and clinopyroxene major element measurements from Miðfell experimental charges.

Two beam spot sizes were used for both glass and plagioclase analyses: 1 and 10 μ m for glass and 1 and 5 μ m for plagioclase. A more diffuse beam was preferentially used, where grain size allowed, to reduce the effect of alkali drift and due to the instability of glass in a focused beam.

Secondary standard corrections

The electron probe calibration was checked using secondary standards of basaltic glasses 113716 and 111240-52, San Carlos olivine, anorthite, augite, and diopside (Jarosewich et al., 1980). Accuracy values are close to 100% and precision is below 2% for the major oxides within each phase (Table 2.2). A secondary standard correction was applied to experimental charge glass measurements to make the analyses consistent across the three EPMA sessions. Repeat analyses of glass from experiment MID2014-20 were used to compare the corrected values and provide an estimate of analytical precision. Secondary standard corrections are detailed in Appendix A.

2.4.4 Uncertainty analysis

Precision and accuracy

Tables 2.1 and 2.2 give estimates of precision and accuracy for EPMA and SIMS measurements of major element oxides, trace and volatile elements. They have been calculated using repeat analyses from a number of secondary standards as well as the matrix glass, which is



Figure 2.6 Calculated relative error from repeat analyses against measured trace element concentrations of secondary standards and repeated unknowns; (a) Nb, (b) Ba, (c) Ce, and (d) CO₂. Red curve is fit(1), green curve is fit(2). Grey dashed line is an average value for CO₂.

assumed to be homogeneous in composition. CO_2 accuracy was calculated using a jackknife method, where accuracy values were calculated from each calibration by leaving out one of the standards in the primary calibration and using it for accuracy calculation.

Relative error dependency on concentration

Precision can be described in terms of standard deviation, σ , or relative error (standard deviation divided by the mean; σ/\bar{x}). Repeats of standards are used to calculate the standard deviation around the mean of repeat measurements, and the relative error can be calculated using the mean value. Standards of different elemental concentrations can give different estimates of precision and often the unknowns measured will have a composition different to that of the standards. Comparing relative error estimates to average secondary standard concentration and fitting a curve to this relationship is an attempt to better interpolate the error of unknowns based on the error of measured standards.

Figure 2.6 shows the relationship between relative error and standard concentration for Nb, Ba, Ce and CO_2 , based on repeat analyses of standards during the Miðfell SIMS sessions. The general trend of the data shows that relative errors are greatest at low measured

concentrations, showing a clear dependency of relative error on the sample concentration. Curves have been fitted to the data with equations of the form:

$$fit(1): Y = a\left(\frac{1}{X}\right) + b; \quad fit(2): Y = dX^e + f,$$

where Y = relative error, X = concentration of measured species, a, b, d, e and f are all fitted constants.

Figures 2.6a–b show that for Nb and Ba both equations fit the data reasonably well, with large relative errors at low trace element concentrations. Both equations give the same fit for Ce and match the data very well (Figure 2.6c). The relationship does not hold for CO₂ standards, and as a result an average value of relative error, independent of concentration, has been calculated (Figure 2.6d). This relationship can be further explored by analysing the ¹²C count data for the secondary standards to see if the carbon signal was varying across successive cycles. For all carbon standards, the count data shows low variability (< 4% relative error) with the first 1–3 cycles often recording slightly higher counts than the following 12. The count data therefore suggests that ¹²C was measured consistently across the cycles used to calibrate the data, and is not the cause of variation in secondary standard relative error. Equations for *fit*(1) and *fit*(2) can be used to calculate the relative error of measured unknowns, providing the fit to the standards data is good. These equations have been used to estimate trace element uncertainties for Miðfell melt inclusions in Chapters 5 and 6, based on the compositions of an enriched, an average, and a depleted melt inclusion. Best fit equation coefficients are given in Appendix A.

2.4.5 Scanning electron microscopy

Scanning Electron Microscopy (SEM) is a non-destructive analytical and imaging technique. A beam of electrons is fired at, and scanned across, the surface of a sample. Some electrons are deflected back off the surface, with the mean atomic number of the atoms interacting with the electron beam controlling the amount of electron absorption, and hence the intensity of back scattered electrons picked up by the detector. A flat sample surface is required to reduce topographic electron scattering effects that could otherwise be confused with mean atomic mass variation, and the sample must be coated to prevent charging and subsequent deflection of the electron beam (Reed, 2005).

Experimental charges were imaged using SEM to help identify different silicate phases and to more clearly observe the microstructure within each experimental charge.



Figure 2.7 Micro-Raman spectrum of a CO₂-bearing bubble in an olivine-hosted melt inclusion from Hartley et al. (2014). Separation of the Fermi diads, Δ , is proportional to CO₂ density (Kawakami et al., 2003).

2.4.6 Vapour bubble-melt inclusion CO₂ reconstructions

Vapour bubbles form in some melt inclusions upon decompression of their olivine host, reducing the concentration of CO₂ dissolved within the melt inclusion glass. To reconstruct the initial dissolved CO₂ content of the melt inclusion, the mass of CO₂ within the vapour bubble must be added to the dissolved portion in the melt inclusion glass. This reconstruction is done by measuring the vapour bubble volume using X-ray tomography and the CO₂ density using Raman spectroscopy to calculate the mass of exsolved CO₂. The presence of CO₂ is confirmed by the Fermi diad on a Raman spectrum (two peaks at ~ 1285 and ~ 1388 cm⁻¹, bounded by hot bands), with the separation of these bands (Δ ; Figure 2.7) proportional to CO₂ density based on the equation of Kawakami et al. (2003). The higher the degree of separation, the higher the CO₂ density.

Thirteen melt inclusions with vapour bubbles from Miðfell were measured at WHOI in order to reconstruct their total CO_2 contents, by combining the CO_2 within the melt inclusion glass (SIMS) and the CO_2 within the vapour bubble (Raman spectroscopy). The combined X-ray tomography, Raman spectroscopy and SIMS technique for these melt inclusion reconstructions has been developed by Glen Gaetani, Véronique Le Roux, and Frieder Klein (pers. comm.).

X-ray Tomography

X-ray micro-computed analyses were performed by Le Roux *et al.* using a Skyscan 1272 table-top system at WHOI. Olivine grains were securely mounted and scanned for ~ 10

hours at 70 kV and 142 μ A. The pixel resolution ranged from 1.0–1.4 μ m. The 3-D data were reconstructed using NRecon software, and the melt inclusion and vapour bubble volumes were calculated using CTAn and Avizo 3-D data processing software.

Raman Spectroscopy

Raman spectra of CO₂ in vapour bubbles were obtained by Klein *et al.* using a confocal Raman spectrometer (Horiba LabRam HR) equipped with a thermoelectric cooled charge-coupled device (CCD) detector (Synapse[®], 1024 × 256 pixels, 26 µm × 26 µm pixel size) and a spectrometer with a focal length of 800 mm. A 532 nm laser was used with a grating of 1800 grooves per mm, a slit size of 100 µm, and a 100× long working distance objective with a numerical aperture of 0.8 for all analyses. The instrument was calibrated daily with pure silicon and diamond standards. Peak drift was monitored using the Raman bands of a Hg-Ne lamp. The distance of the Fermi doublet (bands at ~ 1285 cm⁻¹ and ~ 1388 cm⁻¹) was measured to calculate the density of CO₂.

2.5 Data collation

2.5.1 Database of Icelandic melt compositions

Chapter 5 makes use of a large database of Icelandic melt inclusion compositions. The database contains 692 analyses on melt inclusions from 16 different eruptions, with details of major, trace element, and host olivine major element concentrations. A list of sources can be found in Chapter 5, Table 5.3.

This database is used to put the Kistufell and Miðfell eruptions into the wider context of Iceland. Comparisons can be made across the different rift zones, to other melt inclusion suites, and to other primitive melts.

2.5.2 Database of global melt inclusion compositions

Chapters 5 and 6 make use of a large database of published melt inclusion compositions, which are mainly olivine-hosted, from around the world. The database contains eruption locality information, major and trace element compositions of the melt inclusions, host crystal major element compositions, and details on vapour bubble reconstructions, melt inclusion re-homogenisation, and post-entrapment crystallisation corrections. It has been compiled from 23 different published sources, details of which can be found in Appendix B, and has 1609 melt inclusion analyses and 51 mid-ocean ridge basalt (MORB) glass measurements.

Chapter 3

Experimental petrology methods

3.1 Introduction

Igneous petrology sets out to understand the structure and processes of magmatic systems using observations of petrography and geochemistry on igneous rocks. One of the key questions in the field of igneous petrology is what mechanisms are responsible for the formation of crust (Henstock et al., 1993; Kelemen et al., 1997b; Annen et al., 2006; Coogan, 2014). This question can be explored with knowledge of the depths at which magma is ponding in reservoirs and crystallising silicate phases. Geophysical techniques have enabled the imaging of melt within the crust in Iceland with the use of microseismicity (Brandsdóttir et al., 1997; Lees, 2007; Mitchell et al., 2013; Greenfield and White, 2015; Greenfield et al., 2016; Kim et al., 2017). However, it is the pressure of magma on the surrounding rock that produces the seismicity, limiting the structural imaging to the confining shape as magmatic pressure changes or dykes intrude. Seismic velocities of the material imaged within the crust can give information on the mechanical properties of crustal material, however these values are an average of the assemblage making it difficult to tease apart the finer details of the mineral assemblage (Maclennan et al., 2001b).

Observations of the geochemistry and petrology of erupted volcanic products can complement the seismic imaging constraints on the structure of the oceanic crust. Igneous barometry aims to constrain the pressure at which melts are crystallising silicate phases, which happens as magma cools in a melt reservoir. Pressures are calculated from the equilibrium relationships of igneous phases (Carmichael et al., 1970; Holland and Blundy, 1994; Nimis, 1995; Putirka et al., 1996; Putirka, 2008b; Neave and Putirka, 2017). These barometers are calibrated on the results of pressurised melting experiments, which provide the compositions of silicate phases in equilibrium with each other at the experimental conditions. There are a range of barometers available to the igneous petrologist, however they vary in reliability, can differ in recovered pressure, and have large uncertainties (Yang et al., 1996; Maclennan et al., 2003a; Winpenny and Maclennan, 2011; Neave and Putirka, 2017). To better constrain pressures of equilibration between igneous phases, and hence depths of crystallisation, improvements to barometry calibrations are required from more experiments.

Masotta et al. (2013); Neave and Putirka (2017) and others have started to recalibrate igneous barometers based on the results of new experiments and more careful selection of previous experiments. However, the key component missing in these calibrations is experiments conducted at mid-crustal pressures (3–7 kbar). Melting experiments have been conducted at atmospheric pressure and > 10 kbar for several decades (Boyd and England, 1960; Johannes et al., 1971; Gaetani and Grove, 1998). However, this pressure range misses out the majority of the crust, which is a region of fundamental interest for igneous petrology. Mid-crustal pressures have previously been difficult to stabilise in piston-cylinder pressure vessels due to problems with incomplete compression of the experimental assemblage, thermocouple instability, and the use of a low friction pressure medium (Moore et al., 2008). The piston-cylinder method requires refinement to be able to conduct more experiments equilibrated at crustal pressures, which would provide an expanded dataset capable of producing reliable mid-crustal calibrations of igneous barometers.

This chapter describes the experimental methods developed and used at Woods Hole Oceanographic Institution (WHOI) to equilibrate basaltic glass from the Miðfell eruption at 0.001, 5, and 10 kbar and over a range of temperatures. The aim of these experiments was to produce silicate phase equilibration data that could aid the re-calibration of the clinopyroxeneliquid barometer, and to test previously established phase equilibrium modelling tools, such as Petrolog and MELTS (Asimow and Ghiorso, 1998; Putirka, 2008b; Danyushevsky and Plechov, 2011).

3.2 Starting material

The starting composition used for crystallisation experiments is based on Miðfell basaltic glass, sampled from the glassy rims of pillow basalts (MID2014-01; see Chapter 2 for details of sample location and Table 3.1 for bulk composition). The presence of phenocrysts within the Miðfell eruption suggests that the chosen starting glass composition has experienced fractionation and is therefore not a primitive melt composition. Justification for using Miðfell glass is as follows: (i) The aim of the experiments was to produce data capable of aiding recalibration of the jadeite-in-clinopyroxene barometer, therefore a primitive melt composition is not necessary, but rather that clinopyroxene appears on the liquidus. Experiments were not aiming to recreate the past crystallisation history of Miðfell. (ii)

There was insufficient time at WHOI to both make a synthetic composition and develop a 5 kbar experimental method. The primitive Miðfell melt composition could have been estimated from olivine-hosted melt inclusions, or modelled by back-fractionation of the Miðfell matrix glass. (iii) The depleted nature of Miðfell glass complements more enriched Icelandic compositions that other researchers were experimenting on at the time.

3.2.1 Preparation

Glass chips were picked under a microscope from hand-crushed glassy Miðfell pillow rim material at the University of Cambridge. Care was taken to avoid phenocrysts, crystalline glass, or altered glass as incorporation of these phases into the bulk starting composition could have modified its chemistry.

Around 10 g of starting material, in the form of glass chips, was sent over to WHOI for further sample preparation. The glass chips were cleaned in an ultrasonic bath, first with de-ionised water, and then in ethanol to remove any dirt or dust that may have been coating the glass. A small glass chip was tested in 0.1 M HCl solution to be certain that HCl cleaning would not dissolve away the glass before the rest of the starting material was cleaned in HCl. There was no significant mass change to the small glass chip (mass before 0.03033 g, mass after 0.03024 g), the small amount of mass loss was due to the loss of fine-grained suspension in the HCl solution. The rest of the starting material was cleaned in 0.1 M HCl in a platinum mortar (mass before 10.01365 g, mass after 9.79795 g).

An agate ball mill was cleaned with crushed silica sand before the Miðfell glass chips were crushed to a powder. The powder was further homogenised in ethanol using an automatic pestle and mortar over four one-hour periods. After each run the ethanol was allowed to evaporate and the powder was returned to the base of the mortar. Once dried for the final time, the powder was broken up with a plastic ball in a plastic vial using a floor shaker. The total amount of Miðfell glass powder available for experiments was 9.60208 g. When not in use, the powder was stored in a vacuum oven at 120 $^{\circ}$ C to keep the powder dry.

 Table 3.1 Bulk starting composition of Miðfell basaltic glass used for all experiments.

n = 20	SiO ₂	TiO ₂	Al_2O_3	Cr ₂ O ₃	FeO	MgO	CaO	MnO	Na ₂ O	K ₂ O	P_2O_5	Total
\overline{x}	47.06	0.80	14.87	0.12	9.52	10.80	14.20	0.17	1.51	0.02	0.03	99.12
σ	0.35	0.02	0.17	0.03	0.15	0.14	0.14	0.03	0.03	0.01	0.01	-

Twenty glass analyses, n, were used to give an average bulk starting composition, \bar{x} , and a standard deviation, σ .

3.3 Experiments at 0.001 kbar

The first set of experiments was conducted at atmospheric pressure (1 atm = 0.001 kbar) using a Deltech gas mixing furnace, which allowed for the control of temperature and oxygen fugacity conditions (Figure 3.1).

3.3.1 Sample preparation

In a Deltech gas mixing furnace, the experimental starting composition hangs as a molten ball from a loop of Re wire (Figure 3.1). Therefore, careful preparation of the starting material is required to achieve experimental stability. 0.200 g of starting composition glass powder was placed in a 5 mm die with two drops of polyvinyl alcohol solution (< 10 ml) and compressed by a weight of two tonnes for five minutes. This process formed a 5 mm (diameter) pellet of basaltic glass powder. A Re loop was formed by twisting a short length of Re wire around a 4.4 mm drill blank. The powder pellet was balanced on the Re loop and melted into place using a blowtorch.

3.3.2 Experimental set-up

Before 0.001 kbar experiments could be carried out, the hotspot of the Deltech furnace had to be located. Once the Deltech had been brought up to the required experimental temperature, the Pt-Pt₉₀Rh₁₀ thermocouple was incrementally lowered towards the centre of the Deltech, recording the measured temperature and exposure length of the thermocouple at each step. Once the peak temperature had been reached, the thermocouple, holder and oxygen sensor assemblage was removed from the furnace and allowed to cool. The thermocouple was then returned to the height at which it recorded the highest temperature. The ZrO_2 –CaO electrolyte oxygen sensor was adjusted to the same height, and the sample holder moved to just above this position, so that the sample would hang from the Re loop into the temperature hotspot (Figure 3.1).

3.3.3 Experimental procedure

All 0.001 kbar experiments followed a standard experimental procedure (Bucholz et al., 2013). The Re loop and starting glass powder pellet were suspended from the sample holder by a short length of thin Pt wire. The Deltech furnace was brought up to experimental temperature, with water flowing through the cooling brackets to keep the ends of the ceramic furnace casing cool. A mixture of CO and CO_2 was set flowing through the column to maintain



Figure 3.1 Schematic of the 0.001 kbar experimental set-up using a Deltech gas mixing furnace with the sample suspended from a Re loop by Pt wire in the hotspot of the furnace. Temperature monitored by a $Pt-Pt_{90}Rh_{10}$ thermocouple. Oxygen fugacity monitored by a ZrO_2 -CaO electrolyte oxygen sensor, and controlled by CO and CO₂ gas flow rates.

stable oxygen fugacity. A ZrO₂-CaO electrolyte oxygen sensor, calibrated against the Fe-FeO and NNO buffers, was used to monitor the oxygen fugacity conditions throughout the experiment. An experimental oxygen fugacity of -1.5 $\Delta \log fO_2$ (FMQ) required an oxygen sensor output of -688 mV, which was checked regularly. This oxygen fugacity was chosen to be consistent with that of the piston-cylinder experiments, where the graphite capsule buffers oxygen fugacity based on the CCO (graphite-oxygen) buffer. Oxygen fugacity for the pistoncylinder experiments was predicted using THERMOCALC to model the pressure-temperature dependency of the CCO buffer (Figure 3.2).

Gas flow rates were initially high to maintain a low oxygen fugacity as the experimental assemblage was lowered into the furnace, taking care not the touch the sides of the ceramic furnace, or to shake the basaltic pellet loose as it neared the temperature hotspot. Once in place, the assemblage was checked from below using a mirror, the bottom bung was replaced, and the gas flow rate was reduced to prevent Na-loss from the molten basaltic glass



Figure 3.2 Calculated oxygen fugacity for the three experimental pressures. 0.001 kbar (purple) held at -1.5 $\Delta \log fO_2$ (FMQ) by the mixing of CO and CO₂ gases. 5 kbar (red) and 10 kbar (blue) experimental oxygen fugacity buffered by the graphite capsule, i.e. the CCO buffer. Diamonds mark the highest and lowest temperature experiments at each pressure.

(Figure 3.1). Oxygen fugacity and temperature were checked every few hours and recorded. Gas flow rates were adjusted if necessary to maintain a constant oxygen fugacity. After a run time of 24 hours, to keep Na-loss to a minimum, the basaltic charge was quenched into de-ionised water placed beneath the furnace by passing a current through the thin Pt wire suspending the assemblage and causing the wire to melt. The quenched glassy sphere was removed from the water, dried, crushed, and mounted in a one-inch round of epoxy. The fragments were polished using silicon carbide papers and water-based diamond suspension down to 1 μ m (Chapter 2).

3.4 Experiments at 10 kbar

Experiments at equilibration pressures of 10 kbar were conducted using a 12.7 mm (1/2-inch) diameter experimental assemblage and a piston-cylinder apparatus (Figures 3.3 and 3.4, respectively; Boyd and England 1960; Mirwald et al. 1975; Gaetani and Grove 1998).

3.4.1 Assemblage preparation

The assemblage shown in Figure 3.3 is valid for 10 kbar experiments as well as 5 kbar experiments, but with a few modifications. The 10 kbar assemblage measures 12.7 mm in diameter and the pressure medium is CaF_2 . Preparation of the experimental assemblage required the following steps:

The graphite capsule was packed with starting composition powder and then a thin layer of graphite powder compacted on top. It was heated in a muffle furnace for > 4 hours at 400 °C to ensure that the capsule was as dry as possible. Crushable MgO spacers, the MgO thermocouple sleeve and fused quartz tube were cut to length using a diamond-edged bench saw. The MgO pieces were placed in a muffle furnace at 1000 °C for 24 hours to dry. Once all the pieces had been prepared, including pressure medium sleeves (see below), they were assembled as shown in Figure 3.3. Pb foil was cut to size to wrap around the sides and cover the base of the 12.7 mm diameter assemblage, ready to insert into the pressure vessel.

CaF₂ pressure medium sleeves

 CaF_2 was first baked in a muffle furnace overnight to drive off any water. The CaF_2 was then crushed and sieved to remove any lumps. 6.78 g of CaF_2 was measured out to make each sleeve (three required per experiment). A small press was used to compress the CaF_2 inside the 12.7 mm (1/2-inch) die. The sleeves were sintered at 800 °C for > 30 minutes in a muffle furnace, then removed and placed in a vacuum oven at 120 °C until required.

3.4.2 Experimental procedure

Once the experimental assemblage was ready, the experimental stack was constructed (Figure 3.4). A $W_3Re_{97}-W_{25}Re_{75}$ thermocouple was made using the two wires connected to the piston-cylinder output power controller and four-holed ceramic tube (Holtz, 2001). Zirconium oxide cement was inserted into the ceramic tube holes to protect the wires from oxidation at high experimental temperatures (Moore et al., 2008). The experimental assemblage was greased and loaded into the pressure vessel, before being placed over the main ram at the bottom of the stack. The other stack components were loaded on top of the pressure vessel as shown in Figure 3.4.

Water was set flowing through the two cooling brackets and N_2 gas was piped slowly over the thermocouple wires, again to prevent oxidation of the wires, which could result in temperature fluctuations and ultimately experimental failure. For the 10 kbar experiments, experimental conditions were reached using the cold-piston-in method (Johannes et al., 1971). The pressure of the end load was increased to 1149 bars, and then the main ram was taken up to 51 bars, both read off the analogue pressure dials; equivalent to 10 kbar experienced by the experimental charge. The output program was set running, which ramped the temperature up to experimental conditions at a rate of 60 °C per minute. Pressure was maintained on the main ram during the temperature ramping. Once stable at experimental conditions, the



Figure 3.3 Schematic of the experimental assemblage for 19.1 mm (3/4-inch) piston-cylinder experiments. 10 kbar experiments are 12.7 mm (1/2-inch) in diameter, use CaF_2 sleeves as the pressure medium, and do not contain the short graphite furnace sleeve. The whole assemblage is wrapped in Pb foil and inserted into the pressure vessel.



Figure 3.4 Schematic of the stack experimental set-up for 5 and 10 kbar experiments. Pressure conditions are maintained by the main ram and end load pistons, with dials showing their respective pressures. Temperature is monitored by a $W_3Re_{97}-W_{25}Re_{75}$ thermocouple inserted just above the capsule and controlled by the discharge of the capacitor through the base plug. Oxygen fugacity is held at CCO buffer conditions by the graphite capsule. Water pumped through cooling jackets kept the stack cool.

experiment was checked every few hours, recording conditions and adjusting the pressure of the main ram if necessary.

10 kbar experiments were run for 48 hours to allow time for equilibration between phases, after which they were quenched to room temperature by shutting off the power to the stack capacitor. Extracted capsules were sectioned longitudinally, mounted in epoxy, and polished using silicon carbide papers and water-based diamond suspension down to 1 µm grade.

3.4.3 Pressure and temperature uncertainty

A friction correction had previously been determined for the experimental assemblage at 12 to 14 kbar and 1300 °C using the breakdown of Ca-Tschermakite to the assemblage anorthite, gehlenite, and corundum (Hays, 1966; Bohlen, 1984). It was found to be within the pressure uncertainty of the experiment (\pm 0.5 kbar), therefore no friction correction was applied to reported 10 kbar pressures. The temperature uncertainty on the piston-cylinder apparatus is \pm 5 °C, which comes from the uncertainty on the thermocouple calibration (Holtz, 2001). The thermocouple measures the temperature at the MgO spacer above the graphite capsule, which has been converted into the temperature at the centre of the capsule by calibration with salt melting experiments (Figures 3.3 and 3.4; Siewert et al. 1998; Masotta et al. 2012). The

temperature recorded by the output is different to the experimental temperature due to: (i) the thermocouple wire calibration, which converts electromotive force to temperature by a theoretical relationship, and (ii) the thermal gradient from the MgO spacer to the capsule hotspot (one quarter to one third down from the capsule lid). Based on WHOI experimental lab calibrations, the 12.7 mm (1/2-inch) assembly thermocouple measures a temperature 12 °C cooler than that at the capsule hotspot, giving a thermal gradient across the capsule of 5–10 °C. For the 19.1 mm (3/4-inch) assembly, the thermocouple output measures a temperature 42 °C cooler than that at the capsule hotspot, which translates to a 10–20 °C thermal gradient through the capsule (Glenn Gaetani, pers. comm.). Thermal gradients are the same magnitude as the temperature spacings of the experimental runs, therefore some overlap of experimental conditions is conceivable, which could produce runs of different experimental temperatures that have the same crystalline textures.

3.5 Experiments at 5 kbar

Experiments at equilibration pressures of 5 kbar were conducted using a 19.1 mm (3/4-inch) diameter experimental assemblage and a piston-cylinder apparatus (Figures 3.3 and 3.4; Hudon et al. 1994; Manning and Boettcher 1994). Very few experiments had previously been conducted successfully at this pressure due to difficulties in stabilising the experimental set-up (Moore et al., 2008; Masotta et al., 2012; Iacovino et al., 2013). At lower pressures within a piston-cylinder assemblage, friction and material compressibility has a significant impact upon the pressure experienced by the capsule. Therefore, soft materials are required, particularly for the pressure medium, to get the central capsule to the correct pressure, however these materials can be more unstable at high temperatures. Precision is needed when making the experimental set-up to ensure that the assemblage materials remain in place and behave structurally as they should. The stability of the thermocouple can also be a problem, as it must be held in place just above the capsule for the duration of the experiment, but this is harder at low pressures. The procedure described below was developed over a period of three months to overcome the challenge of stabilising piston-cylinder experiments at high temperature, low pressure for 48-hour runs.

3.5.1 Assemblage preparation

The experimental charge assemblage construction is the same as detailed above for the 10 kbar experiments and as shown in Figure 3.3. Graphite capsules were formed by hollowing out a

6.4 mm graphite rod with a 4.4 mm drill bit, then cutting disks off the same rod to form a graphite wafer lid.

NaCl pressure medium sleeves

NaCl was first baked in a muffle furnace overnight to dry it out. The salt was then crushed and sieved to remove any lumps. 4.08 g of NaCl was measured out for each salt sleeve. A small press and the 19.1 mm (3/4-inch) diameter die were used to compress the salt at 4500 psi for one minute. The salt sleeves were annealed in the muffle furnace by ramping up 100 °C per hour from 110 °C to 700 °C, held at 700 °C for 10 hours, and then ramped down at the same rate back to 110 °C. They were then stored in a vacuum oven at 120 °C.

3.5.2 Experimental procedure development

Prior to this study, very few experiments had been conducted at crustal pressures on basaltic compositions, and none were from Iceland. At WHOI, only one set of experiments had been successfully conducted at < 10 kbar using a 19.1 mm assemblage with NaCl pressure medium. However, recent developments to experimental procedures at other institutions had shown that mid-crustal pressures could be achieved consistently using a piston-cylinder apparatus (Moore et al., 2008; Masotta et al., 2012). The success of other authors provided the motivation to develop an experimental procedure to stabilise 5 kbar in a piston-cylinder set-up, if achieved then the experimental results could be an important dataset for igneous barometer recalibration.

The first experiments conducted at 5 kbar used a basaltic glass composition from Borgarhraun, Northern Volcanic Zone (sample FH9845). Once the experimental procedure had been developed enough that the probability of success began to improve, then the Miðfell starting composition was used for experiments at 5 kbar.

Initial experimental procedure

Experiment FH9845-42 was prepared as detailed above. The cold piston in method was used with a 20% overpressure to achieve experimental pressure conditions (Johannes et al., 1971). Previous 5 kbar experiments in the WHOI lab had been over-pressured to stabilise the low pressure assemblage, therefore the initial procedure for this set of experiments used an overpressure. The temperature was then ramped up at a rate of 60 °C per minute by the output power program until 1298 °C was displayed on the output screen. This display temperature, as measured by the thermocouple, is equivalent to 1340 °C within the centre of

the capsule. The experiment deviated from experimental conditions after one minute, causing the output power controller to turn off power to the experiment, i.e. the experiment failed.

Cracking was heard as the assemblage was initially pressurised, suggesting that the fused quartz tube, separating the pressure medium from the graphite furnace, had fractured/broken. Maintaining the integrity of the quartz tube was initially thought to be the way to stabilise the experiment. The method of approaching pressure-temperature conditions was altered by ramping both pressure and temperature together to give the quartz tube time to soften, and hopefully achieve stability.

Development: hot piston in

Experiment FH9845-43 followed the hot piston in method. Temperature was ramped up at a rate of 20 °C per minute to give the quartz tube time to soften, but with two-minute dwells every 260 °C, allowing the pressure on the main ram to be increased by 12 Pa ($\sim 20\%$ total main ram pressure). Experimental conditions were stabilised, but the experiment failed overnight with deviation alarm, over-temperature and over-power lights illuminated. Temperature had been observed to be oscillating by ± 5 °C during the experiment, suggesting a problem with the thermocouple, however N₂ gas had been flowing over the wires to prevent oxidation.

Experiment FH9845-44 was conducted using the same procedure as FH9845-43, with much the same result. Temperature was not being maintained causing oscillations to increase out of control until failure of the experiment. This power oscillation suggested that the Proportional-Integral-Derivative (PID) controller settings, which set the output power to the experiment and maintain the temperature, were incorrect for the 19.1 mm experimental assemblage. The PID controller was auto-tuned on the red press piston-cylinder to get the PID settings required to maintain temperature during each run of the 19.1 mm experimental assemblage. Experiment MID2014-01, the first conducted at 5 kbar using the Miðfell starting composition, was set running with the new PID settings and using the same procedure as FH9845-44. After reaching stability, the experiment failed overnight with deviation alarm, over-temperature and over-power lights illuminated.

Comparison of the assemblage materials used in experiment FH9845-44 to published experimental assemblages that were successful at 5 kbar suggested that other materials could be used for the 19.1 mm set-up. Modifications based on the material used by other authors were made to subsequent trial procedures (Hudon et al., 1994; Laporte et al., 2004; Lambart et al., 2009a).

Development: different assemblage materials

Experiment MID2014-04 used a Pyrex glass (borosilicate) tube instead of fused quartz tube to test whether the different tube material would not crack during pressurisation and stabilise the assemblage, as used in the assemblages of other experiments (Hudon et al., 1994; Laporte et al., 2004; Lambart et al., 2009a). The same procedure was used as for MID2014-01, however it failed after two minutes at PT conditions.

KCl salt sleeves were used in experiment FH9845-45, with PID settings re-tuned to cope with the different pressure medium. Fused quartz was again used and the same procedure as MID2014-01 was followed. The experiment failed just before experimental temperature was reached. Experiment FH9845-46 was also conducted with KCl and the same method as FH9845-45, and also resulted in failure.

A pyrophyllite wafer replaced the MgO wafer above the graphite capsule in experiment FH9845-48 as it was considered that the MgO wafer contacting with the thermocouple at low pressures was causing temperature measurement problems (Figure 3.3). FH9845-48 was conducted using the same procedure as MID2014-01, however the experiment failed while ramping up to PT conditions. The hydrous nature of pyrophyllite could have affected the oxygen fugacity of the experimental assemblage, potentially creating more oxidised conditions that could damage the integrity of the thermocouple.

Alternative materials were not improving the stability of the assemblage, suggesting that something more fundamental needed to be changed to the experimental procedure. As experiments at higher pressure are known to be stable, it was decided to go back to over-pressuring the assemblage and return down to the desired pressure as experimental temperature was reached.

Development: over-pressuring the assemblage

Moore et al. (2008) suggest that low pressure experiments can fail due to incomplete compression of the experimental assemblage and problems with the thermocouple. Therefore over-pressurising the experimental assemblage could stabilise the experiment before returning back down to equilibration pressure. Nelson and Montana (1992) showed with their set of experiments that pressure could be reduced during an experimental run to change the pressure of equilibration, and in their case induce plagioclase melting.

Experiment FH9845-47 used the ramping up PT method with overpressure of $\sim 20\%$, however failure occurred before PT conditions were reached. Experiment FH9845-49 had an overpressure of 100% on the main ram (115 bar), and was then returned to 58 bar for the final 300 °C increase up to experimental temperature. The same temperature ramping rate with

dwell times as MID2014-01 was used. The experiment remained stable and quenched at the end of the 48 hour program. A repeat of the same experimental procedure as FH9845-49 for experiment FH9845-50 resulted in failure upon pressure reduction after over-pressuring. The power output increased, which increased the assemblage temperature. As a result, the power output dropped in response to measuring a high temperature, which caused the temperature to drop to failure.

It was thought that FH9845-50 failed due to the overpressure not being high enough, or the pressure reduction from overpressure occurred too soon. The large changes in output power and fluctuations in temperature for [FH9845-50] led to the conclusion that the thermocouple was feeding back varying temperatures to the output controller. It was concluded that during the successful run the overpressure had caused the base plug to collapse around the thermocouple and hold it in place (Figure 3.4). During the failed runs, the thermocouple was not held in place, allowing it to move vertically in a large temperature gradient. Movements would be translated into different measured temperatures, which the output controller would respond to by altering the output power. If the oscillations were too large then the experiment would fail. Collapsing of the base plug to maintain a constant position for the thermocouple was key to the stabilisation of these 5 kbar runs.

Using this new assertion that the base plug needed to be collapsed to stabilise conditions, experiment FH9845-51 was run successfully using the same procedure as FH9845-50. Experiment MID2014-30 failed as the pressure was dropping slowly from 64 bar to 56 bar (main ram) over the course of the experiment. Experiment FH9845-53 was stable for 45 hours before an over temperature lead to failure. Experiments MID2014-31, MID2014-32, MID2014-34, and MID2014-35 were all successful. MID2014-33 failed after 24 hours.

3.5.3 Final experimental procedure

The experimental procedure with the highest number of successful runs was as follows. The end load pressure required for the experiment was loaded and ~ 6 bars onto the main ram to maintain a good connection between the capacitor stack, the base plug, and the graphite furnace (Figure 3.4). Temperature was ramped up at a rate of 20 °C per minute, with two-minute dwells at every 260–300 °C increase in temperature. Once above 900 °C the assemblage was over-pressured to 10 kbar (115 bars, main ram). As temperature was increased again, the overpressure was reduced by 10 bars on the main ram every 50 °C until experimental conditions were reached. N₂ gas was piped over the thermocouple wires to prevent oxidation. All spacers were made from crushable MgO. NaCl was used as the pressure medium.

3.5.4 Pressure and temperature uncertainty

Salt melting experiments were conducted to test the equilibration pressure of the experimental assemblage after PT conditions were achieved using the method described above. This method of pressure calibration uses pairs of experiments that bracket the NaCl liquidus at a number of experimental pressures and the known PT relationship for the NaCl melting curve (Bohlen, 1984; Siewert et al., 1998; Masotta et al., 2012). The texture of the salt within the capsule after quenching reveals whether the salt melted or not. The salt melting experiments conducted at 5 kbar pressure all failed. However, it can be argued that as the pressure medium is molten then a reduction in pressure by bleeding the main ram should reduce the assemblage pressure. It is uncertain whether the capsule at the centre decompresses to the same extent, hence experimental confirmation is required. The temperature uncertainty from the thermocouple is ± 5 °C (Holtz, 2001).

3.5.5 Na-gain to the starting bulk composition

After SEM imaging and EPMA of the 5 kbar experiments, it became apparent that these capsules had gained Na₂O during the experimental run. Major element oxide concentrations ranged 4.5–8.0 wt% Na₂O, 3.0–6.5 wt% higher than the starting bulk composition (Table 3.1). Experiment MID2014-32 had the highest Na₂O EPMA measurement and also showed visual evidence of NaCl salt crystals under SEM. Figure 3.5 shows the NaCl crystals present in the centre of the experimental capsule, clearly indicating infiltration of the NaCl pressure medium. Energy-dispersive X-ray Spectroscopy (EDS) on the SEM confirmed the composition of the NaCl crystals present in this charge. While not all of the 5 kbar experiments show such clear NaCl crystals, the elevated Na₂O EPMA measurements suggest that all four experiments have been affected to varying extents by Na-gain.

In conclusion, care must be taken when using the results of these experimental runs, as although the experimental PT and oxygen fugacity conditions are known, the bulk composition is now an unknown.

3.5.6 Further refinement to the experimental procedure

Eliminating the need to overpressure

The likely cause of failure for the majority of the initial 5 kbar experiments was due to movement of the thermocouple. Vertical movements of the thermocouple within the thermal gradient at the edge of the graphite capsule feedback temperature fluctuations to the output power controller, which responds by altering the output power and changing the temperature



Figure 3.5 (a) Electron back scatter and (b) QEMSCAN images of experiment MID2014-32, showing crystalline NaCl present within the centre of the experimental charge. 1 mm white bar for scale.

further (Figure 3.4). Sometimes temperature oscillates out of control to the point of failure. Over-pressurising the experimental assemblage allowed the base plug to collapse around the thermocouple, holding it in place at a constant height above the capsule. A specialist base plug that has a small ridge around the thermocouple has been developed to eliminate the need for over-pressurising the experimental assemblage.

Preventing Na-gain

NaCl pressure medium is molten at equilibration temperatures used for the 5 kbar experiments, allowing for easy diffusion of Na⁺ and Cl⁻ ions through the experimental assemblage and causing Na-gain to the starting bulk composition. This problem could be overcome by using a double capsule arrangement, where the graphite capsule containing the starting material is surrounded by a noble metal capsule (e.g. Pt). The presence of a sealed noble metal outer layer to the capsule would prevent diffusion of Na⁺ into the basaltic experimental material; it would also reduce H₂O exchange across the capsule walls. This technique has been used by other authors experimenting with a NaCl pressure medium (Manning and Boettcher, 1994; Laporte et al., 2004; Lambart et al., 2009a,b; Iacovino et al., 2013).

Calibration of experimental pressure

Successful salt melting experiments are required to check the equilibration pressure of the experimental procedure, particularly as the capsule might not have returned back to 5 kbar after being over-pressured (Siewert et al., 1998; Baker, 2004; Masotta et al., 2012). However, if the experimental procedure can be developed without over-pressurising the assemblage, then previous experiments suggest that the use of NaCl pressure medium requires no frictional pressure correction (Boettcher et al., 1981).

Chapter 4

Crystallisation experiments using the Miðfell glass composition

4.1 Introduction

Igneous barometry can provide information about the pressures, and therefore depths, of melt crystallisation within a magmatic system and can in consequence be used to infer the depths of magma reservoirs. One of the potential applications of this is the ability to relate magma reservoir depth to the modes of volcanic activity observed at the surface. Seismicity, ground deformation, and volatile degassing are all surficial expressions of magma storage and transport within the crust (Edmonds, 2008; Sigmundsson et al., 2010; Tarasewicz et al., 2014). The use of such observations, along with knowledge of magma reservoir depths, has the potential to aid volcanic eruption forecasting, particularly in predicting the eruption style and identifying the potential hazard of such eruptions (McNutt, 1996; Greenfield et al., 2016; Geiger et al., 2016).

Igneous barometers

Several types of igneous barometer are employed by petrologists to estimate equilibration pressures within a volcanic system, with the majority of these pressures assumed to represent magma reservoir depths. One set of barometers utilises the pressure sensitivity of volatile solubility in silicate melt (Newman and Lowenstern, 2002; Moore, 2008; Shishkina et al., 2010; Witham et al., 2012). Assuming that measured samples were saturated in volatiles, then observed volatile concentrations can be converted to saturation pressures. This method has been used on olivine-hosted melt inclusions to estimate entrapment pressures, however processes such as post-entrapment crystallisation and olivine decrepitation can modify

the dissolved volatile content of these inclusions, resulting in lower pressure estimates (Danyushevsky et al., 2002; Bucholz et al., 2013; Moore, 2014; Maclennan, 2017).

Barometry methods that involve equilibria between crystallised silicate phases are limited due to the low number of minerals crystallised from mafic silicate melts. Assemblages of olivine \pm plagioclase \pm clinopyroxene \pm spinel dominate mafic igneous rocks, so endmember equilibria are restricted to those compositions (Grove et al., 1992). However, recent developments of activity-composition (*a*-*x*) models for mafic phases have enabled a new barometer to be created using the Holland and Powell (2011) thermodynamic dataset, which uses the compositions of equilibrium silicate phases to estimate pressure (Jennings and Holland, 2015; Green et al., 2016; Ziberna et al., 2017).

Olivine-plagioclase-augite-melt (OPAM) barometry also requires equilibrium between igneous phases, but pressure is estimated using the equilibrium composition of melt coexisting with these phases. Several OPAM barometer parameterisations make use of the pressure sensitivity of the olivine-plagioclase-augite ternary point liquid composition (Yang et al., 1996; Kelley and Barton, 2008; Voigt et al., 2017).

The most useful pressure dependent chemical exchange for barometry of mafic igneous systems is the incorporation of jadeite (Jd; NaAlSi₂O₆) within clinopyroxene, which has been known for decades within the igneous petrology community (Putirka et al., 1996; Holland and Powell, 1998). The original calibration of this clinopyroxene-liquid barometer used experiments conducted on basalt at pressures of 0.001 kbar and > 7 kbar, with subsequent re-calibrations aiming to improve barometer reliability (Putirka et al., 1996; Putirka, 1999; Putirka et al., 2003; Putirka, 2008b). Testing of earlier calibrations, using assemblages equilibrated at known conditions, has shown that pressures < 7 kbar are overestimated, which is concerning as these pressures are arguably those of most interest to igneous petrologists (Neave and Putirka, 2017). However, improvements to experimental procedures, such as reducing Na-loss in gas-mixing furnaces, have produced low pressure datasets that are more reliable, and useful for barometer calibration (Yang et al., 1996; Berndt et al., 2002; Moore et al., 2008). These improvements, along with careful selection of calibration data, have enabled the development of an improved clinopyroxene-liquid barometer (Neave and Putirka, 2017).

Use of igneous barometers in Iceland

Igneous barometry has been a key component of volcanic system interpretation in Iceland over the past few decades, with clinopyroxene-liquid barometry providing lower crustal magma chamber estimates for Borgarhraun, which were further supported by OPAM estimates (Yang et al., 1996; Maclennan et al., 2003b; Winpenny and Maclennan, 2011; Maclennan et al.,
2012). Petrological observations from more recent eruptions, such as those at Eyjafjallajökull and Holahraun, have been able to estimate magma storage depths and link them to the observed eruption style and behaviour (Keiding and Sigmarsson, 2012; Geiger et al., 2016). The plumbing systems of the southern volcanoes Gírmsvötn-Laki and Katla have also been investigated using igneous barometry to constrain magma chamber depth within the crust (Neave et al., 2013, 2015; Budd et al., 2016). Despite this interest in using igneous barometry on Icelandic eruptions, very few experiments have been conducted on Icelandic basalt compositions (cf. Maclennan et al. 2012; Neave et al. 2017).

This chapter presents experiments conducted on basaltic glass from the Miðfell eruption to augment the dataset of Icelandic experimental compositions; methods are described in Chapter 3. It was originally hoped that the results of these experiments would lead to recalibration of the clinopyroxene-liquid barometer, however this was prevented by: (i) not being able to conduct enough experiments, and (ii) modification of starting bulk composition during 5 kbar experiments. Instead the experiments presented here provide a dataset of compositions to test the latest parameterisations of igneous barometers, namely clinopyroxene-liquid, OPAM, and multi-reaction average pressure (Yang et al., 1996; Kelley and Barton, 2008; Neave and Putirka, 2017; Voigt et al., 2017; Ziberna et al., 2017). These experiments can also be used to test some of the crystallisation models that are often used to interpret the crystallisation history of magmatic systems; Petrolog and MELTS (Ghiorso and Sack, 1995; Asimow and Ghiorso, 1998; Danyushevsky and Plechov, 2011). This chapter concludes by applying the tested igneous barometers to natural samples from Mðfell, the eruption that will be the main focus of subsequent chapters.

4.2 Petrology

This section presents petrological observations from crystallisation experiments carried out on the Miðfell starting bulk composition (Table 3.1). Thirty-five experiments were conducted, of which, 27 reached the end of their run time at stable pressure-temperature conditions and were therefore considered successful. Experiments were equilibrated at three pressures; 0.001 kbar, 5 kbar, and 10 kbar, with each requiring a slightly different experimental procedure as detailed in Chapter 3.

Experimental charges have been imaged using SEM and QEMSCAN to provide textural observations of the mineral assemblages (Figures 4.1–4.3). EPMA was used to measure phase compositions, which are presented along with calculated phase proportions and a consideration of the mineral-melt partitioning behaviour recorded by each experiment (Table 4.1).



Figure 4.1 SEM images of 0.001 kbar experiments (scale bar 100 μ m). In order of decreasing equilibration temperature: (a) MID2014-16, 1245 °C; (b) MID2014-11, 1240 °C; (c) MID2014-07, 1230 °C; (d) MID2014-18, 1225 °C; (e) MID2014-13, 1220 °C; (f) MID2014-08, 1210 °C; (g) MID2014-24, 1205 °C; (h) MID2014-20, 1200 °C; (i) MID2014-09, 1190 °C; and (j) MID2014-12, 1170 °C.

4.2.1 Run products

Experiments at 0.001 kbar

Figure 4.1 shows SEM and QEMSCAN images of experimental assemblages equilibrated at 0.001 kbar. These images show that the proportion of crystallised phases increases as equilibration temperature decreases, with silicate phases arriving on the liquidus in the order: olivine (1250–1245 °C), plagioclase (1225–1220 °C), and then clinopyroxene (1205–1200 °C). Cr-spinel can be seen in some of the higher temperature experimental charges (Figures 4.1c,d & f), but is too small to be measured by EPMA (< 20 μ m). Its composition was confirmed by energy-dispersive X-ray spectroscopy (EDS) using SEM. No Cr-spinel was observed in any of the other 0.001 kbar experiments, and therefore was assumed not to be a significant crystallising phase in this suite of experiments.

The grain length of olivine and plagioclase is > 50 μ m for equilibrium temperatures 1245–1220 °C (Figures 4.1a–e), however below this temperature the mineral assemblages become much finer-grained (< 50 μ m), suggesting that the assemblage is being controlled by a higher nucleation rate (Figures 4.1f–j). As the equilibrium temperature gets further below the bulk liquidus, the assemblage texture is increasingly dominated by higher nucleation rates. All the 0.001 kbar experiments were run for 24 hours, which is only long enough for a limited amount of Ostwald ripening to occur, hence small grain sizes are preserved in the lower temperature experiments. These charges preserve textures akin to fast cooling rates in basaltic lavas, where large steps of temperature under-cooling below the liquidus result in high nucleation rates.

In the low temperature assemblages the plagioclase grains become more needle-like, which is likely the result of faster growth where morphology is controlled by the rate of diffusion to growth surfaces. The highest crystallinity charges, equilibrated < $1200 \,^{\circ}$ C, do not show any clear textural evidence for zonation or fractionation within the experiment. The fine-grained phases appear to be randomly distributed, however due to the fragmentation of the glass charge upon quenching, information regarding whole charge structure has been lost. Textural information cannot be gleaned from these charges and related to the thermal structure of the experimental set-up, however it is assumed that thermal gradient effects are minimal for a gas-mixing furnace.

Glass and crystalline phase EPMA oxide data was measured from phases coexisting together locally in the assemblage. This method was employed to minimise the effect of potential disequilibrium and chemical zonation in the higher crystallinity experiments. Therefore, mineral phase compositions can be assumed to be in equilibrium with each other, giving meaning to any partition coefficients recovered from the data.



Figure 4.2 Full capsule SEM and QEMSCAN images (left; scale bar 1 mm), and an enlarged SEM image (right; scale bar 200 μ m) of 5 kbar experiments. In order of decreasing equilibration temperature: (a) MID2014-31, 1250 °C; (b) MID2014-32, 1230 °C; (c) MID2014-35, 1210 °C; and (d) MID2014-34, 1190 °C. QEMSCAN colours: light green, olivine; dark green, clinopyroxene; blue, plagioclase; peach, glass.

Experiments at 5 kbar

Experiments conducted at 5 kbar again show an increasing degree of crystallinity as equilibration temperature decreases (Figure 4.2). Only four 'successful' 5 kbar runs were achieved, hence the equilibrium temperatures are separated by 20 °C intervals and there is no super-liquidus experiment. The arrival of silicate phases onto the liquidus can be loosely constrained by the experiments detailed here, with phases crystallising in the order: olivine (> 1250 °C), then clinopyroxene (1230–1210 °C), and then plagioclase (1210–1190 °C). No Cr-spinel was observed in any of the 5 kbar experiments, which could be the result of the non-primitive nature of the starting bulk composition.

In the high temperature experiments (Figure 4.2a–b), olivine is mainly found at the side of the capsule and the very centre. This distribution of phases could be the result of the temperature gradient across the capsule, with slightly lower temperatures just below the centre of the capsule and towards the bottom. Or perhaps the distribution reflects the ability of olivine to nucleate heterogeneously on the side of the capsule. These high temperature charges also show the largest grain sizes, with some olivine crystals > 100 μ m long. As equilibrium temperature decreases, so too does the average grain size, which is < 50 μ m when clinopyroxene and plagioclase arrive on the liquidus. The lowest equilibrium temperature experiment shows a highly crystalline texture with glass only present around the edge of the capsule (Figure 4.2). Due to the high crystallinity of this charge it could be expected that the phases would show compositional variability from the centre of the charge out towards the rim. However, such variability is not clear from the SEM images due to the small grain-size of the phases and the similarity in greyscale between olivine and clinopyroxene.

The good coverage of glass across the three higher temperature experiments enabled phases to be analysed by EPMA in areas where they were all coexisting. It was hoped that this method of data acquisition would provide mineral compositions that were in equilibrium with each other as they had experienced similar crystallisation temperatures within the charge. The high crystallinity of the lowest temperature experiment limited the glass measurements that could be made. Crystalline phases were measured from rim to core of the capsule, but no systematic variation in crystal composition could be seen. Equilibrium between phases in the core and glass at the rim cannot be assumed for this capsule.

In some of the charges NaCl crystals are visible using SEM and QEMSCAN, which suggests that some of the NaCl pressure medium, used to contain the experimental assemblage, leached through the graphite capsule and into the experimental material. The presence of NaCl crystals further suggests that the capsule was saturated in NaCl to enable grains to crystallise, therefore it is likely that there has been significant addition of NaCl to these experimental charges (MID2014-32).



Figure 4.3 Full capsule SEM and QEMSCAN images of 10 kbar experiments (scale bar 1 mm). In order of decreasing equilibration temperature: (a) MID2014-25, 1310 °C; (b) MID2014-10, 1300 °C; and (c) MID2014-22, 1290 °C. QEMSCAN colours: light green, olivine; dark green, clinopyroxene; blue, plagioclase; peach, glass.



Figure 4.3 continued Full capsule SEM and QEMSCAN images of 10 kbar experiments (scale bar 1 mm). In order of decreasing equilibration temperature: (d) MID2014-14 (and MID2014-29), 1280 °C; (e) MID2014-27, 1270 °C; and (f) MID2014-23, 1240 °C. QEMSCAN colours: light green, olivine; dark green, clinopyroxene; blue, plagioclase; peach, glass.

Experiments at 10 kbar

The experimental charges equilibrated at 10 kbar do not follow the clear textural trend of crystallinity increasing as temperature decreases seen in the 0.001 and 5 kbar experiments. The highest temperature assemblage, MID2014-25 (1310 °C, Figure 4.3a), contains a large clinopyroxene grain > 1 mm wide containing several small olivine crystals (< 200 μ m) situated at the bottom of the capsule. This assemblage contrasts with the lower temperature assemblage of MID2014-10, which equilibrated 10 °C cooler and contains only 100 μ m long olivine grains along the base of the capsule (Figure 4.3b). MID2014-22 (1290 °C, Figure 4.3c) is again coarsely crystalline, bearing a resemblance to the texture seen in

the highest temperature assemblage, showing a 200 μ m wide olivine grain encased in a clinopyroxene grain > 500 μ m wide. These are the only two crystals present in the capsule, suggesting low nucleation rates favouring crystal growth. Given the likely temperature gradient across the capsule, it is conceivable that these phases have all equilibrated at lower temperatures than the set experimental temperature. However, all phases have crystallised at the same place in the capsule, suggesting that they were all exposed to the same point along the capsule temperature gradient and therefore remain offset in equilibrium temperature. For these three experiments, major element oxides have been measured for the clinopyroxene cores and rims, olivine cores, and the glass near the clinopyroxene faces. The olivine could be out of equilibrium with the glass due to the surrounding clinopyroxene.

Two experiments were equilibrated at 10 kbar and 1280 °C, MID2014-14 and -29, which provides a way of testing the reproducibility of experimental conditions by piston cylinder apparatus. Both experiments show distinct layering within the capsule, which is texturally different to experiments equilibrated at higher temperatures (Figure 4.3d). At the top of both capsules is just glass, below is an assemblage of glass and clinopyroxene forming grains < 100 μ m, and below that glass, olivine and clinopyroxene coexist. The bottom layer is almost completely crystalline, but with plagioclase present in the assemblage as ~ 400 μ m long needles, surrounded by, and trapping olivine and clinopyroxene grains. The patches of glass in the lower section of the capsule are too small to be measured by EPMA. The structure of these charges may be the result of the thermal gradient across the capsule, with highest crystallinity at the base where temperatures were lowest. Glass was only measured in the uppermost section of this capsule, while all crystalline phases were measured within each of the layers defined above (Figure 4.3d).

The texture of MID2014-27 resembles that of a sub-solidus assemblage, with a completely crystalline mixture of olivine, clinopyroxene and plagioclase crystals, all < 50 μ m in size (Figure 4.3e). The small patches of glass that can be found are again too small for EPMA, however the other phases could be measured across the capsule. There is an apparently homogeneous distribution of these phases through the capsule, with no obvious layering as seen in assemblages equilibrated 10 °C hotter, apart from a glass and clinopyroxene only assemblage at the top.

MID2014-23 (1240 °C) again shows distinct layering within the capsule, similar to that of MID2014-14 (and -29), however the plagioclase crystals are coarser in this capsule; $\sim 400 \text{ }\mu\text{m} \log \text{ and} \sim 50 \text{ }\mu\text{m} \text{ wide}$ (Figure 4.3f). Olivine and clinopyroxene crystals remain fine-grained < 50 µm, and more glass appears to be present compared to MID2014-27. The difference in texture between MID2014-23 and MID2014-27 suggests that the latter has equilibrated at a lower temperature than the former, or perhaps a higher pressure. MID2014-



Figure 4.4 Pressure-temperature phase diagram for Miðfell basaltic glass based on experimental equilibrium assemblages. Colour indicates pressure of equilibration; 0.001 kbar, purple; 5 kbar, red; 10 kbar, blue. Shape indicates phases present; diamond, glass (gl) only; circle, gl + olivine (ol); triangle, gl + ol + plagioclase (pl); inverted triangle, gl + ol + clinopyroxene (cpx); square, gl + ol + pl + cpx.

27 does not fit into the sequence of textural observations unless it is placed as the lowest temperature assemblage, which is not what was programmed into the piston cylinder output power computer when running this experiment. If temperature was held correctly for the duration of the experiment, then perhaps the pressure was higher than required. The run sheet for this experiment suggests that pressure was held at 10 kbar for the duration, along with the temperature and output power remaining constant. For this experiment, glass was measured in the two uppermost layers while crystalline phases were measured in layers two and three (Figure 4.3f).

As with the 5 kbar experiments, no Cr-spinel was observed in any of the 10 kbar assemblages, again suggesting that it was not a dominant crystallising phase for this starting bulk composition.

Pressure-temperature phase diagram

The equilibrium phases described above can be used to plot a pressure-temperature phase diagram to show the conditions at which certain phases appear on the liquidus (Figure 4.4). The 0.001 kbar experiments constrain the temperatures at which olivine, plagioclase, and clinopyroxene each arrive on the liquidus to within 5 $^{\circ}$ C. The larger temperature spacing

of the 5 kbar experiments brackets the clinopyroxene and plagioclase liquidi to within $20 \,^{\circ}$ C, however the olivine liquidus only has a lower temperature limit as no super-liquidus experiment was conducted.

The 10 kbar experiments do not show the same clear progression of crystallised phases as the other two pressures, however if MID2014-25 (1310 °C) is ignored then the phase liquidi can be estimated as shown in Figure 4.4. The phase diagram suggests that all three phases appear on the liquidus at similar temperatures, as shown by the association of olivine and clinopyroxene for the high temperature assemblages (Figures 4.3c–d). Justification for olivine arriving on the liquidus before clinopyroxene comes from: (i) the textural observation that olivine grains are trapped within clinopyroxene crystals in MID2014-25 and MID2014-22 (Figures 4.3a & c); and (ii) the 10 kbar experiment with lowest crystallinity is olivine only (MID2014-10; Figure 4.3b). The non-primitive nature of the starting bulk composition suggests that it has already undergone significant olivine crystallisation, therefore it is unsurprising that all three silicate phases appear on the liquidus over such a small temperature range.

The complete diagram shows that the liquidus (ol-in) increases in temperature with pressure, with the liquidi of the other phases also increasing in temperature. The order of crystallisation also changes with increasing pressure, with plagioclase appearing before clinopyroxene at 0.001 kbar, but the reverse is true for pressures > 5 kbar.

4.2.2 Glass compositions

Experimental glass compositions have been measured by EPMA, with measurements corrected for each session based on analyses of Smithsonian basaltic glass 113498-1; correction details presented in Appendix A (Jarosewich et al., 1980). This correction altered the SiO₂ content of the experimental glasses and was necessary to minimise the effects of inter-session instrumental drift. MID2014-21 was equilibrated at 10 kbar and 1420 °C to provide a homogeneous glass sample of the starting bulk composition. This experimental charge was measured during each EPMA session to provide another standard for checking measurement reproducibility. After SiO₂ corrections were applied, repeat analyses of MID2014-21 showed that SiO₂ content still varied by ~ 0.5 wt% between sessions, indicating that there is still some residual systematic error in the glass SiO₂ measurements. Figure 4.5a shows that the five super-liquidus experiments (MID2014-02, -03, -05, -06, and -28) are offset to higher MgO and SiO₂ than the bulk composition (also measured by EPMA). This offset is likely a systematic error in SiO₂ quantification for these glass-only experiments, as it is unlikely that a phase has been missed during SEM imaging, which could have crystallised to enrich the remaining liquid in MgO and SiO₂.

ass magnesium number (Mg# ^{Liq}),	t# ^{Cpx}). Compositional variability	and solution fit. * <i>Cr-sp present</i> .
age compositions of the phases present.	nd clinopyroxene magnesium number (each mineral phase, an estimate of Na lo
tions for successful runs, along with aver), plagioclase anorthite content (X_{An}^{Pl}) , a	harge given by σ . Modal proportions of ϵ
Table 4.1 Experimental condit	olivine forsterite content (X_{Fo}^{Ol})	within each experimental is ch

	χ^2 fit	30.1	21.8	15.9	4.4	7.1	6.7	7.1	5.4	8.6	5.0	6.0	6.3	na	4.8	3516	3806	1454	253	0.0	4.2	1.8	1.2	2.4	1.6	na	na	3.5
: (wt%)	$\Delta Na_2 O$	-0.49	-0.41	-0.34	-0.14	-0.21	-0.19	-0.22	-0.15	-0.23	-0.16	-0.18	-0.18	na	-0.16	5.51	5.63	3.54	1.49	0.00	-0.02	-0.01	-0.08	-0.08	-0.03	na	na	-0.10
Δoxide	ΔFeO	0.13	0.33	0.15	0.05	-0.04	-0.14	0.04	-0.08	-0.09	-0.10	-0.04	-0.21	na	-0.10	-3.09	-4.73	-3.36	-0.44	0.00	0.11	0.02	-0.08	-0.33	-0.07	na	na	0.12
(%)	Cpx	ı	ı	ı	ı	ı	I	ı	ı	ı	I	ı	1.3	na	19.8	1	I	18.1	25.6	ı	ı	21.0	ı	4.3	24.7	na	na	32.9
ortions	Ρl	ı	ı	I	ı	I	I	I	ı	1.6	6.8	10.2	13.2	na	20.0	1	I	ı	16.3	ı	I	I	I	I	5.9	na	na	17.1
lal prop	0I	ı	ı	ı	ı	0.5	1.7	2.7	3.1	4.1	6.4	7.6	8.8	na	11.2	4.6	7.4	5.5	9.2	1	ı	0.9	0.6	0.5	2.3	na	na	5.4
Mod	ß	100	100	100	100	99.5	98.3	97.3	96.9	94.2	86.8	82.2	76.7	na	49.0	95.4	92.6	76.3	48.9	100	100	78.1	99.4	95.2	67.0	na	na	44.6
x	ь	I	ı	ı	ı	ı	ı	ı	ı	ı	ı	ı	1.1	na	13.8	1	ı	3.3	4.2	1	ı	0.7	ı	0.4	1.4	na	2.6	5.8
Cp	Mg#	ı	ı	ı	ı	ı	ı	ı	ı	ı	ı	ı	86.3	na	71.0		ı	83.7	74.7	'	ı	86.0	ı	85.6	83.8	na	79.9	81.3
60	ь	ı	ı	ı	ı	ı	ı	ı	ı	1.2	0.6	2.5	1.5	1.9	2.0		ı	ı	3.4	1	ı	ı	ı	ı	1.4	0.9	1.0	1.3
Pla	\mathbf{X}_{An}	ı	I	ı	I	I	I	I	I	88.7	87.4	85.7	86.1	83.2	82.6	ı	I	I	71.4	ı	ı	ı	ı	ı	80.0	79.2	75.9	78.2
	ь	ı	ı	ı	ı	0.1	0.2	0.2	0.2	0.1	2.6	0.1	0.2	0.2	0.2	0.6	0.6	1.1	2.0	1	ı	0.2	0.2	0.0	0.7	0.4	0.5	1.0
Õ	\mathbf{X}_{Fo}	ı	ı	ı	ı	87.2	87.2	86.6	86.4	86.0	84.4	84.5	83.8	82.0	79.9	89.6	91.8	88.7	79.2	1	ı	85.9	86.4	85.8	80.8	81.0	72.6	76.8
	Fe^{3+}	4.1	4.1	4.1	4.2	4.2	4.2	4.2	4.2	4.2	4.3	4.3	4.4	4.4	4.3	7.2	7.2	7.4	6.7	6.4	6.1	5.8	6.2	6.1	5.7	5.6	na	5.4
Glass	ь	0.4	0.5	0.5	0.6	0.6	0.5	0.5	0.6	0.4	0.6	0.6	0.6	1.7	0.5	0.6	0.6	1.2	0.8	0.5	0.4	0.4	0.4	1.1	0.5	0.3	na	1.8
	Mg#	69.3	68.7	69.2	69.1	68.5	68.1	67.0	66.8	66.2	64.5	63.5	62.7	60.1	57.0	74.4	78.8	71.6	56.8	69.8	70.0	62.3	69.7	69.7	61.0	61.1	na	52.5
ions	T (°C)	1310	1290	1270	1250	1245	1240	1230	1225	1220	1210	1205	1200	1190	1170	1250	1230	1210	1190	1412	1320	1310	1300	1290	1280	1280	1270	1240
Condit	P (kbar)	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	S	5	5	5	10	10	10	10	10	10	10	10	10
	Experiment	mid02	mid03	mid05	mid06	mid16	mid11	mid07*	mid18*	mid13	mid08*	mid24	mid20	mid09	mid12	mid31	mid32	mid35	mid34	mid21	mid28	mid25	mid10	mid22	mid14	mid29	mid27	mid23

The lack of a precious metal capsule in the piston-cylinder experimental assemblages suggests that it would be possible for volatiles, namely H_2O and CO_2 to diffuse into the nominally volatile-free starting bulk composition. If this process had occurred, then it could be expected that the EPMA glass totals would reflect the presence of an unmeasured oxide, i.e. recover low oxide totals, as volatiles are incompatible in mafic silicate phases. The 0.001 kbar experiments were conducted in a gas-mixing furnace and are degassed, therefore they can provide a comparison for the piston-cylinder experiments. Glass major element oxide totals are generally > 99%, which is the expected recovery for University of Cambridge silicate glass EPMA data. There is no difference between the 0.001 kbar and piston-cylinder experiment totals, further suggesting that there has not been considerable volatile-gain to the starting bulk composition during the piston-cylinder runs. This observation is important, as addition of volatiles to the experimental composition would affect the position of the liquidus and the crystallinity of the assemblage recovered from each run.

Major element oxide variability with MgO content

The Harker plots shown in Figure 4.5 show the variation of major element oxides with MgO content of the experimental glass. Decreasing MgO content is a proxy for crystallisation as the first phase to appear on the liquidus, olivine, decreases the MgO concentration of the remaining liquid (Langmuir et al., 1992). Data symbols in Figure 4.5 indicate which phases are present in each experimental charge and are therefore in equilibrium with the glass composition plotted. The use of Harker plots partly obscures compositional variability in the glass (and silicate phases in subsequent plots), however individual compositions have been plotted as smaller grey symbols to show the extent of compositional variability seen within the capsules. The aim of these figures is to show cooling and crystallisation relationships, average glass and mineral compositions are appropriate to show these trends.

Figure 4.5a shows the variation of SiO₂ with MgO in the glass. For the 0.001 kbar experiments SiO₂ content increases as crystallisation progresses, until clinopyroxene arrives on the liquidus and the SiO₂ content plateaus. The 5 kbar glasses are highly variable in SiO₂, with the most crystalline assemblages showing the lowest SiO₂ content glass, which is unexpected given that olivine and clinopyroxene crystallisation should have driven the SiO₂ content up. Experiments equilibrated at 10 kbar shown very little SiO₂ content variation and show large gaps in MgO content between clusters of experiments, making it more difficult to highlight trends between the oxides.

Both the 0.001 and 10 kbar glasses show an increase in FeO content as crystallisation progresses, particularly once plagioclase arrives on the liquidus (Figure 4.5b). The 5 kbar experiments again show a different trend, with much lower FeO contents that the other

two pressures. Figures 4.5c–d show clearly how Al_2O_3 contents and CaO/Al_2O_3 ratios are affected by different phases crystallising. Al_2O_3 slowly increases and CaO/Al_2O_3 stays level while olivine is the only phase on the liquidus at 0.001 kbar. When plagioclase starts to crystallise at MgO ~ 10 wt% the Al_2O_3 content of the glass rapidly decreases, increasing the CaO/Al_2O_3 ratio. Then at MgO ~ 9 wt% clinopyroxene also starts to crystallise, causing the Al_2O_3 content to stay constant and CaO/Al_2O_3 to start to decrease. CaO/Al_2O_3 decreases in the 10 kbar glasses because clinopyroxene crystallisation dominates at high pressures. This same relationship between phase crystallisation and changes in oxide concentration is not as pronounced for CaO (Figure 4.5e).

Glass Na₂O contents can be seen to increase slightly with crystallisation for 0.001 and 10 kbar experiments, as expected given the slightly incompatible nature of Na in silicate phases (apart from sodic plagioclase). However, the 5 kbar glasses show very high concentrations of Na₂O, over 5 wt% higher than the starting bulk composition, suggesting that these experiments have gained Na, as stated previously.

Major element oxide variability with temperature

The previous section presented the variation of major element oxides with MgO content of the experimental glass, but Figure 4.6 shows the variation of oxides with equilibrium temperature. Glass MgO content can be seen to decrease with temperature for all experimental pressures, indicating the validity of the assumption that MgO content is a good proxy for temperature and crystallisation (Figure 4.6a). MID2014-25, equilibrated at 10 kbar and 1310 °C, does not fit this trend as it plots at a MgO content more akin to 1280 °C experiments, which matches the textural observations made on this experiment in the previous section (Figure 4.3a & c). Figures 4.6c–f also highlight the anomalous nature of MID2014-25.

Figures 4.6a–b again highlight that the super-liquidus experiments plot at MgO and SiO₂ contents greater than that of the starting bulk composition (MID2014-21), despite also having not crystallised any phases. Figure 4.6c shows a clear decrease in CaO content as both clinopyroxene and plagioclase appear on the liquidus at 10 kbar, while Figure 4.6d again highlights the Fe-loss experienced by 5 kbar experiments. The variation of Al₂O₃ content and CaO/Al₂O₃ ratio as different phases crystallise is again shown in Figures 4.6e–f, and is as described in the previous section.

Bulk composition modification

The glass compositions of experiments equilibrated at 5 kbar show evolution trends that are unlike the trends shown by the other two pressures. This is shown most clearly by



Figure 4.5 Major element concentrations of the glass within each experimental charge. (a) SiO_2 , (b) FeO, (c) Al_2O_3 , (d) CaO/Al_2O_3 , (e) CaO, and (f) Na₂O. Individual analyses plotted in grey; average 0.001 kbar, purple; 5 kbar, red; and 10 kbar, blue. The bulk starting composition is a navy diamond and shapes indicate the phases present. Error bars show glass repeat analysis variability.



Figure 4.6 Major element concentrations of the glass within each experiment plotted against experimental equilibration temperature. (a) MgO, (b) SiO₂, (c) CaO, (d) FeO, (e) Al₂O₃, and (f) CaO/Al₂O₃. Individual analyses plotted in grey; average 0.001 kbar, purple; 5 kbar, red; and 10 kbar, blue. The bulk starting composition is a navy diamond and shapes indicate the phases present. Error bars show glass repeat analysis variability.

Na₂O, with the 5 kbar experiments having up to 6 wt% more Na₂O than the starting bulk composition and experimental charges from other pressures (Figure 4.5f). The 5 kbar bulk composition appears to have also been modified for FeO, as lower concentrations than expected are measured in the 5 kbar capsules (Figures 4.5b & 4.6d). CaO may also have been lost from these experiments, though the change is not as extreme as the Na-gain or Fe-loss (Figures 4.5e & 4.6c; quantified by mass balance with details below, Table 4.1). These modifications to the starting bulk composition during the course of the experimental procedure have had a knock on effect on the measured concentrations of the other major element oxides.

The Na-gain experienced by the 5 kbar experiments is likely due to the leaching of NaCl pressure medium through the experimental assemblage and into the capsule (Moore et al., 2008). Small crystals of NaCl were observed by SEM in some of these charges, as detailed in the previous section. The presence of such crystals is reflected in the glass chemistry, within which the Na would dissolve during the experiment. The extent of Na-gain is likely different for each 5 kbar experimental run, as there is no reason why the same amount of Na⁺ would diffuse into the graphite capsule each time. Therefore the mineral assemblage for each 5 kbar experiment is the result of a different bulk composition each time, and the evolution of the phase compositions with changing pressure-temperature conditions is also a result of compositional changes.

Fe-loss is generally associated with the dissolution of Fe with the walls of a metal capsule, however in the 5 kbar experimental procedure a metal capsule was not used (Moore et al., 2008; Jakobsson et al., 2014). It is possible that the iron had an affinity for the NaCl pressure medium, causing it to diffuse out of the starting bulk composition at the capsule centre. However, without analysis of the rest of the experimental assemblage, it is difficult to deduce where the iron has gone.

4.2.3 Mineral compositions

This section presents the compositions of the condensed phases assumed to be in equilibrium with the glasses presented in the previous section. Figure 4.7a shows the average forsterite content of olivine plotted against the magnesium number, $Mg\# = X_{Mg}/(X_{Mg} + X_{Fe^{2+}})$, of the host experimental liquid, where X_i is the fraction of cation *i* in the mineral of interest. Mg# has been calculated using $X_{Fe^{2+}}$ assuming a ferric to total iron ratio based on an oxygen fugacity of -1.5 $\Delta \log f O_2$ (FMQ) and the ferric to ferrous iron ratio parameterisation of Kress and Carmichael (1991). Again, the shapes of the data symbols indicate the phases present within the experimental charge.



Figure 4.7 (a) olivine forsterite content (X_{Fo}^{Ol}) against liquid Mg#, lines indicate $K_D = 0.26, 0.30, 0.34$ (red), and 0.38 (b) plagioclase anorthite content (X_{An}^{Pl}) against liquid Ca#, and (c) clinopyroxene Mg# against liquid Mg#, lines indicate $K_D = 0.20, 0.30$ (red), 0.40, and 0.50. Multiple analyses plotted in grey; 0.001 kbar, purple; 5 kbar, red; and 10 kbar, blue. Shapes indicate phases present.

Experiments equilibrated at 0.001 kbar have produced olivine compositions that decrease in forsterite content as crystallisation progresses, from $X_{Fo} = 0.87$ to 0.80 (Table 4.1). These compositions indicate a Mg-Fe partitioning coefficient between olivine-liquid ($K_D^{Ol-Liq}_{D Fe-Mg}$) of 0.32–0.34. This range of values matches those expected from previous observations of olivine-liquid equilibria, indicating that equilibrium between olivine and melt has been achieved within these experimental charges (Putirka, 2008b; Matzen et al., 2011). The 5 kbar olivine compositions lie along a $K_D^{Ol-Liq}_{D Fe-Mg} = 0.34$ line, suggesting equilibrium between olivine and liquid, however the olivine compositions are highly forsteritic; much higher than expected olivine compositions in equilibrium with basaltic liquid (Langmuir et al., 1992). These compositions can be explained by observing that the 5 kbar glasses have lost FeO (Figures 4.5b & 4.6d), therefore the glass Mg# is much higher, more akin to a primary mantle melt, hence the crystallisation of highly forsteritic olivines ($X_{Fo} > 0.92$). The 10 kbar olivines do not show constant partitioning coefficients, which along with olivine compositional variability within some of the low temperature charges, suggests that olivine and liquid have not attain equilibrium in all cases.

Figure 4.7b shows the average anorthite content of plagioclase plotted against the calcium number, $Ca\# = X_{Ca}/(X_{Ca} + X_{Na})$, of the host experimental liquid. The three experimental pressures plot in different regions of liquid Ca# space. 5 kbar experiments have low X_{An} (< 0.75) and low liquid Ca# (< 0.65), which is likely due to the Na-gain experienced by these capsules causing lower liquid Ca# and therefore more albitic plagioclase to crystallise. Experiments crystallised at 0.001 and 10 kbar plot at liquid Ca# > 0.75, lying on the same trend of decreasing anorthite content as liquid Ca# decreases. The difference between the two sets of plagioclase compositions could be a result of different equilibrium pressures affecting the partitioning of Na, Ca, Si, and Al between plagioclase and liquid. However, Na-loss in the 0.001 kbar experiments, which is thought to be low, could have increased the Ca content of the liquid and equilibrium plagioclase if significant enough.

Figure 4.7c shows that the high clinopyroxene compositional variability has resulted in average clinopyroxene compositions that do not conform to definite Fe-Mg partitioning trends between clinopyroxene and liquid ($K_{D \ Fe-Mg}^{Ol-Liq}$). The red line indicates a $K_{D \ Fe-Mg}^{Ol-Liq} = 0.30$, which is within the range of partitioning coefficients expected for clinopyroxene (Putirka, 2008b), however very few experiments follow this line. Instead there is a very general trend of clinopyroxene Mg# decreasing as liquid Mg# decreases, however the gradient for this relationship is different for each equilibration pressure. Identification of these trends is hindered by the small number of experiments saturated in clinopyroxene, however the spread of individual analysis compositions does indicate that there is a degree of disequilibrium within some of the charges.

4.2.4 Liquid-mineral modal proportions

Textural observations, glass chemistry and mineral compositions presented in previous sections all indicate that crystallinity within the experimental charges increases as equilibrium temperature decreases. The proportion of crystalline phases within each experiment can be estimated by mass balancing the major element oxides measured in the experimental phases with the known starting bulk composition (Table 3.1). For each major element, j, of known concentration, X, in each experimental phase, i, there exists a phase proportion, p_i , that satisfies the following:

$$\sum_{i} p_i X_j^i = X_j^{bulk},$$

such that:

$$\sum_{i} p_{i} = 1, \ p_{i} \ge 0, \ \text{and minimising} \ \sum_{j} \left(\sum_{i} (p_{i} X_{j}^{i}) - X_{j}^{bulk} \right)^{2}.$$

This method of least squares under equality and inequality constraints minimises the misfit between the bulk composition calculated from mineral proportions and the true bulk composition by solving simultaneous equations for each major element oxide. These equations were solved in R using the lsei function, which also provides a measure of fit quality, however a separate fit statistic has been calculated:

$$\chi^2 = \sum_j \frac{(X_j^{recalc} - X_j^{bulk})^2}{\sigma_j^2} \text{ where } X_j^{recalc} = \sum_i p_i X_j^i.$$

The χ^2 misfit statistic is the sum of the squared residuals for recalculated and true bulk composition divided by the analytical variance, σ_i^2 . This is converted to the reduced χ^2 statistic by dividing by the number of degrees of freedom, ν , which in this case is one fewer than the number of measured major element oxides. The smaller the reduced χ^2 statistic, the better the fit quality of calculated phase proportions.

The results of such calculations are shown in Table 4.1 and Figure 4.8, with bars centred on experimental equilibrium temperature. Average mineral and glass compositions have been used, as they are the most representative values for mineral chemistry given that some of the charges show considerable compositional variability. Experiments conducted at 0.001 kbar show a smooth increase in crystallinity as equilibrium temperatures are reduced, with the mineral proportions of all condensed phases increasing once they appear on the liquidus. These mineral proportion estimates match the textures observed from SEM imaging (Figure 4.1), with the lowest temperature experiment reaching a crystallinity of 51%, which is dominated by plagioclase and clinopyroxene crystals. Reduced χ^2 statistics for the 0.001 kbar calculations indicate a good quality of fit, apart from the three highest temperature experiments, which are super-liquidus showing only quenched glass.

The phase proportion plot for 5 kbar also shows an increase in crystallinity with decreasing temperature, however the reduced χ^2 statistic indicates a very poor fit for each of these calculations. Therefore, it is surprising to see that crystallinity increases in a similar way to the 0.001 kbar experiments, even though the recalculated bulk composition is quite different to that of the true bulk. Textural observations from these charges (Figure 4.2) confirm that the general trend calculated is true, however the absolute values of mineral proportions perhaps cannot be trusted.

Experiments conducted at 10 kbar have some anomalous results, highlighted by the lack of crystallinity increasing smoothly as equilibrium temperature decreases. The experiments



Figure 4.8 Calculated modal proportions of silicate phases and glass for experiment plotted against temperature. Separate plots for each experimental pressure. Glass, green; olivine, light grey; plagioclase, medium grey; and clinopyroxene, dark grey. Blue diamonds (and numbers) indicate the quality of fit between a phase proportion-estimated bulk and the known bulk composition; quantified by a reduced χ^2 misfit statistic.

equilibrated at 1310 °C (MID2014-25) and 1270 °C (MID2014-27) show higher extents of crystallinity than expected when compared to the other experiments. For MID2014-27 this is because the glass composition could not be measured by EPMA due to the small areas of glass in between condensed phases. Textural observations (Figure 4.3) suggest that MID2014-27 has a higher extent of crystallinity than the lowest temperature experiment, suggesting that it, along with MID2014-25, appears to have equilibrated at a lower temperature than its run conditions. Alternatively crystallinity could be increased by equilibrating at a higher pressure than 10 kbar, which would allow the assemblage to equilibrate further from the liquidus and produce higher crystallinity (Figure 4.4). These two experiments have been removed from the final plot to show the crystallinity as temperature decreases.

Deviations from starting bulk composition

Bulk compositions were recalculated using all major element oxides from all experiments to ascertain whether the starting bulk composition had been modified during the experimental procedure. Na-loss is a common problem in experiments conducted using a gas-mixing furnace as Na volatilises into the gas fluxing past the charge, therefore gas flux is kept to the minimum required to maintain oxygen fugacity (Yang et al., 1996). Recalculating bulk compositions provides a way to constrain the amount of Na-loss experienced by 0.001 kbar experiments.

Results of such calculations indicate that the three highest temperature 0.001 kbar experiments have lost ~ 0.5 wt% Na₂O during the 24 hour experimental run (Table 4.1). These experiments contain only glass and were the first gas-mixing furnace experiments conducted, therefore it is likely that Na-loss was caused by too high a gas flux from experimental procedure inexperience. Some minor Na-loss has been calculated for the rest of the 0.001 kbar experiments, but not as significant as those at super-liquidus temperatures.

Experiments conducted at 5 kbar produced the largest reduced χ^2 statistics, indicating large residuals between calculated and true bulk compositions. Bulk composition calculations indicate that these experiments have experienced Na-gain of 1.5–5.6 wt% Na₂O and Fe-loss of 0.4–4.7 wt% FeO (Table 4.1). The Na-gain is a result of NaCl pressure medium diffusion into the experimental charge centre, while the Fe-loss suggests Fe diffusion out of the capsule. 10 kbar experiments are unaffected by both Na-gain or Fe-loss.

4.2.5 Partitioning of major elements between minerals and melt

At conditions of equilibrium between a crystalline phase and its host liquid, the partitioning of major element cations between the mineral and liquid can be described by an equilibrium partition coefficient. The partition coefficients for olivine (Fe-Mg), plagioclase (Ca-Na), and clinopyroxene (Fe-Mg) are shown below (Putirka, 2008b):

$$K_{D Fe-Mg}^{Ol-Liq} = \frac{X_{Fe^{2+}}^{Ol} X_{Mg}^{Liq}}{X_{Mg}^{Ol} X_{Fe^{2+}}^{Liq}}, \quad K_{D Ca-Na}^{Pl-Liq} = \frac{X_{Ab}^{Pl} X_{Al}^{Liq} X_{Ca}^{Liq}}{X_{An}^{Pl} X_{Na}^{Liq} X_{Si}^{Liq}}, \quad K_{D Fe-Mg}^{Cpx-Liq} = \frac{X_{Fe^{2+}}^{Cpx} X_{Mg}^{Liq}}{X_{Mg}^{Cpx} X_{Fe^{2+}}^{Liq}}$$

Expected values of these partition coefficients have been estimated from experimental phases and natural mineral-liquid pairs. Comparison of partition coefficients calculated from Miðfell experimental phase compositions to expected values is one way of checking for equilibrium between mineral and host liquid.

Fe-Mg exchange between olivine-liquid

Partition coefficients for the exchange of Fe-Mg between olivine-liquid have been calculated for each experimental charge using average glass compositions paired with (i) individual olivine analyses, and (ii) average olivine compositions. The ferric iron content of the glass has been calculated using the $X_{Fe^{3+}}/X_{Fe^{2+}}$ ratio parameterisation of Kress and Carmichael (1991), which is a function of temperature, pressure, oxygen fugacity and glass composition. Oxygen fugacity was kept at -1.5 $\Delta \log fO_2$ (FMQ) during the 0.001 kbar experiments by a mixture of CO and CO₂ gases. For 5 and 10 kbar experiments, the oxygen fugacity was buffered by the graphite capsule, again at around -1.5 $\Delta \log fO_2$ (FMQ). Calculated ferric iron values for each average glass composition are shown in Table 4.1.

If all the iron within the experimental charge is assumed to be Fe²⁺, then the 0.001 kbar olivine-liquid pairs record partition coefficients of $K_D = 0.30 \pm 0.01$ (Figure 4.9a). This matches expected values of K_D based on previous studies (Roedder and Weiblen, 1970; Ford et al., 1983; Bédard, 2005; Putirka, 2008b). However, if glass $X_{Fe^{2+}}$ contents are used that assume ferric iron is present, then $K_D = 0.33 \pm 0.01$, which is much closer to the partition coefficient calculated from 0.001 kbar experiments on synthetic Hawaiian picrite (Matzen et al., 2011). K_D slightly increases as temperature decreases for the 0.001 kbar experiments. Very similar values are calculated for the 5 kbar experiments, despite the Na-gain and Feloss affecting these experiments, which would have disrupted equilibrium between phases (Figure 4.9b). Only one of the 10 kbar experiments plots near expected equilibrium values for K_D . This observation could be the result of non-equilibrium glass composition being used in the calculations, as these experiments exhibit the most layering of phases with the charge, suggesting potential fractionation of crystalline phases from the liquid. MID2014-25 (1310 °C) continues to look out of place with respect to the other 10 kbar experiments, plotting at low K_D .

Ca-Na exchange between plagioclase-liquid

The partitioning coefficient for Ca-Na exchange between plagioclase-liquid is thought to be $K_D = 0.27 \pm 0.18$, however this value is very variable and dependent on a number of factors (Putirka, 2008b; Namur et al., 2012). Individual analyses of 0.001 kbar plagioclase paired with average glass compositions produce a wide range of K_D values, however the average plagioclase compositions show a trend of decreasing K_D with increasing temperature (Figure 4.9e), placing values within the expected equilibrium range. There are too few 5 and 10 kbar experiments saturated with plagioclase to give any meaningful insight into



Figure 4.9 Liquid-mineral major element partition equilibrium constants. (a) and (b) $K_D^{Ol-Liq}_{D Fe-Mg}$ against temperature, and (c) against total alkalis. (d) $K_D^{Cpx-Liq}_{Pe-Mg}$ against temperature. (e) and (f) $K_D^{Pl-Liq}_{D Ca-Na}$ against temperature. Partition coefficients calculated using $X_{Fe^{3+}}/X_{Fe^{2+}}$ from Kress and Carmichael (1991) for individual mineral analyses, light grey circles; average 0.001 kbar, purple; 5 kbar, red; and 10 kbar, blue. Partition coefficients assuming all Fe²⁺ plotted as large dark grey symbols.

the correlation of partitioning behaviour and experimental pressure, however the 10 kbar experiments do plot at the highest K_D values (Figure 4.9f).

Fe-Mg exchange between clinopyroxene-liquid

All the experimental charges saturated in clinopyroxene have at least one clinopyroxeneliquid pair that calculates a partitioning coefficient within the expected range of $K_D = 0.24$ – 0.31 (Putirka, 2008b; Mollo et al., 2013). However, the high variability of clinopyroxene compositions within the low temperature experiments, results in the average clinopyroxeneliquid pairs producing very different K_D values (Figure 4.9d). The high K_D values estimated by some of the clinopyroxene-liquid pairs from the low temperature experiments suggest that there is disequilibrium between clinopyroxene and liquid in these charges, with the extent of disequilibrium increasing as temperature decreases.

4.2.6 Clinopyroxene compositional variability

Previous sections have highlighted that individual clinopyroxene analyses show significant compositional variability within a single experimental charge; this is recorded in their Mg# and calculated K_D values. Table 4.1 shows the standard deviation of clinopyroxene Mg# measured in each charge, which indicates that Mg# variability increases as equilibrium temperature decreases. Clinopyroxene-liquid barometry depends on the equilibrium condition between clinopyroxene grains and host liquid, therefore to successfully test this barometer with the experiments presented here requires the identification of equilibrium clinopyroxene compositions.

Cation fractions

Clinopyroxene cation fractions have been calculated using individual major element oxide analyses and the method of Putirka (2008b), which is on a six-oxygen basis to give the general clinopyroxene formula, $A_2B_2O_6$, with four cations.

Figure 4.10 shows the cation fractions for all individual clinopyroxene analyses measured across all experimental pressures. The 0.001 kbar experiments show the highest compositional variability, with some analyses having up to 0.3 cations of Al on the tetrahedral site (Figure 4.10c). The three pressure suites show parallel negative correlations between X_{Mg} and (a) X_{Al} (Total Al), (b) X_{Na} , (c) $X_{Al(IV)}$, and to a lesser extent (d) X_{Fe} . There is a positive correlation between X_{Mg} and X_{Ca} . The 10 kbar experiments plot at slightly higher X_{Al} cation fractions than the other two pressures, indicating the incorporation of more Al



Figure 4.10 Individual clinopyroxene compositions from each experiment as cation fractions; (a) X_{Al} , (b) X_{Na} , (c) $X_{Al(IV)}$, (d) X_{Fe} , and (e) X_{Ca} all plotted against X_{Mg} . Coloured by equilibrium pressure; 0.001 kbar, purple; 5 kbar, red; 10 kbar, blue. Shapes indicate phases present.

into the clinopyroxene structure at equivalent X_{Mg} fractions. The correlations between cation fractions suggest that Na-Al and Mg-Al are coupled together.

Clinopyroxene compositional variability is clearly dependent on equilibration temperature, as shown in Figure 4.11. At lower temperatures for each pressure suite, the variability in calculated cation fraction increases. In Figure 4.11a the 0.001 kbar experiments form a wedge shape defined by a horizontal line at $X_{Mg} = 0.9$ and a sloped line joining the lowest X_{Mg} analyses. A similar wedge can be drawn in Figure 4.11c for X_{Ca} . Figures 4.11b & d–e show a slightly different relationship, with a horizontal line between the lowest cation fraction measurements, and a sloped line between the highest. At high temperatures, crystallinity is low within the charges and diffusion can occur relatively quickly, therefore mineral compositions are expected to be in equilibrium with the host liquid. At lower temperatures, crystallinity increases and diffusion rate decreases, making it more likely for disequilibrium compositions are likely defined by high X_{Mg} and X_{Ca} , with cation fractions decreasing as disequilibrium increases. In Figures 4.11b & d–e the reverse is true, equilibrium is defined by low X_{AI} and X_{Na} , while disequilibrium gives high cation fractions.

End-member fractions

Clinopyroxene cation fractions have been converted into end-member fractions using the equations detailed by Putirka (2008b) to investigate clinopyroxene variability further. The two experiments showing the most clinopyroxene variability, MID2014-12 (1170 °C, 0.001 kbar) and MID2014-23 (1240 °C, 10 kbar), are shown in Figure 4.12 with compositions recalculated as fractions of jadeite (Jd; NaAlSi₂O₆), Ca-Tschermak clinopyroxene (CaTs; CaAl^{VI}Al^{IV}SiO₆), enstatite-ferrosilite (EnFs; [Mg,Fe]₂Si₂O₆), and diopside-hedenbergite (DiHd; Ca[Mg,Fe]Si₂O₆). Fractions of the Ti-rich (CaTi; CaTiAl₂O₆) and Cr-rich (CrCaTs; CaCr₂SiO₆) clinopyroxenes were also calculated, but have not been plotted due to small end-member fractions.

The large range in DiHd and CaTs fractions measured suggests that the majority of the clinopyroxene variability can be described in terms of X_{DiHd} and X_{CaTs} , implying that the dominant exchange reaction between clinopyroxene compositions is:

$$(Mg,Fe)Si = Al^{IV}Al^{VI}$$
.

DiHd-rich compositions are thought to be equilibrium compositions, based on the observations of cation fractions in the previous section. High Mg and Ca cation fractions are



Figure 4.11 Clinopyroxene cation fractions plotted against temperature; (a) X_{Mg} , (b) X_{Al} , (c) X_{Ca} , (d) $X_{Al(IV)}$, and (e) X_{Na} . Coloured by equilibrium pressure; 0.001 kbar, purple; 5 kbar, red; 10 kbar, blue. Shapes indicate phases present.



Figure 4.12 end-member fractions of clinopyroxene from the two experiments showing the most compositional variability; mid12 (1170 °C, 0.001 kbar) in purple and mid23 (1240 °C, 10 kbar) in blue. Fractions of (a) jadeite (X_{Jd}), (b) Ca–Tschermak (X_{CaTs}), and (c) enstatite-ferrosilite (X_{EnFs}) plotted against fraction of diopside-hedenbergite (X_{DiHd}).

thought to be equilibrium values, hence high DiHd compositions are nearest equilibrium (Figure 4.12).

The highest DiHd clinopyroxenes in Figure 4.12a suggest that high pressure clinopyroxenes are more Jd-rich than lower pressure clinopyroxene, although there is a temperature difference between these two experiments. This observation tentatively supports observations of the pressure-sensitive incorporation of jadeite into clinopyroxene, as exploited by clinopyroxene-liquid barometers (Blundy et al., 1995; Putirka et al., 1996, 2003; Putirka, 2008b; Neave and Putirka, 2017). However, the full range of Jd fractions seen within the 0.001 kbar experiments is greater, and goes to higher contents, than the 10 kbar experiments. If clinopyroxene-liquid barometry was used on these Jd-rich compositions, then a large pressure overestimate would be calculated (> 20 kbar). Figure 4.12c shows a boomerang-shaped relationship between X_{EnFs} and X_{DiHd} , suggesting that the EnFs fraction initially increases as DiHd decreases, but then at $X_{DiHd} < 0.22$ it decreases. This is likely due to a different clinopyroxene end-member becoming more dominant at low DiHd (disequilibrium) conditions, potentially CaTs as Figure 4.12b shows a change in gradient at $X_{DiHd} = 0.22$, which increases the CaTs fraction.

Equilibrium compositions

Observations in the previous sections suggest that clinopyroxene shows a spectrum of compositions, from those that are in equilibrium with coexisting liquid to disequilibrium compositions. Cation fractions in Figure 4.11 suggest that equilibrium clinopyroxene compositions are characterised by high Mg and Ca, but low Na and Al. These cation fractions suggest that equilibrium compositions in terms of end-members are high in DiHd, and low in Jd and CaTs (Figure 4.12).

These observations can be linked to the clinopyroxene Mg# shown in Figure 4.7c and K_D values calculated for Figure 4.9d. Equilibrium compositions are high in X_{Mg} , therefore equilibrium Mg# is expected to also be high, as X_{Mg} varies more than X_{Fe} so primarily controls Mg# variability. Equilibrium partition coefficient values are expected to be low, K_D = 0.24–0.31, with disequilibrium compositions low in X_{Mg} giving high K_D values (Putirka, 2008b).

The variability of clinopyroxene-liquid K_D highlights a method for identifying equilibrium clinopyroxene values. Clinopyroxene-liquid pairs can be used to calculate K_D , but also temperature and pressure estimates, for comparison with expected values (Putirka, 2008b; Neave and Putirka, 2017). For natural samples, where equilibrium pressure and temperature are unknown, then comparison values could come from a different set of thermobarometers, with clinopyroxene-free thermometers more common than the equivalent barometer (Sugawara, 2000; Shejwalkar and Coogan, 2013). Another way of testing for the equilibrium compositions of clinopyroxene is to compare to expected end-member proportions, calculated using the equations of Putirka (1999).

Figures 4.13a–d and 4.14a–d compare observed clinopyroxene end-member proportions to those calculated for all experiments saturated in clinopyroxene. The coloured symbols have used known experimental equilibration conditions for end-member calculation, grey symbols have been calculated using an iterated pressure-temperature estimate using the Neave and Putirka (2017) barometer and Putirka (2008b) thermometer.

All four 0.001 kbar plots show that the high temperature experiment has produced clinopyroxene end-member compositions that plot close to a 1:1 line with expected values, indicating equilibrium between the crystalline phase and host liquid. The low temperature



Figure 4.13 Comparison of calculated and measured clinopyroxene end-member compositions crystallised at 0.001 kbar; (a) diopside-hedenbergite (X_{DiHd}), (b) jadeite (X_{Jd}), (c) Ca-Tschermak (X_{CaTs}), and (d) enstatite-ferrosilite (X_{EnFs}); estimates of (e) equilibrium temperature using Putirka (2008b) thermometer, and (f) Fe-Mg clinopyroxene-liquid partitioning equilibrium constant for each clinopyroxene and liquid composition. 1170 °C in white, 1200 °C in purple.

experiment produces observed compositions that start on the 1:1 line, but move away from it. Observed DiHd compositions extend to lower proportions than expected (Figure 4.13a), observed Jd and CaTs move to higher proportions (Figure 4.13b–c). Observed EnFs is fairly constant, however the expected values show more variability. This is due to calculated X_{EnFs} being dependent on observed X_{DiHd} , which shows high variability away from equilibrium compositions. Therefore, all the observed X_{EnFs} are likely close to true expected equilibrium values if DiHd variability is ignored (Figure 4.13d). Calculated CaTs is also sensitive to observed DiHd, but not to the same extent as calculated EnFs.

Grey points in Figures 4.13a–d show that when estimated pressure and temperature are used, more of the clinopyroxene compositions plot on the 1:1 expected-observed composition line, particularly for DiHd. This observation suggests the importance of having an independent temperature or pressure estimate to prevent the misidentification of equilibrium clinopyroxene within natural samples. If $X_{DiHd} = 0.4$ was assumed to be the equilibrium composition from the grey data points and then used for thermobarometry, temperature would be overestimated by ~ 200 °C (Figure 4.13e) and pressure would be overestimated by ~ 20 kbar overestimate (Figure 4.18a).

The 10 kbar experiments show a similar relationship between expected and observed endmember proportions, however for DiHd proportions the 1:1 composition is not the highest DiHd proportion observed, but rather a middle value equal to the composition measured in the highest temperature experiment (Figure 4.14a). Again the grey points plot closer to the 1:1 line, suggesting a range of expected compositions, rather than just one.

Clinopyroxene-liquid thermometry produces the most accurate temperature estimates for DiHd compositions that lie on, or close to, the 1:1 expected-observed line (Figures 4.13e and 4.14e). Expected clinopyroxene-liquid partition coefficients also best match the values calculated from these compositions, further supporting the use of expected end-member compositions as a test for clinopyroxene-liquid equilibria (Figures 4.13f and 4.14f).

Clinopyroxene disequilibrium

Previous studies have observed disequilibrium compositions of clinopyroxene, attributing their formation to rapid cooling rates and sector zoning within clinopyroxene grains (Vernon, 2004; Mollo et al., 2010, 2013; Welsch et al., 2016; Neave and Putirka, 2017).

Mollo et al. (2010, 2013) conducted experiments on an alkali basalt composition from Mt Etna volcano, Italy to quantify the effect of cooling rate on clinopyroxene-liquid equilibrium. Their experiments have shown that as cooling rate is increased, clinopyroxene compositions become more Al- and Na-rich, but lower in Mg and Ca, which equates to higher CaTs and lower DiHd end-member compositions. These observations match those made on the



Figure 4.14 Comparison of calculated and measured clinopyroxene end-member compositions crystallised at 10 kbar; (a) diopside-hedenbergite (X_{DiHd}), (b) jadeite (X_{Jd}), (c) Ca-Tschermak (X_{CaTs}), and (d) enstatite-ferrosilite (X_{EnFs}); estimates of (e) equilibrium temperature using the Putirka (2008b) thermometer, and (f) Fe-Mg clinopyroxene-liquid partitioning equilibrium constant for each clinopyroxene and liquid composition. 1240 °C in white, 1280 °C in light blue, 1290 °C in navy.

Miðfell composition experiments (Figures 4.11, 4.13, and 4.14). Al-rich clinopyroxenes are crystallised as a result of rapid undercooling, which causes high nucleation rates and rapid growth out of equilibrium. It has been observed that Al-rich clinopyroxenes can take a long time to be re-equilibrated, hence their compositions can be preserved (Mollo et al., 2013).

The experiments presented in this chapter have a different experimental procedure than Mollo et al. (2013). Rather than each experiment heating up to above the liquidus and cooling, Miðfell experiments were heated up to the equilibrium temperature of interest and allowed to crystallise. Therefore, the lowest temperature experiments stayed tens of degrees below the bulk liquidus.

The starting assemblage is powdered glass, which is metastable, and will melt or crystallise silicate phases as soon as temperature and diffusion will allow. Due to the thermal profile of the experimental capsule, with the hotspot near the top, it is conceivable that during heating the glass undergoes several stages of partial melting. Melt will segregate from the residue and migrate to the top of the capsule, where it could start crystallising silicate phases. If heating is slow enough then this process of partial melting, migration, and crystallisation could produce chemical layering within the capsule. Experimental assemblages were heated to temperature over 10–15 minutes to reduce the potential of this process causing chemical segregation within the capsule. Also the thermal gradient across the 10 kbar experiments is sufficiently small to reduce this process.

Rapid crystallisation of clinopyroxene from the partially melted glass-residue mixture could have created disequilibrium clinopyroxene compositions (Mollo et al., 2013). As crystallinity of the assemblage increases and equilibrium is approached, mineral growth will slow to allow equilibrium clinopyroxene compositions to crystallise (Vernon, 2004).

Figures 4.1–4.3 show the textures of all experimental assemblages. The most crystalline experiments display textures that are all quite different, however each charge displays considerable variability in clinopyroxene compositions, highlighted by Mg#. MID2014-12 (Figure 4.1j) shows a fairly homogeneous texture throughout the charge, but clinopyroxene compositions show variability. MID2013-23 (Figure 4.3f) has compositional analyses taken from only one layer of the capsule, but again it shows clinopyroxene variability. MID2014-14 (Figure 4.3d) has distinct layering of different assemblages present going down through the capsule. Clinopyroxenes vary within each distinct layer, however the full composition range can only been seen when considering the whole capsule. MID2014-34 (Figure 4.2d) shows a crystalline interior of olivine, plagioclase, and clinopyroxene, with a glassy rim. Clinopyroxene compositions are seen to vary in an unsystematic way through the crystalline interior.

Observations of other mineral compositions suggest that fractionation within the layered experiments is responsible for some of the compositional variability, with mineral phases becoming separated from the liquid and no longer being able to equilibrate. This variation could be the result of the melt fractionation process described above. However, in MID2014-34 there is no systematic change in mineral composition going from the capsule side towards the centre, which suggests that fractionation disequilibrium is not clear cut in this capsule. Fractionation is expected cause the minerals most isolated from the melt to be furthest from equilibrium with the liquid composition. Therefore, it is likely that the extreme clinopyroxene compositions and disequilibrium are in part the result of rapid growth as suggested by Mollo et al. (2013).

It is more difficult to ascertain whether this disequilibrium growth has manifested as sector-zoned clinopyroxenes with Al-rich and Mg-rich regions, or whether the phases are homogeneous in composition with a population of crystal compositions present. This distinction cannot easily be made from the images in Figures 4.2 & 4.3 as the clinopyroxene tend to be very small. The lack of obvious zones in larger grains could either be due to the orientation of the crystal, or due to the lack of zoning. Disequilibrium clinopyroxenes are apparent from compositional analyses, however it remains unclear as to whether these crystals are zoned or not. Given the rapid growth required to produce Al-rich clinopyroxene compositions it seems more likely that the whole crystal would be out of equilibrium, i.e. homogeneous, rather than a just a particular zone of the crystal, i.e. zoned.

4.3 Crystallisation models

Petrolog and MELTS are programs that combine multiple melt-mineral equilibrium models to simulate the processes of silicate melting and crystallisation (Danyushevsky and Plechov, 2011; Asimow and Ghiorso, 1998). They can be used to model mantle melting and the composition of primary mantle melts to predict the crystallisation path of a melt and the compositions of equilibrium silicate phases. They can also be used to reverse the processes of crystallisation on evolved melts, such as melt inclusions that have undergone post-entrapment crystallisation. Given the usefulness of these two programs in providing simple modelling checks to the nuanced history of igneous rocks, it is appropriate to test their ability to reproduce the experimental assemblages as detailed in previous sections. The initial Miðfell bulk starting composition has been run through Petrolog and MELTS for the pressures at which the experiments were conducted to test the reliability of these programs.

4.3.1 Petrolog

Petrolog offers an algorithm for combining several mineral-melt equilibrium models for major and trace elements to describe crystallisation and reverse crystallisation processes (Danyushevsky and Plechov, 2011). The mineral-melt equilibrium models used to test the reliability of Petrolog in this section are those of Danyushevsky (2001), though it must be noted that Petrolog offers several parameterisations for each silicate phase.

Figure 4.15a compares the observed chemistry of experimental olivine to that predicted by Petrolog. It shows that there is a reasonably good match between observed compositions and Petrolog predictions, with both datasets forming approximately linear arrays in X^{Ol}_{Fo}-Mg# ^{Liq} space. This observation suggests that the partition coefficient, $K_{D Fe-Mg}^{Ol-Liq}$, used in the Petrolog parameterisation describes the relationship between olivine forsterite and liquid composition well. For experiments conducted at 0.001 kbar, olivine forsterite compositions have been matched very well by Petrolog, however the 5 kbar experiments bear no resemblance to predicted compositions. This difference is due to the Fe-loss experienced by the 5 kbar runs, resulting in higher Mg# ^{Liq} and therefore more forsteritic olivines as a result. The 5 kbar olivines lie on the extrapolated Petrolog line, suggesting that the equilibrium relationship, as seen at other pressures, holds true. The 10 kbar experiments have crystallised olivine at much higher Mg#^{Liq} compositions than predicted by Petrolog. When compared to clinopyroxene in Figure 4.15e it can be seen that at 10 kbar clinopyroxene is predicted to arrive on the liquidus before olivine, i.e. at higher Mg# ^{Liq}, hence the discrepancy between 10 kbar observed and predicted olivine. However, as discussed previously, the experiments suggest that olivine crystallised before clinopyroxene, which is shown by: (i) textures of olivine surrounded by clinopyroxene, and (ii) the lowest crystallinity assemblage being olivine only. The two phases are both close to the liquidus at the highest 10 kbar experimental pressures, hence the slight ambiguity in crystallisation order.

Predicted plagioclase compositions do not match observed compositions as successfully as olivine (Figure 4.15c). Again there is reasonable agreement for 0.001 kbar experiments between observed and predicted compositions, with the gradient between X_{An}^{Pl} and Ca^{# Liq} matching well. Unlike olivine K_D , the relationship between X_{An}^{Pl} and Ca^{# Liq} is dependent on $K_{D Ca-Na}^{Pl-Liq}$, X_{Si}^{Liq} , and X_{Al}^{Liq} due to the coupled exchange reaction between anorthite-albite plagioclase end-members, making it harder to predict. H₂O also has an effect on this relationship, though it is expected that the H₂O content of these experiments is low. The difference for 5 kbar experiments can be explained by Na-gain, as described in previous sections. Increased Na in the liquid has reduced the Ca^{# Liq}, shifting all 5 kbar plagioclase compositions to the left on Figure 4.15c. Anorthite content for this experiment is at the lower end of compositions predicted by Petrolog. Predicted 10 kbar plagioclase compositions



Figure 4.15 Comparison of observed mineralogy from experiments and predicted mineralogy from Petrolog. (a) X_{Fo} against liquid Mg#, (b) liquid Mg# against temperature, (c) X_{An} against liquid Ca#, (d) modal proportion of liquid against temperature, and (e) clinopyroxene Mg# against liquid Mg#. Coloured by equilibrium pressure; 0.001 kbar, purple; 5 kbar, red; 10 kbar, blue. Petrolog predictions as lines. Phases present in experimental charges represented by different shapes.
also do not match those observed in experiments, with predictions giving lower anorthite compositions than measured for the same Ca# ^{Liq} compositions. This observation suggests that the Danyushevsky (2001) plagioclase model might not be appropriate for high pressures.

The clinopyroxene compositional variability makes it difficult to judge the fit quality fit between observed and predicted compositions in Figure 4.15e. However, it can be seen that Petrolog can predict the Mg# ^{Liq} at which clinopyroxene appears on the liquidus for 0.001 and 10 kbar. Fe-loss for the 5 kbar experiments means that observed clinopyroxene compositions plot at higher Mg# ^{Liq} than expected, however the clinopyroxene compositions are not high in Mg as seen for 5 kbar olivines. This observation suggests that the $K_D^{Cpx-Liq}$ relationship for clinopyroxene is more complicated than that of olivine, and has potentially been affected by Na-gain for these experiments. If the Petrolog relationship is taken to represent equilibrium conditions, then it suggests that the highest Mg# ^{Cpx} compositions are in equilibrium at 0.001 kbar, as concluded in previous sections. However, for 10 kbar this equilibrium composition is within the range of Mg# ^{Cpx} measured, suggesting that the highest Mg# ^{Cpx} clinopyroxenes are not those in equilibrium with the liquid, again as concluded previously (Figure 4.15e).

Figure 4.15b shows that the predicted Mg# ^{Liq} matches observed compositions reasonably well for 0.001 kbar experiments, however it evolves too steeply at lower temperatures, achieving lower Mg# ^{Liq} than observed. This discrepancy is likely due to predicted crystallinity being greater than observed, therefore the liquid has been modelled to have evolved more than observed (Figure 4.15d). Looking at the curves for higher pressures suggests that liquids of the same composition are predicted at higher temperatures, reflecting the fact that the liquidus temperature increases with pressure. The 5 kbar experiments again do not match expected compositions due to Fe-loss and the 10 kbar experiments plot in the same region as predicted values, but produce some anomalous points due to potentially incorrect equilibration temperatures.

The variation of crystallinity with equilibration temperature agrees with higher liquidus temperatures at higher pressures (Figure 4.15d), though predicted crystallinity curves do not match observations. This could be due to differences in observed and expected mineral liquidus temperatures, or perhaps it is due to incorrect calculation of mineral proportions from phase compositions. It is clear though that Petrolog is overestimating the liquidus temperature at all pressures used in the calculations.

Petrolog can predict mineral compositions reasonably well for the 0.001 kbar experiments, as well as coming close to matching liquid compositions and crystallinity fractions. Modelling further highlights the effect Na-gain and Fe-loss has had on the 5 kbar experiments, causing observed compositions to differ greatly from predictions. Experiments conducted at 10 kbar



Figure 4.16 Expected plagioclase anorthite content based on Namur et al. (2012) model. (a) Comparison of temperature dependent and temperature independent parametrisations of plagioclase anorthite content. Plagioclase anorthite content (X_{An}) plotted against (b) liquid Ca# (c) equilibration temperature. Individual plagioclase analyses in grey. Average plagioclase compositions plotted for: 0.001 kbar, purple; 5 kbar, red; and 10 kbar, blue; shapes indicate phases present within experimental charge. Predicted plagioclase compositions in blue (T independent) and green (T dependent) diamonds.

match Petrolog predictions to some extent, however clinopyroxene is predicted on the liquidus first, which has affected the expected compositions of the other minerals. The plagioclase model was particularly unsuccessful at recovering 10 kbar plagioclase compositions.

4.3.2 Plagioclase-melt equilibrium models

The previous section highlighted the failings of the Danyushevsky (2001) plagioclase model at high pressures, here an alternative plagioclase-melt equilibrium model is discussed for comparison. The Namur et al. (2012) plagioclase model consists of two parameterisations, one of which is temperature dependent, with both calculating the anorthite content of plagioclase in equilibrium with silicate melt. The liquid composition is converted into cation

proportions based on eight oxygens, and then X_{Si} , X_{Al} , X_{Fe} , X_{Mg} , X_{Na} , and X_{Ca} are used to calculate equilibrium plagioclase.

Figure 4.16 shows that the temperature dependent and temperature independent parameterisations produce very similar estimates of anorthite content, but the T-independent estimate is slightly higher than the T-dependent value. Experimental liquid compositions have been used to estimate the equilibrium plagioclase composition based on these two plagioclase models. The T-independent model estimates the 0.001 kbar plagioclase compositions relatively well, however the 10 kbar experiment compositions are overestimated by the plagioclase models. This discrepancy is perhaps unsurprising as the Namur et al. (2012) model was calibrated using a suite of 0.001 kbar experiments, so the parameterisation may not extend well to higher pressures.

A comparison of the Namur et al. (2012) plagioclase model with that of Danyushevsky (2001), suggests that the former better describes equilibrium plagioclase compositions calculated from silicate liquids. However, this might not be a fair comparison as the Namur et al. (2012) model was directly given equilibrium glass compositions, while Petrolog formulated glass compositions based on the crystallisation evolution of the starting bulk composition used in the experiments. Therefore the Namur et al. (2012) model started closer to experimental equilibrium compositions than Petrolog, giving the Petrolog algorithm more room for deviation from observed experimental compositions.

4.3.3 MELTS

The MELTS program combines the mineral-melt models of numerous igneous minerals, making it capable of modelling a vast number of igneous assemblages, from mantle melting of ultramafics at high pressure to acidic melt crystallisation within the crust. The high degree of variance available within MELTS requires careful manipulation to get the program to behave in the way intended by the user. For example, in modelling crystallisation of the Miðfell basaltic glass starting bulk composition at 10 kbar, the crystallisation of garnet had to be suppressed to prevent crystallisation of a 'mantle-like' assemblage. In the same way that Petrolog was tested, the following section compares mineral and melt compositions predicted by MELTS equilibrium crystallisation models to those observed in experimental charges described in previous sections.

Figure 4.17a shows that olivine compositions are poorly predicted by MELTS, particularly at high pressures. 0.001 kbar olivines are close to being recreated by MELTS, with olivine crystallisation starting at the correct Mg#^{Liq} composition, however predicted olivine forsterite concentrations are too low. At 5 and 10 kbar, MELTS predicts that olivine only reaches the liquidus once the melt has evolved to low Mg#^{Liq} compositions, which is not what is

observed within the experimental charges. The reason for this is because MELTS prioritises the crystallisation of both clino- and orthopyroxene at high pressure.

Plagioclase compositions are also poorly predicted by MELTS (Figure 4.17c). At 0.001 kbar there is a reasonable match between predicted and observed plagioclase compositions, however the highest anorthite plagioclase compositions are not calculated by MELTS. High pressure compositions produce the right gradient between X_{An}^{Pl} and Ca# ^{Liq}, but again not for the same compositions as those observed. Again, 5 kbar compositions are an unreliable comparator due to Na-gain and Fe-loss.

MELTS predicts that at 10 kbar multiple clinopyroxene compositions are crystallised (augite and pigeonite) as shown in Figure 4.17e. These compositions match up reasonably well with those measured from 10 kbar experiments, though it was not obvious that two clinopyroxene populations had crystallised; just one with disequilibrium compositional variability. These multiple clinopyroxene compositions are likely replacing the crystallisation of olivine within the MELTS parameterisation at 10 kbar. At lower pressures Mg# ^{Cpx} and Mg# ^{Liq} predicted trends parallel each other, but again do not match observed compositions, with 0.001 kbar clinopyroxene predicted on the liquidus at higher Mg# ^{Liq} than observed. Predictions suggest that crystallisation is dominated by clinopyroxene at high pressures, which causes a mismatch between the predicted compositions of other mineral phases and the observations.

Glass composition and crystallinity curves resemble those predicted by Petrolog, with 0.001 kbar predictions being close to observed values (Figure 4.17b and d). However, again the glass evolution with decreasing temperature is not correct as Mg# ^{Liq} and crystallinity increase too rapidly. The liquidus is predicted to increase in temperature with increasing pressure, as seen by experiments, however MELTS shows larger increases in liquidus temperature as pressure increases.

Overall, given the myriad ways of varying MELTS input, it is probably possible to reproduce the observed compositions to a better extent than displayed here, however sensible parameters were chosen to mimic the choices made when modelling a composition of unknown equilibrium conditions. H₂O content was assumed to be low, but concentrations as low as 0.5 wt% could still decrease the liquidus by $\sim 30^{\circ}$ C. This temperature decrease could be the reason why MELTS predicts a higher solidus temperature than observed for the 10 kbar experiments. When modelling a bulk composition or a certain target assemblage, it becomes more difficult to manipulate MELTS to give a solution that is both feasible and correct.



Figure 4.17 Comparison of observed mineralogy from experiments and predicted mineralogy from MELTS. (a) X_{Fo} against liquid Mg#, (b) liquid Mg# against temperature, (c) X_{An} against liquid Ca#, (d) modal proportion of liquid against temperature, and (e) clinopyroxene Mg# against liquid Mg#. Coloured by equilibrium pressure; 0.001 kbar, purple; 5 kbar, red; 10 kbar, blue. MELTS predictions as lines. Phases present in experimental charges represented by different shapes.

4.4 Igneous barometry

Numerous barometers can be used to calculate equilibrium pressures from igneous mineral and glass assemblages. These pressures can often be inferred to represent crystallisation pressures and therefore the depths at which magma chambers are present within the volcanic system of interest. The majority of igneous barometers utilise the pressure sensitivity of igneous mineral solid solutions and phase equilibria, with the stability of Na-pyroxene end-member jadeite (Jd; NaAlSi₂O₆) in clinopyroxene being the focus of several barometer calibrations (Putirka et al., 1996; Putirka, 1999; Putirka et al., 2003; Putirka, 2008b). Recent studies have further refined the Jd-in-Cpx barometer through recalibration with experimental data, however interest has also extended to barometers involving equilibria between multiple igneous phases and liquid (Yang et al., 1996; Kelley and Barton, 2008; Neave and Putirka, 2017; Ziberna et al., 2017). Multi-reaction thermobarometry has been a common aspect of metamorphic petrology, using the pressure and temperature sensitivity of several phase equilibria in conjunction to provide an average pressure-temperature estimate (Powell and Holland, 2008). It is only recently with the development of activity-composition models for mafic assemblages that multi-reaction barometry can be applied to igneous systems (Holland and Powell, 1998; Jennings and Holland, 2015; Green et al., 2016; Ziberna et al., 2017).

The experiments detailed in previous sections provide a method of testing the accuracy of these igneous barometers as the equilibration conditions of each experimental charge are known. However, it has been noted that some of the experimental charges contain minerals that appear to show disequilibrium compositions or mineral proportions that are out of line with other equilibration conditions. This section will mainly focus on clinopyroxene-liquid barometry, olivine-plagioclase-augite-melt (OPAM) barometry, and multi-reaction barometry, though it will also consider the thermometers used in conjunction with these barometers.

4.4.1 Clinopyroxene-liquid barometry

The pressure-sensitive incorporation of jadeite into clinopyroxene has been the basis for many igneous barometers over the past twenty years (Putirka et al., 1996, 2003; Putirka, 2008b). Tests on these barometer calibrations have shown that they tend to overestimate pressures for assemblages known to have equilibrated at pressures < 7 kbar. This discrepancy is thought to be the result of previous barometers using experiments conducted at pressures > 10 kbar for calibration, and therefore requiring extrapolation to lower pressures to investigate igneous equilibria within the crust. The most recent iteration of the Jd-in-Cpx barometer has been recalibrated using experiments conducted at crustal pressures (0.001-7 kbar), along with

higher pressure experiments (up to 20 kbar), to improve the accuracy of pressure estimation (Neave and Putirka, 2017).

This barometer requires the measurement of equilibrium liquid and clinopyroxene compositions, along with an equilibration temperature estimate to calculate equilibration pressure. Several of the experimental charges, detailed in previous sections, are saturated in clinopyroxene and their equilibration conditions are known, therefore these experiments can be used to test the Jd-in-Cpx barometer.

Figures 4.18a–c show the results of clinopyroxene-liquid barometry on clinopyroxenesaturated experiments. Pressure estimates have been calculated using the full range of clinopyroxenes measured in each experimental charge, along with the average glass composition for each experiment. As discussed in previous sections, this clinopyroxene variability is likely due to rapid crystallisation allowing disequilibrium compositions to crystallise. Clinopyroxenes in equilibrium with the glass should produce a more accurate pressure estimate than those out of equilibrium. These disequilibrium clinopyroxene compositions tend to have high Al and Na cation fractions, resulting in high pressure estimates, as shown by Figures 4.18a–c. At high equilibration temperatures, where crystallinity is low, the phases are more likely to be in equilibrium with the glass and therefore give more accurate pressure estimation. Due to the known experimental conditions of each charge it is possible to identify the equilibrium clinopyroxene composition for each charge, but this can be more difficult if pressure and temperature are unknown.

Figure 4.18a shows that for the 0.001 kbar experiments, pressure can be estimated within error of the barometer using the highest X_{DiHd} content clinopyroxenes equilibrated at 1200 °C, however the 1170 °C experiment overestimates pressure by a couple of kbars, even at highest X_{DiHd} . Either the equilibration clinopyroxene has not been measured, or there is glass compositional variability, again meaning that the measurements used in this barometer are not in equilibrium. Successful pressure recovery at higher temperature rules out a systematic pressure overestimation by the barometer, therefore it is likely that MID2014-12 compositions used are not quite in equilibrium. It is difficult to ascertain if there was zoning within the experimental charge indicating fractionation of phases due to the fragmentation of the basaltic sphere upon quenching from the Deltech gas mixing furnace.

Experiments equilibrated at 5 kbar recover pressure relatively well for both temperatures, with the highest X_{DiHd} content clinopyroxenes giving the best pressure estimates for the 1190 °C experiment. Clinopyroxenes with a similar composition as the 1190 °C experiment produce the best pressure estimates for the 1210 °C experiment (Figure 4.18b). These two experiments are known to have gained Na from the NaCl pressure medium during the experimental run, yet the clinopyroxene and glass compositions still allow for correct pressure



Figure 4.18 Clinopyroxene-liquid barometry using various jd-in-cpx parameterisations. Pressure estimates using Neave and Putirka (2017) barometer against clinopyroxene diopside-hedenbergite content for (a) 0.001 kbar (purple), (b) 5 kbar (red), and (c) 10 kbar (blue) experiments. (d)–(f) Comparison of Neave and Putirka (2017) barometer with other parameterisations.

estimation, suggesting that the two have remained in equilibrium despite compositional modification.

Figure 4.18c shows clearly that the highest X_{DiHd} content clinopyroxenes do not necessarily give the correct pressure estimate for the Jd-in-Cpx barometer. The 10 kbar experiment conducted at 1290 °C calculates equilibration pressure very well, with all clinopyroxene compositions plotting within the error of the barometer. This X_{DiHd} fraction (0.5) in experiments conducted at lower temperatures also recovers the correct pressure, however clinopyroxene compositions within these charges span pressures above and below this value. The highest X_{DiHd} contents (0.75) underestimate the pressure of equilibration by over 5 kbar.

The extreme clinopyroxene variability within low temperature experimental charges makes it difficult to be confident with the basic assumption that the highest X_{DiHd} measured is in equilibrium with the average glass composition, and therefore provides the most accurate pressure estimate. Testing of this barometer on hundreds of clinopyroxene-liquid pairs has concluded that the best test of equilibrium between the two phases is to look at the observed and predicted clinopyroxene end-member compositions (Putirka, 1999; Neave and Putirka, 2017). If the two match, then there is a good indication that the two are in equilibrium. For the experimental charges presented here it is clear that there have been some equilibration problems for lower temperature assemblages, particularly at 10 kbar, meaning that they cannot be reliably used for pressure are due to too many Na and Al cations within the clinopyroxene structure, produced by rapid clinopyroxene crystallisation and sector zoning. Pressure underestimates could be due to too little Na and Al in the clinopyroxene, or too much Na and Al in the glass.

Figures 4.18d–f compare various parameterisations of the jd-in-cpx barometer with the most recent calibration (Neave and Putirka, 2017). At low pressures the iterative combination of Neave and Putirka (2017) barometer and Putirka (2008a) thermometer agree with Neave and Putirka (2017) barometer calculated with known equilibration temperatures. However, at higher pressures the iterative combination produces larger estimates than the known temperature barometer. This overestimation is due to disequilibrium clinopyroxene compositions estimating high pressures, which are increased further by the higher temperature associated with this pressure and composition. The positive feedback between pressure and temperature is enhanced the most for the highest Al clinopyroxenes, which are most out of equilibrium and give the highest pressures. The previous calibrations of the jd-in-cpx barometers produce higher pressure estimates than the Neave and Putirka (2017) barometer for crustal pressures, indicating that the latest calibration has reduced the pressure overestimate for assemblages equilibrated at crustal pressures.

4.4.2 Olivine-plagioclase-augite-melt (OPAM) barometry

Yang et al. (1996) used a database of experimental olivine, plagioclase, augite (OPA) and equilibrium liquid compositions to describe the crystallisation of these three igneous phases as a function of pressure and liquid composition. The concentration of Mg, Ca, and Al within the melt was parametrised as a function of pressure and the other major element concentrations, essentially describing the movement of the OPA-liquid ternary point with changing pressure. This relationship between ternary equilibrium composition and pressure has been developed further as a barometer, by comparing calculated and observed liquid compositions to estimate equilibration pressures.

Kelley and Barton (2008) used these equations for liquid X_{Mg} , X_{Ca} , and X_{Al} contents to calculate equilibrium melt compositions at increments of 1 kbar for comparison with observed melt compositions. Their numerical method for estimating equilibrium pressure involves converting the calculated ternary liquids into normative mineral components, then using a regression to find the pressure dependence of these components, before rearranging the regression function to calculate pressure from observed normative components.

More recently the use of Yang et al. (1996) equations has been improved through the use of a χ^2 misfit function. Again equilibrium melt compositions are calculated, X_i^{calc} , for Mg, Ca, and Al as a function of pressure, and then compared to observed compositions, X_i^{obs} . For each pressure a χ^2 misfit can be calculated using:

$$\chi^2 = \sum_{i=1}^3 \left(\frac{X_i^{obs} - X_i^{calc}}{\sigma_i} \right)^2,$$

where σ_i is the analytical uncertainty based on EPMA measurement of the major element oxide. The χ^2 misfit value can be converted to probability using the χ^2 distribution, which essentially describes the quality of fit between observed and calculated compositions. The results of multiple glasses can be filtered based on the probability of their fit, refining the pressure estimates calculated from a suite of samples from the same eruption. This χ^2 misfit minimisation method was developed by John Maclennan (pers. comm.).

A new set of multiple linear regressions has been developed to include Fe^{3+} and Cr terms to the liquid X_{Mg} , X_{Ca} , and X_{Al} content equations (Voigt et al., 2017). The χ^2 misfit minimisation method for calculating equilibration pressure can also be applied to these equations.

The experiments conducted on Miðfell glass can be used to test how well the Yang et al. (1996); Kelley and Barton (2008); Voigt et al. (2017) parameterisations estimate pressure from basaltic liquid compositions. Also these experiments can provide testing for the quality of fit probability filter used in the χ^2 misfit minimisation method. Figures 4.19a–c show the

results of the Yang et al. (1996) parameterisation using all measurements of experimental glass and the χ^2 misfit minimisation method. Not all of the glass compositions are OPA-saturated, therefore this provides a test to the probability filter, which is set to p > 0.8, to see if pressures far from equilibration are discarded. The probability of fit is coloured for each point in these plots.

Glasses equilibrated at 0.001 kbar produce progressively lower pressure estimates as the MgO content of the liquid decreases, with estimates levelling off at ~ 1.5 kbar for the lowest MgO glasses, which are the lowest temperature and most crystalline (Figure 4.19a). This relationship clearly shows the importance of having all three OPA phases saturated to recover equilibration pressure. The data point colours indicate that the quality of fit increases as MgO decreases, however a probability filter of p > 0.8 would give a large range of pressure estimates (0.6–7.5 kbar); only discarding the highest pressures. Therefore, textural evidence of OPA phase saturation would be required to filter these pressure estimates further, rather than using probability alone. Some of the lowest MgO glasses have a lower quality of fit, suggesting that these charges had glass compositional heterogeneity and therefore a degree of disequilibrium.

Glass compositions equilibrated at 5 kbar are far less numerous, and were shown in previous sections to have gained Na. However, the highest probability pressure estimates, which occur in the more crystalline charges, recover \sim 5 kbar reasonably well with a slight underestimation of pressure (Figure 4.19b). This observation suggests that the OPAM barometer can be used on more alkali basalt compositions, however more systematic testing with variable alkali contents would be required to affirm this claim.

The 10 kbar glasses all cluster at high pressures with the best pressure recovery for the highest MgO glasses (Figure 4.19c). These glasses show the lowest quality of fit and are known to not be saturated in all three OPA phases. The lower MgO glasses drift to higher pressures and probabilities, but now overestimate equilibration pressure. Given the stratified nature and high crystallinity of these experimental charges, the glasses measured might not be in equilibrium with all three OPA phases, resulting in higher pressure estimates as shown by the lower variance assemblages at lower pressure (Figure 4.19a–b).

Pressure estimates using Voigt et al. (2017) equations, which contain terms for Fe³⁺ and Cr, combined with the χ^2 misfit minimisation method are shown in Figures 4.19d–f. The results have the same overall structure as the Yang et al. (1996) parameterisation results, but far fewer are calculated as having a high quality of fit. This means that the probability filtered 0.001 kbar glasses recover an average pressure much closer to the experimental pressure than for Yang et al. (1996), however 5 kbar glass pressures are underestimated and for 10 kbar



Figure 4.19 Olivine–plagioclase–augite–melt (OPAM) barometer calibrations used to estimate crystallisation pressures from experimental liquid compositions. χ^2 misfit minimisation method using Yang et al. (1996) for (a) 0.001 kbar, (b) 5 kbar, and (c) 10 kbar, and Voigt et al. (2017) for (d) 0.001 kbar, (e) 5 kbar, and (f) 10 kbar. Colour of plotted circle indicates fit quality or pressure probability. Pale green region shows liquid compositions that are OPA-saturated. Kelley and Barton (2008) parameterisation for (h) 0.001 kbar, (i) 5 kbar, and (j) 10 kbar.



Figure 4.20 Pressure estimates using OPAM parameterisations plotted as kernel density functions for (a) Yang et al. (1996), (b) Kelley and Barton (2008), and (c) Voigt et al. (2017). Glasses equilibrated at pressures of 0.001 kbar (purple), 5 kbar (red), and 10 kbar (blue) are plotted separately. Voigt et al. (2017) pressure estimates of glasses equilibrated at 0.001 kbar with varying (d) Fe^{3+} , and (e) Cr contents are also shown.

they are overestimated. The addition of Fe³⁺ and Cr terms to the X_i^{calc} equations seems to produce larger residuals, giving lower probabilities of fit.

Kelley and Barton (2008) pressure estimates are not filtered by probability. This parameterisation reproduces 0.001 kbar well, particularly at the lower glass MgO contents, however with no probability filter, textural observations are required to confirm the presence of OPA phases, and therefore confirm the validity of the pressure estimate (Figure 4.19g). Higher equilibration pressure glasses show much more variability in the pressures calculated from their compositions (Figures 4.19h–i). OPA-saturated glasses give a 12 kbar pressure estimate range for the 5 kbar equilibration pressure, and overestimate pressure at 10 kbar.

Figure 4.20 compares the pressure estimates from the three OPAM parameterisations. Without a way of filtering the Kelley and Barton (2008) parameterisation pressure estimates, the kernel density functions for the three equilibration pressures overlap and span 0–15 kbar, which is greater than the thickness of the Icelandic crust. Therefore, this method of pressure

estimation seems unreliable for use on natural samples. The Yang et al. (1996) equations overestimate 0.001 kbar and 10 kbar glasses, but slightly underestimate those equilibrated at 5 kbar. Voigt et al. (2017) is similar for 5 kbar glasses, again slightly underestimating the pressure. The 0.001 kbar glasses predict a low pressure of equilibration (< 1 kbar), but 10 kbar glasses are overestimated by over 4 kbar. It is therefore expected that at equilibration pressures > 4 kbar Voigt et al. (2017) will calculate higher pressures than Yang et al. (1996).

Figures 4.20d–e show the effect of changing Fe^{3+} and Cr contents of glasses run through the Voigt et al. (2017) parameterisation. Addition of both Fe^{3+} and Cr to the glass decreases the pressure estimate recovered from the OPAM barometer. Fe^{3+} is difficult to measure, so not all glasses will have appropriate Fe^{3+} data to satisfy the Voigt et al. (2017) parameterisation. Omission of Fe^{3+} from the pressure calculation will likely result in an overestimate of the equilibration pressure for a natural sample. Cr has more of an effect on pressure estimation than Fe^{3+} , as addition of 0.1 wt% Cr_2O_3 decreases pressure estimates by ~ 4 kbar (Figure 4.20e), while addition of 1.0 wt% Fe_2O_3 decreases pressure estimates by ~ 2 kbar. This suggests that quantification of Fe^{3+} and Cr is very important when using the Voigt et al. (2017) parametrisation.

When applying the OPAM barometer to natural samples it is important that the glass measured is in equilibration with all three OPA phases. Textural observations are key to confirming this and the validity of the pressure estimate calculated by the barometer.

4.4.3 Multi-reaction barometry

The barometers presented earlier in this section used single reaction equilibria between igneous phases to calculate a pressure estimate for the clinopyroxene-liquid pair or OPAM assemblage. However, multi-reaction barometry uses the position of equilibrium for several reactions in combination to establish an average pressure estimate. This method has been used in metamorphic petrology for several years to establish the equilibration pressure-temperature conditions of mineral assemblages (Holland and Powell, 1998; Powell and Holland, 2008; Holland et al., 2013). The Holland and Powell (2011) thermodynamic dataset is used with various mineral activity-composition (a-x) models to calculate equilibrium conditions for a given mineral assemblage. Recent advancements to a-x models for mafic minerals have extended the capabilities of average pressure (avP) calculations to mafic igneous compositions.

Multi-reaction barometry in mafic igneous systems uses reactions between olivine, plagioclase, clinopyroxene, and spinel solid solution end-members (Ziberna et al., 2017). Calibration of this barometer required the modification of previously published *a-x* models for olivine (Jennings and Holland, 2015), plagioclase (Holland and Powell, 2011), clinopyroxene (Green et al., 2016), and spinel (Bryndzia and Wood, 1990), to get calculated pressure estimates of gabbros to match known experimental conditions. Experiments conducted at 0.001 kbar were not used in the calibration process due to the high level of clinopyroxene variability, and therefore disequilibrium within these charges. This technique targets gabbroic assemblages, therefore it has been calibrated at pressures typical of gabbro formation (Ziberna et al., 2017).

Compositional measurements of the igneous phases listed above are converted into endmember activities using *a-x* models. These activities are combined with the Holland and Powell (2011) thermodynamic dataset in THERMOCALC to calculate the equilibrium pressures of a number of end-member reactions, which combined give an average pressure estimate. The significance of this pressure estimate is dependent on how well the multiple reactions agree within the uncertainty of the mineral measurements. An estimate of temperature is required for these calculations, with Ziberna et al. (2017) using the Ca-in-olivine thermometer of Shejwalkar and Coogan (2013), but the Opx-Cpx thermometer of Putirka (2008b) could be used as an alternative. Ziberna et al. (2017) presents two barometers, the choice of which is dependent on the igneous phases in equilibrium; clinopyroxene, olivine, and plagioclase either with or without spinel. As spinel has not been measured in any of the Miðfell experimental charges, only the clinopyroxene-olivine-plagioclase barometer is tested below.

Clinopyroxene-olivine-plagioclase average pressure (COIP avP)

For a three-phase assemblage of clinopyroxene-olivine-plagioclase, three end-member reactions can be written between these phases:

$$\begin{split} Mg_2Si_2O_6 + CaAl_2SiO_6 &= Mg_2SiO_4 + CaAl_2Si_2O_8\\ Fe_2Si_2O_6 + CaAl_2SiO_6 &= Fe_2SiO_4 + CaAl_2Si_2O_8\\ Mg_2Si_2O_6 + NaAlSiO_6 &= Mg_2SiO_4 + NaAlSi_3O_8 \end{split}$$

where $Mg_2Si_2O_6$ is high-T clinoenstatite (Cenh), $CaAl_2SiO_6$ is Ca-Tschermak clinopyroxene (CaTs), Mg_2SiO_4 is forsterite (Fo), $CaAl_2Si_2O_8$ is anorthite (An), $Fe_2Si_2O_6$ is clinoferrosilite (Cfs), Fe_2SiO_4 is fayalite (Fa), $NaAlSiO_6$ is jadeite (Jd), and $NaAlSi_3O_8$ is high-T albite (Abh). Use of these three reactions to produce average pressure estimates is thought to underestimate pressure by ~ 1.5 kbar for pressures < 5 kbar, but reproduce higher pressures reasonably well (Ziberna et al., 2017).

Miðfell experiments that saturated in clinopyroxene, olivine, and plagioclase have be used to test the multi-reaction barometry calibration. Figures 4.21a–c show average pressure estimates calculated from MID2014-12 clinopyroxene, average olivine and average



Figure 4.21 Estimates of pressure using multi-reaction barometry on experimental olivine–plagioclase– clinopyroxene assemblages. Pressure estimates for mid12 (1170 °C, 0.001 kbar) against (a) X_{DiHd} , (b) X_{CaTs} , and (c) X_{Jd} content of the clinopyroxenes used in the calculation. Mid23 (1240 °C, 10 kbar) pressure estimates against the same clinopyroxene end-members in (d), (e) and (f).

plagioclase compositions. Individual olivine and plagioclase analyses were not used as (i) EMPA measurements were not made in systematic three phase groupings, and (ii) the compositional variability of olivine and plagioclase is not as significant as clinopyroxene variability. However, there is still a chance that these mineral compositions will not be representative of any of the compositions in equilibrium with clinopyroxene.

As clinopyroxene compositions approach expected equilibrium values (high X_{DiHd} , low X_{CaTs} , low X_{Jd}) then the estimated pressure from avP calculations is within error of the experimental equilibration pressure (Figures 4.21a–c). Disequilibrium clinopyroxene compositions, Al-rich, produce pressure estimates with very large uncertainties and significance of fit values, indicating that these pressure estimates are unreliable. Ziberna et al. (2017) chooses to ignore 0.001 kbar experiments due to the strong effect disequilibrium Al contents can have on calculating pressure, which have been shown to produce large pressure overestimates (Fig. 4.21a-c). However, if the clinopyroxene composition is carefully chosen, which is more difficult with natural samples, then the avP method can reproduce low equilibration pressure estimates.

			Experim	nental T	1		Calcula	ted T	
Experiment	Р	T _{exp.}	avP	σP	sigfit	T _{calc.}	avP	σP	sigfit
mid20 mid12	0.001 0.001	1200 1170	2.32 10.25	1.85 6.15	0.79 0.94	1311 1295	1.22 10.02	1.98 6.68	0.53 0.93
mid34	5	1190	8.54	4.92	1.30	1286	8.07	5.00	1.24
mid14 mid27 mid23	10 10 10	1280 1270 1240	6.52 6.94 6.96	2.38 2.22 2.97	0.25 0.58 0.16	1323 1247 1238	6.36 7.00 6.96	2.45 2.19 2.97	0.20 0.59 0.16

Table 4.2 Pressure estimates from multi-reaction geobarometry using average olivine, plagioclase and clinopyroxene compositions for each three-phase saturated experiment. Pressures calculated using the known experimental temperature (T_{exp}) and calculated using the Ca-in-olivine thermometer. Uncertainty estimates of each pressure given by σP and an estimate of reliability by sigfit.

Figures 4.21d–f show average pressure calculations for MID2014-23 clinopyroxenes, again with average olivine and plagioclase compositions. Unlike the 0.001 kbar experiment, the highest X_{DiHd} clinopyroxenes do not produce the closest pressure estimates to the known experimental conditions.

Slightly lower compositions, $X_{DiHd} = 0.50$, produce estimates with the lowest uncertainty. The pressure estimate may be an underestimate of the true equilibrium value, but it is within the uncertainty of the calculation. The clinopyroxene composition for this estimate matches the clinopyroxene producing the most accurate pressure estimate for the clinopyroxene-liquid barometer, further supporting that the equilibrium clinopyroxene composition for the 10 kbar experiments is $X_{DiHd} = 0.50$. However, the high crystallinity of this experiment, coupled with some variability in olivine and plagioclase compositions suggest that perfect equilibrium of the phases is not a correct assumption. Therefore, the average pressure method may be working correctly, but our assumed equilibrium assemblage is wrong, hence the underestimation of pressure.

Pressure estimates calculated using average mineral compositions for experiments containing clinopyroxene-olivine-plagioclase are detailed in Table 4.2. Both the known experimental equilibration temperature and Ca-in-olivine thermometer temperature have been used to calculate average pressure. Calculated temperatures tend to be higher than known equilibrium conditions, therefore pressure calculations with these values produce lower pressure estimates. The 0.001 kbar experiment with the lowest clinopyroxene compositional variability (MID2014-20; 1200 °C) produces the pressure estimate closest to the true experimental pressure. Experiment MID2014-34 has a large uncertainty associated with its pressure estimate, which could be due to Na-gain giving unusual mineral compositions, for example Figure 4.7a shows that olivines within this assemblage have high forsterite contents ($X_{Fo} > 0.90$). All three 10 kbar experiments give consistent pressure estimates, however they underestimate the experimental pressure by \sim 3 kbar. These estimates are close to being within error of the true pressure, but again perhaps the assumption that the mineral compositions used are in equilibrium is not valid.

Improved average pressure estimation is possible if spinel is present within the assemblage as it (i) provides three more end-member reactions to produce pressure estimates, and (ii) hosts Al, potentially reducing the amount of high-Al disequilibrium clinopyroxene (Ziberna et al., 2017).

4.4.4 Summary of igneous barometers

The results from testing the three igneous barometers presented in the sections above clearly demonstrate the importance of being able to identify and use equilibrium phase compositions when attempting barometry. The equilibrium composition of clinopyroxene is crucial to the barometers considered here, with disequilibrium clinopyroxenes displaying anomalously high Al contents producing pressure estimates significantly different from true equilibration conditions, often overpressure estimates.

Clinopyroxene-liquid barometry requires the careful choice of clinopyroxene composition, however this can be estimated through calculation of expected clinopyroxene end-member components (?Neave and Putirka, 2017). In general the observed clinopyroxene compositions that were able to estimate the true equilibrium pressure, were close to predicted clinopyroxene compositions. Comparison of calculated and observed clinopyroxene components should be used to help validate pressure estimates from clinopyroxene-liquid barometry. Textural observations of clinopyroxene and glass pairs can be used to compliment the results of end-member comparison to support equilibrium between the two phases.

The OPAM barometer parameterisation of Yang et al. (1996) used with the χ^2 misfit minimisation method of Maclennan (pers. comm.) produces more accurate pressure estimates when a probability filter is applied to pressure estimates. The Kelley and Barton (2008) parameterisation, which has no such probability filter, was unreliable at estimating equilibration pressures > 0.001 kbar. The recent Voigt et al. (2017) OPAM parameterisation seems to work best for low pressures of equilibration, while Yang et al. (1996) is more suited to higher pressures (> 5 kbar), both in combination with the χ^2 misfit minimisation method.

Multi-reaction barometry requires equilibrium between three or four crystalline igneous phases to calculate pressure. This equilibrium condition can be quite difficult to confirm when analysing a volcanic rock with phenocrysts, however it is easier for plutonic rocks. Estimates are believed to be more reliable if spinel has been measured, as it allows for extra reactions to be involved in the average pressure calculation. For the experimental charges presented here, this multi-reaction barometer was restricted to just three phases. Again equilibrium clinopyroxene composition was key, producing the most accurate pressure estimates at expected clinopyroxene end-member compositions, however pressures were underestimated in charges equilibrated at 10 kbar.

4.5 Application to natural Miðfell samples

The previous section focused on the testing of igneous barometers using the results of experiments conducted on basaltic glass from the Miðfell eruption. Here the same barometers are applied to natural samples, providing pressure constraints on the magmatic plumbing system beneath Miðfell.

The majority of previous pressure estimates from Miðfell have been calculated using clinopyroxene and liquid compositions from gabbroic xenoliths found within one layer of the eruption (Gurenko and Sobolev, 2006). Other authors used clinopyroxene-liquid barometry to provide a low pressure estimate for Miðfell, 0.4 ± 1.2 kbar, suggesting that the final equilibration between these two phases occurred shallow in the crust, potentially as low as subglacial depths (Danyushevsky et al., 1996; Gurenko and Sobolev, 2006). However, single-pyroxene barometry calculates a greater pressure, 3.4 ± 1.6 kbar, which has been interpreted as equilibration during initial clinopyroxene crystallisation deep in the crust and at lower pressures en route to eruption (Nimis, 1995; Gurenko and Sobolev, 2006). This pressure estimate has been supported by an olivine-hosted fluid inclusion study and OPAM barometry, which give pressure estimates of 2.2–3.0 kbar and 1.0 kbar, but the same range of pressures has not been calculated by others (Hansteen, 1991; Kelley and Barton, 2008).

Equilibration temperatures have previously been estimated as 1210 ± 15 °C, calculated empirically using olivine-liquid and clinopyroxene-liquid thermometers (Ford et al., 1983; Danyushevsky, 2001, and references therein). OPAM thermometry, as parameterised by Yang et al. (1996), produced slightly lower temperature estimates, 1187 ± 15 °C (Kelley and Barton, 2008).

4.5.1 Thermometry

The olivine-liquid thermometer of Sugawara (2000) has been used on glass compositions from this study, as well as glasses from Gurenko and Chaussidon (1995); Gurenko and Sobolev (2006), which are all are assumed to be in equilibrium with olivine, so can be used with this thermometer. Temperature estimates from this thermometer are expected to represent the temperature of the carrier liquid when it last equilibrated with olivine. Temperature

estimates range from ~ 1220–1250 °C (Table 4.3). All glasses associated with gabbroic xenoliths, from interstitial to pillow rim glasses, give the same temperature estimate within error of the thermometer (~ 1225 °C). Pillow glasses from this study give a slightly higher temperature estimate of 1236 °C, but again within thermometer uncertainty. Recalculation of this estimate increases to 1250 °C when an alternative Sugawara (2000) formulation is used, which requires multiple glass oxide components and a pressure estimate.

The Ca-in-olivine thermometer from Shejwalkar and Coogan (2013) produces a similar temperature of 1236 °C for olivine phenocrysts sampled in this study. Temperature estimates from this thermometer could represent the temperature at which (i) olivine first crystallised, if core compositions are used, (ii) olivine last equilibrated with melt if rim compositions are used, or (iii) a later diffusion event occurred to modify the olivine chemistry. Olivines from gabbroic nodules record a lower temperature of 1165 °C (Risku-Norja, 1985). This lower temperature estimate could represent cooler equilibrium conditions for gabbroic xenoliths, or it is more likely that the CaO and other major element oxides were poorly constrained. Apart from thermometry on the xenolith olivines, all other temperature estimates are in good agreement with each other, suggesting that prior to eruption Miðfell liquid was at 1225 ± 15 °C.

4.5.2 **OPAM barometry**

The OPAM parameterisation of Yang et al. (1996) has been used on all Miðfell glass compositions measured in this and previous studies. The Voigt et al. (2017) barometer equations have only been used on measurements from this study, as it is the only set of glass analyses with Fe^3 + and Cr concentrations. The results of barometer testing suggest that the Kelley and Barton (2008) parameterisation is unreliable, so has not been used on natural samples.

The most recent Miðfell glass measurements presented in this study record the greatest equilibration pressures according to the OPAM barometer (Figure 4.22a; Table 4.3), suggesting olivine-plagioclase-augite-melt equilibration deep in the crust (6.7 kbar; 25 km). As predicted for equilibration pressures > 5 kbar, the Voigt et al. (2017) parameterisation predicts a higher pressure than the Yang et al. (1996) equations for the same set of glass compositions. Previously measured glass compositions, some of which are associated with gabbroic xenoliths, produce lower pressure estimates (Gurenko and Sobolev, 2006). Estimates range from ~ 4.3 kbar for pillow glasses to 2.4 kbar for interstitial glasses within the gabbroic xenoliths. This progressive decrease in pressure associated with moving from carrier glass into the xenolith, could be due to re-equilibration between the liquid and xenolith assemblage during transport of the xenolith to the surface, i.e. during decompression.

ad are: (i) olivine-plagioclase-augite-melt Yang et al. (1996); Voigt et al. (2017); (ii) clinopyroxene-liquid (Neave and Putirka, 2017); and (iii) alti-reaction average P using spinel-clinopyroxene-olivine-plagioclase and clinopyroxene-olivine-plagioclase (Ziberna et al., 2017). Uncertainty	able 4.3 Temperature and pressure estimates of equilibration for a range of natural Midfell samples. Thermometers used are: (i) MgO-in-liquid
	ulti-reaction average P using spinel-clinopyroxene-olivine-plagioclase and clinopyroxene-olivine-plagioclase (Ziberna et al., 2017). Uncertainty

Reference	Sample	${ m T}_{ m Sugawara2000}$ $(\pm 30~^{\circ}{ m C})$	$T_{Putirka2008}$ (± 45 °C)	${ m T}_{ m Shejwalkar2013} (\pm 20~^{\circ}{ m C})$	Popam	${ m P}_{ m Neavel7}$ (\pm 1.4 kbar)	$\begin{array}{l} P_{avP \; SCOIP} \\ (\pm \; kbar) \end{array}$	$\begin{array}{l} P_{avP \ COIP} \\ (\pm \ kbar) \end{array}$
This study	Pillow glass	1236; 1250 ^a	· ·	1	$6.74 \ ^{b}; 7.30 \ ^{c}$	1		
	lO	I	ı	1236		ı	ı	ı
	Melt inclusions	I	I	I	I	I	I	I
Gurenko	Pillow glass	1230		1	4.33	1		
(1995,2006)	Glass crust	1225	I	1	3.01	ı	·	'
	Interstitial glass	1228	ı	ı	2.39	ı	ı	'
	Cpx-embayment	1221	1238	ı	'	4.21	ı	'
	Xenolith ^d	I	I	1236	I	I	I	5.56 (4.00)
Risku-Norja	Xenolith	1	1	1164	1	1	3.21 (0.93)	2.94 (1.24)
(1985)	Phenocrysts	I	I	1165	1	ı	4.16 (0.87)	4.08 (1.02)
	Xenolith ^e	'	ı	1236	'	'	2.65 (0.94)	2.36 (1.08)
	Phenocrysts e	ı	I	1236	ı	ı	3.66 (0.90)	3.57 (0.94)

^c Average pressure estimate using Voigt et al. (2017) parameterisation. ^d Temperature and pressure estimated using olivine phenocryst measurements from this study in combination with Gurenko and Sobolev (2006) clinopyroxene and gabbro measurements. ^e Pressures recalculated using the Sugawara (2000) ^a Temperature estimate using multiple liquid components and the OPAM pressure estimate. ^b Average pressure estimate using Yang et al. (1996) parameterisation. temperature estimate from this study.



Figure 4.22 Pressure estimates from natural Miðfell samples plotted as kernel density functions using (a) Yang et al. (1996); Voigt et al. (2017) OPAM barometry, and (b) clinopyroxene-liquid barometry (Neave and Putirka, 2017) and multi-reaction barometry (Ziberna et al., 2017). Compositions from this study and Risku-Norja (1985); Gurenko and Chaussidon (1995); Gurenko and Sobolev (2006).

If there has been significant interaction between xenolith and interstitial glass then the OPAM barometer method loses some validity, as the xenoliths studied are not observed to contain olivine, though olivine is seen in other Miðfell xenoliths (Gurenko and Sobolev, 2006). Therefore, the requirement for OPAM barometer liquid to be in equilibrium with all three OPA phases is no longer fulfilled. Pressure estimate differences between pillow glass and interstitial glass might not be the direct result of pressure on an OPAM assemblage, but rather the interaction of liquid with clinopyroxene and plagioclase.

It is strange that the two suites of pillow glasses give such different pressure estimates, the main difference in their chemistry is that glasses from this study record lower SiO₂ (0.5 wt%) and CaO (1.0 wt%) than Gurenko and Sobolev (2006) glasses. Perhaps this difference, which could be due to true melt heterogeneity or analytical uncertainty, is enough to produce an average pressure discrepancy of ~ 2 kbar. The two suites are from different parts of the Miðfell eruption, as the basalts sampled for this study did not contain gabbroic xenoliths, therefore the pillow glasses from the xenolith layer might have been slightly modified by

xenoliths. If the lowest OPAM pressures can be trusted, then perhaps they record the pressure pertaining to a shallow crustal magma chamber within which the melt resided before eruption. While higher pressures indicate deep crustal equilibration and potential magma reservoirs.

4.5.3 Clinopyroxene-liquid barometry

Barometer testing showed the importance of using equilibrium phases for igneous barometry, therefore only clinopyroxene-melt embayment pairs have been used with the clinopyroxene-liquid barometer (Gurenko and Sobolev, 2006). No clinopyroxene phenocrysts were measured in this study, so the Miðfell clinopyroxene dataset has not been extended.

To ensure that equilibrium compositions were used in this barometer, clinopyroxene analyses with cation totals < 3.99 or > 4.02 were discarded, along with any containing $X_{Jd} < 0.01$ as this is below the EPMA detection limit (Putirka, 2008b; Neave and Putirka, 2017). Observed end-member proportions were also compared to expected end-member compositions to check that the two matched (Putirka et al., 2003; Neave and Putirka, 2017).

The clinopyroxene-liquid pairs presented by Gurenko and Sobolev (2006) were used to produce a range of pressure estimates, with a main grouping around 5 kbar, but extending to lower pressures (Figure 4.22b). The distribution of pressure estimates does not change if pressures are recalculated using interstitial or xenolith crust glass compositions instead of embayment glasses. The range of pressures again suggests re-equilibration between clinopyroxene and melt during decompression, though not all pairs have been reset at the most shallow depths. The highest pressures estimated by clinopyroxene-liquid barometry are similar to the OPAM barometer estimates for pillow glasses from the xenolith-bearing layer of Miðfell. This suggests that initial interaction between melt and gabbro occurred at pressures around 5 kbar, or depths > 15 km.

4.5.4 Multi-reaction barometry

Average pressure multi-reaction barometry has been carried out on published spinel, clinopyroxene, olivine, and plagioclase compositions from Risku-Norja (1985), using phenocryst and gabbroic xenolith phases. Xenolith phases, using average compositions, produce an equilibrium pressure of 3.21 ± 0.93 (2.94 ± 1.24 using COIP avP), while phenocryst phases give 4.16 ± 0.87 (4.08 ± 1.02 using COIP avP). Firstly, these values highlight the fact that COIP avP produces slightly lower pressure estimates with higher uncertainty than SCOIP avP. Secondly, it is unclear to what extent these phases are in equilibrium with each other. The uncertainty and significance of fit associated with these estimates are within the limits of acceptable values, suggesting equilibrium. However, Risku-Norja (1985) does not state clearly which phase measurements are in association with others. The temperature estimate used for these pressure calculations is 1165 °C, which is lower than the other temperature estimates for Miðfell. Recalculation of avP using the temperature 1236°C lowers the pressure estimates slightly (Figure 4.22b; Table 4.3). These pressures roughly match those estimated by OPAM on xenolith glasses and some of the clinopyroxene-liquid barometry.

Plagioclase and clinopyroxene compositions from gabbroic xenoliths have been combined with olivine phenocryst compositions, measured as part of this study, to provide another pressure estimate using COIP avP, 5.56 ± 4.00 (Gurenko and Sobolev, 2006). The large uncertainty and significance of fit reflects the fact that xenolith phases and olivine phenocrysts did not crystallise together and are not in equilibrium. However, disequilibrium has not created a pressure estimate that is unreasonable for Miðfell. Olivine compositions from Risku-Norja (1985) have also been combined with Gurenko and Sobolev (2006) clinopyroxene and plagioclase to give a pressure estimate of 5.99 ± 3.58 . The difference in pressure is mainly due to the lower equilibrium temperature estimate from the Ca-in-olivine thermometer, however the uncertainty on the estimate remains high, again indicating a non-equilibrium assemblage of phases.

Multi-reaction barometry has the potential to be very useful in providing igneous pressure estimates, however the method relies upon equilibrium between three or four solid phases. This equilibrium can be difficult to confirm in volcanic rocks, so gabbroic xenoliths are perhaps the best sample for pressure estimation. Phenocryst compositions could be providing pressure information, though they are much more likely to undergo re-equilibration during decompression. Cores of minerals, assumed to be in equilibrium during initial crystallisation are perhaps the best candidates for recording deep magma reservoir pressures.

4.6 Conclusions

The experiments presented in this chapter are inappropriate for a full re-calibration of the clinopyroxene-liquid barometer (Putirka et al., 1996; Neave and Putirka, 2017), however they have been used to test a number of igneous barometers and crystallisation models. Caution was required when interpreting any results involving the 5 kbar experiments as they have undergone Na-gain and Fe-loss during the experimental procedure. Some of the 10 kbar experiments appear to have equilibrated at pressure-temperature conditions different to those of the experimental run, however the 0.001 kbar have equilibrated well and shown expected crystallisation trends.

Clinopyroxene compositions are highly variable within the lowest temperature experiments presented here. This variability is likely due to the rapid crystallisation of clinopyroxene well below the bulk liquidus temperature, and in part the result of melt segregation upon heating. It is important that the correct clinopyroxene compositions are used for thermobarometry as disequilibrium high-Al contents cause pressure and temperature to be poorly estimated. Expected clinopyroxene end-member proportions and K_D values can be compared to those observed to identify equilibrium clinopyroxene compositions for use in thermobarometry.

Testing of Petrolog with the experimental starting bulk composition showed that the program did well at 0.001 kbar to recreate observed mineral chemistry, but was poor at higher pressure. This discrepancy could be due to the uncertainties associated with the 5 and 10 kbar experiments, rather than Petrolog itself. MELTS performed more poorly, mainly due to the prediction of high levels of clinopyroxene crystallisation at high pressures.

The igneous barometers tested were able to reproduce equilibrium pressures within the uncertainty of their parameterisations using experimental compositions. The Yang et al. (1996) OPAM equations coupled with the χ^2 misfit minimisation method worked well, as did the Voigt et al. (2017) parameterisation, however it requires measurement of ferric iron. The clinopyroxene-liquid barometer of Neave and Putirka (2017) was able to estimate equilibrium pressures well, but required the careful choice of equilibrium clinopyroxene composition, otherwise pressure was often overestimated. Multi-reaction average pressure barometry was less accurate at estimating pressure than the other methods, but again required careful identification of the equilibrium assemblage.

Use of the barometers on natural Miðfell samples suggests that melt could have been in equilibrium with all three OPA phases at a pressure of 6.7 kbar (20 km), however barometry methods involving the use of gabbroic xenolith compositions predicted lower pressures of 2.7–4.3 kbar (8–13 km). Overall these pressure estimates suggest that crystallisation happened deep in the crust beneath Miðfell (12–20 km), and could have been the depth of entrapment for olivine-hosted melt inclusions. Magma and gabbroic xenoliths were then transported to shallower depths, where xenoliths could also have been acquired, before eruption.

Chapter 5

Geochemical heterogeneity within olivine-hosted melt inclusions from Kistufell and Miðfell

5.1 Introduction

Observations of high ${}^{3}\text{He}/{}^{4}\text{He}$ ratios (> 8 R/R_a) in certain Icelandic eruptions, such as Kistufell and Miðfell, have long been held as evidence for the presence of a primordial noble gas reservoir beneath Iceland, which has been brought into the Icelandic melting region by a mantle plume (Harrison et al., 1999; Breddam et al., 2000; Brandon et al., 2007; Füri et al., 2010). The primordial nature of Miðfell He has encouraged other studies to access the heavy noble gas signature of this eruption, such detailed study makes it uniquely well-characterised among Icelandic eruptions (Harrison et al., 1999; Trieloff and Kunz, 2005; Füri et al., 2010; Mukhopadhyay, 2012). Ne isotopic ratios that resemble solar values, and low radiogenic to non-radiogenic Xe isotopes further support a relatively undegassed primordial reservoir in the Miðfell source mantle.

The primordial origin of noble gases from Miðfell, and to a certain extent Kistufell, raises the question of whether these eruptions could be preserving lithophile and volatile element signatures that can also be attributed to an ancient lower mantle reservoir. Other Icelandic eruptions show lithophile element variability and enrichment that has been interpreted as the result of melting different mantle lithologies, such as recycled oceanic crust pyroxenite or primitive mantle peridotite (Thirlwall, 1995; Stracke et al., 2003; Maclennan, 2008b; Shorttle and Maclennan, 2011; Shorttle et al., 2014). Measurements of volatile elements have been combined with incompatible trace element (ITE) concentrations to infer the volatile content

of Icelandic source mantle, and at other localities along the mid-ocean ridge (MOR) system (Michael, 1995; Saal et al., 2002; Michael and Graham, 2015; Le Voyer et al., 2017; Hauri et al., 2017).

Measurements of olivine-hosted melt inclusions are central to several of the studies referenced above, as they can preserve more compositional variability than related lavas and therefore are assumed to represent melt heterogeneity present deep within the magmatic system of an eruption (Maclennan, 2008a). This process gives olivine-hosted melt inclusions suites the potential to preserve primitive melt compositions, which could reflect both the variability and enrichment of mantle melts. Observed lithophile element variability within melt inclusion suites is much lower than predicted primary variability from mantle fractional melting models, indicating a reduction in variability between melt formation and entrapment (Rudge et al., 2013; Jennings et al., 2017). This chapter sets out to investigate the processes controlling the extent of lithophile and volatile element variability preserved in olivine-hosted melt inclusions.

Several processes are known to be responsible for the modification of lithophile and volatile element concentrations within melts and melt inclusions, occurring from initial melt formation in the mantle to eruption at the surface. These processes can be disentangled by working backwards from eruption and lava solidification to identify the contribution of each process to melt variability observed at the surface.

Post-entrapment crystallisation (PEC) modifies both lithophile and volatile element concentrations upon cooling and decompression of the host olivine, as olivine crystallises from the trapped melt on the walls of the inclusion (Danyushevsky et al., 2002; Kress and Ghiorso, 2004). Changes to volatile concentrations, coupled with decompression can lead to vapour bubble formation, olivine decrepitation, and potentially partial degassing of the melt inclusion, with CO₂ particularly susceptible to this process (Lowenstern, 1995; Aster et al., 2016; Maclennan, 2017). Modification of H₂O can also occur by diffusive re-equilibration of H⁺ ions between the carrier liquid and melt inclusion through the olivine lattice (Gaetani et al., 2012; Bucholz et al., 2013; Hartley et al., 2015).

Volatile degassing has the potential to modify melt concentration prior to entrapment within minerals, with CO_2 the most prone to this process due to its low solubility within silicate melt (Dixon and Stolper, 1995; Shishkina et al., 2010, 2014). CO₂ saturation and subsequent degassing means that dissolved CO_2 content can be at saturation, and therefore reflect saturation pressure, which can be estimated from CO_2 content using solubility models (Witham et al., 2012; Newman and Lowenstern, 2002). Within magma reservoirs in the crust, the process of concurrent melt mixing and crystallisation causes melt evolution, with melt heterogeneity decreasing as compositions are mixed together (Maclennan et al., 2003a;

Maclennan, 2008a; Neave et al., 2013). Also within the crust, the assimilation of crustal material mainly affects lithophile elements, though volatiles can also be modified if the assimilated material is volatile-bearing (Nicholson et al., 1991; Gurenko and Sobolev, 2006; Eason and Sinton, 2009; Brounce et al., 2012).

The main processes responsible for melt modification in the mantle are intimately linked; source heterogeneity and fraction of melting. Some lithologies contributing to source heterogeneity are the result of previous mantle melting events, with residues providing depleted refractory peridotites, and melts resulting in enriched compositions. Source enrichment ITE signatures are therefore similar to those of low fraction melts, making the two difficult to decipher, even when isotope data is available. However, some ITEs can be modified during the recycling process, for example large ion lithophile elements, such as Sr and Rb, are fluid mobile and could be depleted from recycled crustal material during subduction.

Fractional melting of a heterogeneous mantle source is expected to produce a diverse range of melts. Chapter 6 investigates carbon heterogeneity within the Miðfell mantle source region, but it requires the careful assessment of crustal and mantle melt modification processes discussed in this chapter to support its conclusions. This chapter begins by outlining the petrological observations made on the Miðfell and Kistufell eruptions, before discussing the processes responsible for creating the compositional variability observed within these two melt inclusion suites. It concludes with a wider assessment of trace element variability and enrichment within Icelandic olivine-hosted melt inclusion suites.

5.2 Petrological observations

This section presents textural and geochemical observations of whole rock, matrix glass, phenocrysts, and xenoliths from the Miðfell and Kistufell eruptions to provide a petrological context for the olivine-hosted melt inclusion measurements discussed within this chapter. The observations below have been made from thin sections of Miðfell pillow basalt, gabbroic xenoliths, and published data (Risku-Norja, 1985; Trønnes, 1990; Gurenko and Chaussidon, 1995; Gurenko and Sobolev, 2006). The Kistufell eruption has been characterised using pillow basalt rims and published data (Breddam, 2002). Melt inclusion analyses have been carried out on olivines picked from pillow basalt rims to ensure that melt inclusions are glassy, and have therefore undergone a minimum amount of compositional modification by post-entrapment crystallisation.



Figure 5.1 Photomicrographs of olivine-hosted melt inclusions from Kistufell in (a) plane-polarised light (ppl), (b) reflected light (rl); from Miðfell in (c) ppl, (d) crossed-polarised light (xpl), (e) rl, and (f) showing an olivine with fluid inclusions (FI) in ppl. The scale bar in each photo is 100 µm.

5.2.1 Kistufell

Samples were collected from the Kistufell localities outlined in Chapter 2. Kistufell pillow basalts and hyaloclastite contain abundant olivine (Ol) phenocrysts, with some spinel (Sp) and plagioclase (Pl; confirmed from observations when picking olivine phenocrysts from crusted pillow rim glass). The same three phases also make up the microcrystalline (< 100 μ m) groundmass.

Olivine phenocrysts are often unfragmented, euhedral to subhedral in shape, and sometimes contain spinel inclusions. Small clusters of olivine grains can be found, but are generally < 500 μ m in size. Olivine compositions are high in forsterite (X_{Fo} = 0.879–0.897).

Plagioclase phenocrysts are euhedral to subhedral in shape with Ca-rich compositions ($X_{An} = 0.757-0.823$). Spinel occurs as a phenocryst phase among olivine clusters, but has a low modal proportion in the rock.

Olivine-hosted melt inclusions

Breddam (2002) stated that some olivine phenocrysts (< 1%) contained melt and fluid inclusions, but were also devoid of spinel inclusions. In picking olivines for analysis, it was observed that melt inclusions were not as rare as previously suggested, and that some olivines containing melt inclusions also contained spinel inclusions (Figure 5.1a).

Melt inclusions were observed to vary in size $(20-100 \ \mu m \ longest \ axis)$ and have an ovoid shape (Figures 5.1a–b). The majority of the melt inclusions were glassy, free of spinel inclusions and vapour bubbles. A few melt inclusions were seen to contain vapour bubbles and were not analysed, which could introduce a bias on the dataset, however these inclusions were rare and the range of olivine compositions measured by Breddam (2002) has been analysed in this study.

5.2.2 Miðfell

Samples were collected from localities around the Miðfell eruption as detailed in Chapter 2. Pillow basalts from Miðfell are strongly porphyritic with clearly visible olivine, plagioclase and clinopyroxene (Cpx) phenocrysts forming $\sim 50\%$ of the rock (Figure 5.2 left). Olivine phenocrysts are variable in size (0.5–5 mm in length) and have euhedral to subhedral shapes (Figures 5.3a–b). They are seen as individual grains, clustered in groups (not optically continuous) and as polyhedral grains (optically continuous). Petrographic observations indicate that the olivines are highly forsteritic, containing melt inclusions and spinel inclusions.

Plagioclase phenocrysts, 2–4 mm in length, are less abundant than the olivine grains and are mainly found as individual crystals, though some clusters of grains have been observed.



Figure 5.2 (l) Hand specimen of Miðfell pillow basalt interior, hand for scale. (r) QEMSCAN image of a Miðfell gabbroic xenolith (left of image) in association with its carrier assemblage (right of image). Olivine, yellow; plagioclase, blue; clinopyroxene, green. Image is 10 mm across.

They often contain small raggedly-shaped melt inclusions that resemble a sieve texture. The melt inclusions appear in regular groupings suggesting entrapment of melt along a common plane, possibly at the same stage crystal growth. Extinction angles of plagioclase twins suggest Ca-rich compositions ($X_{An} > 0.70$).

Clinopyroxene phenocrysts are often large (1–5 mm in length) and are light green colour in thin section under plane-polarised light (Figure 5.3a). Some clinopyroxene phenocrysts contain melt inclusions, spinel inclusions, and large crystalline embayments. Crystallisation of a paler, lower birefringence pyroxene that is optically continuous to the main phenocryst is observed around some of the larger ones. Observation of $2V_{\gamma}$ and simple twinning in cross-polarised light suggests a Ca-rich clinopyroxene, as expected given the Ca-rich nature of the plagioclase and mafic melt composition. The largest grains are subhedral, while smaller grains (< 2 mm) often show euhedral faces.

Phenocrysts (60% Ol, 30% Cpx and 10% Pl) form 30–40% of the assemblage; the rest is vesicular microcrystalline groundmass (Figure 5.2 right). These phenocryst observations match those of Gurenko and Chaussidon (1995) on Miðfell and Mælifell picrites from the Hengill volcanic system. $X_{Fo}^{Ol} = 0.857-0.915$, high Mg# ^{Cpx} = 0.8–0.9, and high-Ca plagioclase $X_{An}^{Pl} = 0.83-0.88$. Gurenko and Chaussidon (1995) observed that spinels are low in TiO₂, have a Cr# ^{Sp} = 0.31–0.53, and Mg# ^{Sp} = 0.68–0.77.

The crystalline groundmass contains numerous vesicles (< 1 mm), long needles of plagioclase (< 1 mm) showing a clear twin, and an interstitial brown pyroxene that is

dendritic in places. Very small (< 100μ m) colourless olivines grains are seen, distinctive due to their bright second order birefringence. Small spinels and Fe-Ti oxides are also present.

Olivine-hosted melt inclusions

Olivines have been selected from pillow basalt rims where the matrix glass is almost purely basaltic glass (a few microlites of plagioclase can be seen on the SEM). Melt inclusions are typically > 50 μ m in size, glassy, some contain small spinel inclusions, and the rare inclusion has a vapour bubble present (Figures 5.1c–f). Melt inclusions are not aligned in planes, suggesting that they have not been formed by the sealing of cracks in the host olivine. They are ellipsoidal with a 'negative crystal' shape. They contain no obvious inclusions of other mineral grains, apart from the occasional spinel.

Gabbroic xenoliths

Gabbroic xenoliths have also been found in the Miðfell eruption, but within a layer of pillow basalts above those sampled for this study. They consist predominantly of large plagioclase and clinopyroxene crystals (> 3 mm), though some xenoliths do contain olivine (Gurenko and Sobolev, 2006). The xenoliths are made up of glomerocrysts of plagioclase and clinopyroxene that show curved grain boundaries where crystals are fused together (Figures 5.3c-d). Some contain interstitial glass that has de-vitrified to a dendritic brown phase, often seen radiating off the surface of the clinopyroxenes. Needles of very fine plagioclase have grown from the surface of large plagioclase phenocrysts during this process. The corona of phases seen around the clinopyroxene crystals looks like a reaction rim, with arrowheads of the newly crystallised phase from the matrix glass penetrating the surface of the clinopyroxene, making it appear ragged (Figures 5.3e-f). Some plagioclase phenocrysts contain melt inclusions, resembling a sort of sieve texture, which suggests that the plagioclase was partially dissolved (Figure 5.3e–f). This could have been due to decompression, reheating from interaction with a hotter melt, or chemical disequilibrium with the carrier liquid (Nelson and Montana, 1992; Tepley et al., 1999). Whatever the cause, it suggests that the plagioclase experienced conditions that were different to the original equilibrium conditions forming the assemblage. The QEMSCAN image shows melt, now microcrystalline, along grain boundaries in the xenolith to give the rock a slightly disaggregated appearance (Figure 5.2 right).

5.2.3 Whole rock, matrix glass, and melt inclusion geochemistry

Kistufell and Miðfell were both erupted during the last glacial (Weichselian), though it has been suggested that the Kistufell eruption took place near the end of this period (Trønnes,



Figure 5.3 Photomicrographs of Miðfell pillow basalt in (a) plane-polarised light (ppl), (b) cross-polarised light (xpl); and of Miðfell gabbroic xenoliths in (c) and (e) ppl, (d) and (f) xpl. The scale bar is 1 mm.



Figure 5.4 Total alkalis ($Na_2O + K_2O$) against SiO₂ for Miðfell melt inclusions (red circles), matrix glass (red diamonds) and whole rock (red squares); Kistufell melt inclusions (purple circles) and matrix glass (purple diamonds). Indicates the basic nature of Kistufell and Miðfell glasses, and picritic composition of Miðfell whole rock.

1990; Breddam, 2002; Gurenko and Chaussidon, 1995). It is thought that during deglaciation, the pressure decrease from ice sheet thinning led to an increased eruption rate across Iceland, producing more picritic eruptions containing high modal proportions of olivine (Maclennan et al., 2002).

Figure 5.4 shows the total alkali (Na₂O + K₂O) and SiO₂ compositions of Kistufell and Miðfell samples. Melt inclusions and matrix glass compositions are within the basalt field, however Miðfell whole rock data (Gurenko and Chaussidon, 1995) is classified as a picrite due to olivine addition lowering the SiO₂ content of the rock and diluting the alkalis.

5.3 Compositional heterogeneity of olivine-hosted melt inclusions

Olivine-hosted melt inclusions and matrix glasses from Kistufell and Miðfell have been measured for major, trace and volatile element concentrations. SIMS was used to measure volatile and trace element concentrations, while major elements were quantified by EPMA, as detailed in Chapter 2. The Kistufell sample suite consists of 129 melt inclusion analyses from 97 host olivines, and 13 matrix glass measurements. The Miðfell sample suite contains 106

melt inclusion analyses from 69 host olivines, and 17 matrix glass measurements. Thirteen melt inclusions were analysed at Woods Hole Oceanographic Institution (WHOI) in order to reconstruct their original CO_2 concentration by combining vapour bubble CO_2 with CO_2 dissolved in the melt inclusion glass.

Melt inclusion suites have previously been shown to preserve high compositional variability, which has been interpreted to in part reflect the supply of heterogeneous melts into a magmatic system (Maclennan et al., 2003b; Neave et al., 2014; Hartley et al., 2015). As melts migrate through the mantle and crust, several processes reduce their compositional variability. Olivine-hosted melt inclusions have the potential to provide information on the geochemical characteristics of primitive mantle melts, allowing petrologists to see past some of the later crustal melt modification processes to observe signatures of mantle melting and source heterogeneity.

5.3.1 Significance of melt inclusion compositional variability

Before exploring the potential controls on the extent of compositional variability within olivine-hosted melt inclusions, it is important to characterise the variability observed to firstly ascertain that it is not the result of analytical noise. Compositional variability within the Kistufell and Miðfell melt inclusions suites can be shown to be greater than the variability associated with analytical noise by using a χ^2 statistical test (Maclennan et al., 2003b). Tables 5.1 and 5.2 present the results of calculations testing the natural variance of the Kistufell and Miðfell melt inclusion compositions.

The mean composition for each major, trace or volatile element, \bar{x} , has been calculated for each melt inclusion dataset, along with the standard deviation, σ_0 . Analytical variance has been calculated using repeat analyses of matrix glass from each eruption and presented as a standard deviation, σ_r . The true standard deviation of each melt inclusion suite can be calculated from $\sigma_t^2 = \sigma_0^2 - \sigma_r^2$, which represents the amount of variance observed that is not due to analytical imprecision. The signal-to-noise ratio (σ_t/σ_r) quantifies this true variance with respect to the level of variance from analytical noise. The higher the calculated signal-to-noise ratio, the higher the confidence that natural compositonal variability can be resolved from analytical noise (Slater et al., 2001; Maclennan et al., 2003b).

The χ^2 distribution can be used to quantify the confidence in the assertion that observed chemical variability for each element is more significant than analytical noise. σ_r is an estimate of the true analytical error, σ_R , based on a limited number of repeat analyses. The null hypothesis is that all the variability observed is due to analytical error, or rather natural variability cannot be resolved from the uncertainty in each chemical analysis, $\sigma_R \ge \sigma_o$. The alternative hypothesis is therefore $\sigma_R < \sigma_o$, i.e. natural variability can be resolved from
	\bar{x}	$\sigma_{\rm o}$	$\sigma_{\rm r}$	σ_{t}	σ_t/σ_r	P%	V%	Prob.%
SiO ₂	48.2	1.18	0.36	1.12	3.11	0.8	2.3	100
TiO ₂	0.99	0.17	0.02	0.17	8.61	2.1	17.2	100
Al_2O_3	16.9	0.71	0.11	0.70	6.24	0.7	4.2	100
Cr_2O_3	0.09	0.38	0.02	0.38	16.1	48.4	406	100
FeO	9.08	0.34	0.14	0.31	2.26	1.5	3.4	100
MnO	0.16	0.03	0.03	na	na	19.6	na	44
MgO	8.55	1.33	0.08	1.33	16.7	0.8	15.6	100
CaO	13.5	0.65	0.13	0.63	4.77	1.0	4.7	100
Na ₂ O	1.77	0.13	0.04	0.13	2.99	2.4	7.1	100
K_2O	0.07	0.03	0.01	0.03	2.15	17.2	35.9	100
P_2O_5	0.07	0.02	0.01	0.02	1.66	15.1	23.6	100
CO ₂	580	305	170	254	1.49	17.3	43.8	100
H_2O	0.14	0.02	0.03	na	na	19.0	na	19
F	104	14.2	72.2	na	na	149	na	0
Cl	60.2	52.0	27.5	44.1	1.61	68.7	73.3	100
Κ	572	186	47.4	180	3.79	10.1	31.4	100
Ti	5507	965	112	959	8.55	2.4	17.4	100
Sr	117	12.0	4.00	11.3	2.83	4.0	9.7	100
Y	17.1	1.74	0.33	1.70	5.09	2.1	10.0	100
Zr	39.2	6.68	1.31	6.55	5.00	3.4	16.7	100
Nb	2.86	0.71	0.29	0.65	2.27	10.7	22.8	100
Ba	14.0	3.41	1.06	3.24	3.06	9.3	23.2	100
La	2.57	0.54	0.11	0.53	4.79	4.7	20.7	100
Ce	6.54	1.32	0.16	1.31	8.15	2.8	20.0	100
Pr	1.00	0.19	0.09	0.17	1.92	9.7	16.5	100
Nd	5.66	0.99	0.41	0.90	2.18	8.0	15.9	100
Sm	1.88	0.32	0.12	0.30	2.44	7.3	16.0	100
Eu	0.75	0.11	0.06	0.09	1.67	8.2	12.5	100
Gd	2.35	0.40	0.24	0.32	1.35	10.8	13.7	100
Tb	0.44	0.06	0.05	0.04	0.90	12.1	9.5	97
Dy	2.98	0.47	0.29	0.37	1.30	10.7	12.5	100
Но	0.65	0.09	0.05	0.08	1.50	9.0	12.0	100
Er	1.86	0.24	0.19	0.14	0.73	11.4	7.5	91
Tm	0.27	0.05	0.03	0.03	1.02	13.3	12.0	99
Yb	1.91	0.31	0.15	0.27	1.85	8.5	14.2	100
Lu	0.26	0.05	0.03	0.04	1.63	10.9	16.2	100

Table 5.1 Geochemical variability in Kistufell melt-inclusions. Major elements measured by EPMA, trace and volatile elements measured by SIMS.

 \bar{x} , mean of samples; σ_0 , observed standard deviation from the mean of samples; σ_r , repeat standard deviation based on multiple repeat analyses; σ_t , true standard deviation; σ_t/σ_r , signal-to-noise ratio; P%, 1 σ relative precision as a percentage; V%, variability of sample set as a percentage of the sample mean; Prob.%, probability that observed variance is greater than analytical variance is rounded to three decimal places as a percentage.

	ī	6	6	6	ala	D07-	V107-	Drob 0%
	X	00	Or	Ot	O_t/O_r	F %	V 70	F100.70
SiO ₂	48.1	0.74	0.25	0.69	2.72	0.5	1.4	100
TiO ₂	0.87	0.27	0.03	0.27	8.67	2.0	31.0	100
Al_2O_3	14.9	0.56	0.05	0.55	11.7	0.3	3.7	100
Cr_2O_3	0.07	0.05	0.02	0.04	1.73	78.8	53.2	99
FeO	9.44	1.05	0.14	1.04	7.39	1.3	11.0	100
MgO	11.0	0.66	0.07	0.66	9.17	1.0	6.0	100
CaO	13.8	0.73	0.06	0.73	11.6	0.6	5.3	100
MnO	0.17	0.03	0.04	na	na	16.6	na	3
Na ₂ O	1.59	0.19	0.07	0.18	2.50	2.7	11.2	100
K ₂ O	0.06	0.09	0.02	0.08	4.62	14.2	130	100
P_2O_5	0.06	0.07	0.01	0.07	5.49	6.9	104	100
F	0.01	0.01	0.01	na	na	49.1	na	6
Cl	0.005	0.006	0.001	0.005	3.89	18.3	112	100
SO_2	0.16	0.06	0.00	0.06	21.5	1.0	36.3	100
CO ₂	468	304	110	283	2.57	4.1	60.6	100
H_2O	0.07	0.01	0.01	0.00	0.09	20.1	1.9	46
Li	2.53	0.21	0.08	0.19	2.37	4.1	7.6	100
F	85.9	18.0	13.5	11.8	0.87	14.1	13.8	91
Cl	52.9	95.0	17.3	93.4	5.39	77.6	176	100
Κ	440	659	7.99	658	82.4	4.7	150	100
Ti	5205	1542	166	1533	9.26	3.8	29.5	100
Sr	115	46.0	3.47	45.9	13.2	3.3	39.9	100
Y	15.7	2.24	0.41	2.20	5.35	3.2	14.0	100
Zr	29.2	21.7	0.71	21.7	30.5	4.1	74.5	100
Nb	2.15	3.83	0.09	3.83	43.9	11.3	178	100
Ba	13.8	21.4	0.25	21.4	86.4	4.9	154	100
La	1.96	2.82	0.09	2.82	32.1	9.8	144	100
Ce	5.16	6.13	0.22	6.13	27.3	8.8	119	100
Pr	0.81	0.76	0.05	0.76	13.9	10.7	93.3	100
Nd	4.73	3.33	0.34	3.31	9.60	11.3	70.0	100
Sm	1.70	0.61	0.17	0.59	3.46	13.7	34.6	100
Eu	0.72	0.20	0.07	0.18	2.63	12.9	25.6	100
Gd	2.27	0.57	0.30	0.48	1.59	16.7	21.3	100
Tb	0.41	0.10	0.06	0.08	1.36	16.4	18.7	99
Dy	2.87	0.54	0.23	0.49	2.08	10.1	17.0	100
Ho	0.61	0.11	0.07	0.09	1.30	13.0	14.3	99
Er	1.79	0.33	0.17	0.28	1.69	12.6	15.8	100
Tm	0.24	0.05	0.03	0.04	1.33	13.6	15.1	99
Yb	1.68	0.30	0.22	0.20	0.89	15.3	11.8	92
Lu	0.24	0.04	0.04	0.02	0.45	19.6	7.7	65

Table 5.2 Geochemical variability in Miðfell melt-inclusions. Major elements measured by EPMA, trace and volatile elements measured by SIMS.

 \bar{x} , mean of samples; σ_0 , observed standard deviation from the mean of samples; σ_r , repeat standard deviation based on multiple repeat analyses; σ_t , true standard deviation; σ_t/σ_r , signal-to-noise ratio; P%, 1 σ relative precision as a percentage; V%, variability of sample set as a percentage of the sample mean; Prob.%, probability that observed variance is greater than analytical variance is rounded to three decimal places as a percentage.

analytical error. The number of degrees of freedom, f, is one less than the number of repeat analyses used to estimate σ_r . The critical value for the χ^2 distribution, $c = f\sigma_r^2/\sigma_o^2$, can be used to calculate the probability of accepting the null hypothesis, $\chi^2(c, f)$. The probability that the null hypothesis can be rejected, i.e. the confidence level to which natural variability can be resolved from analytical error is $1 - \chi^2(c, f)$.

Table 5.1 shows that the Kistufell major elements measured by EPMA, apart from MnO, show natural variation to a confidence of > 99%. For the volatile and trace elements measured by SIMS, H₂O, F, Tb, Er, and Tm have a natural variation confidence probability of < 99%. Table 5.2 shows that the Miðfell major elements and volatiles measured by EPMA show natural variation to a confidence of > 99%, apart from Cr_2O_3 , MnO, and F. SIMS-measured trace elements and volatiles show a very high confidence in natural variability, apart from H₂O, F, Yb, and Lu. Therefore, it can be concluded that the Kistufell and Miðfell geochemical data used in this chapter, and subsequent chapters, describes variability caused by natural processes rather than analytical imprecision.

5.3.2 Principal component analysis

Principal component analysis (PCA) has been performed on the Kistufell and Miðfell trace and volatile element melt inclusion datasets to help explain the covariance structure of these multivariate data. Each dataset has been reduced to a smaller number of independent variables called principal components that describe the covariance of the dataset in fewer variables than measured.

In practice this statistical procedure involves calculating the $n \times n$ covariance matrix, **C**, where *n* is the number of observed variables, and then the eigenvectors and eigenvalues of this matrix are computed such that:

$$\mathbf{V}^{-1}\mathbf{C}\mathbf{V}=\mathbf{D}$$

where **V** is the $n \times n$ matrix of eigenvectors, $\mathbf{v}_1, ..., \mathbf{v}_n$; and **D** is the $n \times n$ diagonal matrix of eigenvalues, $\lambda_1, ..., \lambda_n$.

The eigenvector with the highest eigenvalue is principal component one (PC1), as this it defines the largest variance of all the principal components. By definition PCs are orthogonal vectors, with the first few describing the majority of the dataset variance. PC loadings on each of the observed variables, i.e. trace and volatile elements, give an indication of how these variables change with respect to each other. Elements that behaved similarly during processes that formed the dataset will have similar loadings for PC1, PC2 and potentially

PC3. The sign of the loading on each element is important, as it gives the sense of correlation between the elements in PC space.

Elements in the $n \times n$ covariance matrix can be ordered such that those with similar PC1 and PC2 weightings are adjacent to each other, giving a visual comparison of elemental behaviour. It is expected that elements with similar compatibility, e.g. the highly incompatible trace elements Ba, Nb, and La, will appear next to each other in an 'angle of eigenvector'ordered covariance matrix.

Principal components and ordered covariance matrices are used to infer the processes that have created the variability preserved within the dataset. For Miðfell and Kistufell, these processes are related to mantle melting, melt transport and crustal processing, which are discussed in the following sections.

5.3.3 Correcting major element oxides for post-entrapment crystallisation

Post-entrapment crystallisation (PEC) has the potential to modify melt inclusion compositions after entrapment, during cooling and decompression of the host olivine. Decreasing temperature causes the crystallisation of olivine on the walls of the melt inclusion, which alters the composition of the melt. MgO, and to a lesser extent FeO, decrease in concentration, while major and trace elements incompatible in olivine become slightly more concentrated in the remaining melt. Although absolute concentrations of incompatible trace elements are changing, the ratios between these elements are much less sensitive to PEC, therefore trace element ratios can be useful for assessing processes that happened prior to PEC. Models have been developed to describe the process of PEC, to better reverse its effect on melt inclusion chemistry and reconstruct the original entrapment compositions; Petrolog is one such model (Danyushevsky and Plechov, 2011).

The composition of the crystallising olivine and the evolution of the melt inclusion is dependent on the partitioning of Fe-Mg between olivine and liquid (Roedder and Weiblen, 1970; Ford et al., 1983; Bédard, 2005; Matzen et al., 2011). Post-entrapment crystallisation causes the Mg# of the melt to decrease, taking it out of equilibrium with its host olivine. Correction of this process can be achieved by adding olivine back into the melt inclusion until it returns to equilibrium with the olivine host. Olivines that are believed to have cooled rapidly can be PEC-corrected by simple olivine addition, as Fe-Mg diffusion is unlikely to have significantly affected the melt inclusion chemistry. However, if cooling is slow enough to allow diffusion after PEC and before quenching, then Fe-Mg diffusion can occur between

the melt inclusion and host olivine to re-equilibrate the two (Danyushevsky et al., 2000). This process can cause Fe-loss from the melt inclusion.

After the crystallisation of a progressively more Fe-rich olivine rim on the wall of the melt inclusion, a chemical gradient in Fe-Mg exists across this boundary. If the olivine remains at a high enough temperature, diffusion of Fe-Mg between the melt inclusion and host olivine can modify the composition of both. Equilibrium between olivine and melt is achieved based on the Fe-Mg olivine-liquid partitioning coefficient, $K_{D \ Fe-Mg}^{Ol-Liq}$, which relates melt Mg# to olivine forsterite content. To achieve equilibrium after PEC, Fe diffuses out of the melt inclusion, increasing the Mg# of the trapped liquid until it reaches equilibrium with the host olivine, reducing the chemical gradient. A simple olivine addition PEC correction on melt inclusions that have undergone diffusive Fe-loss results in the addition of too little olivine back into the melt inclusion, as Fe-Mg diffusion increased the liquid Mg# closer to its equilibrium value based on $K_{D \ Fe-Mg}^{Ol-Liq}$ (Danyushevsky et al., 2002). Therefore, reconstructed melt inclusions would have lower FeO and MgO than the original melt trapped as an inclusion.

Petrolog can be used to correct for Fe-loss as it involves Fe-Mg diffusion as well as olivine addition to reverse PEC in a two-step process (Danyushevsky and Plechov, 2011). Step one allows Fe-Mg diffusion to equilibrate the melt inclusion with the host olivine at constant temperature. Step two increases the temperature of the system, allowing the host olivine to melt and keeping the melt inclusion and host olivine in equilibrium. The process stops when the melt inclusion reaches an initial FeO* content as specified by the user.

This method of post-entrapment crystallisation correction requires the knowledge, or estimation, of the initial melt inclusion FeO content. Danyushevsky et al. (2002) suggests that this can be achieved by using the FeO* against MgO trend for whole rock, groundmass, and glass compositions. This process, however, results in the extrapolation of all melt inclusions back to a single FeO* content. Given the variability seen in other major element oxides and trace element concentrations, as well as the observation that MORB glasses have variable FeO* content that are interpreted as an indicator for melting pressure, it seems unrealistic to give all melt inclusions the same FeO* content (Langmuir et al., 1992).

Figure 5.5 shows the results of correcting for post-entrapment crystallisation on the Miðfell melt inclusion suite using (a) simple olivine addition, (b) Petrolog Fe-Mg reequilibration (Ford et al., 1983, olivine model), and (c) Petrolog Fe-Mg re-equilibration (Danyushevsky et al., 2002, olivine model). All three corrections produce similar distributions of melt inclusion chemistry for Al_2O_3 , and SiO_2 ; two oxides that behaved differently during olivine PEC. Figure 5.5b shows the difference in FeO* reconstruction for the two approaches, with simple olivine addition preserving FeO* variability and the Petrolog methods eliminating it. If the suite of melt inclusions has undergone Fe-loss then these values are FeO* content minima, with some of the variability across the suite due to Fe-Mg re-equilibration.

The Petrolog corrections assume that some of the FeO* exists as Fe₂O₃, based on the oxygen fugacity of the quartz-fayalite-magnetite oxygen buffer (QFM), and $K_D^{Ol-Liq}_{D Fe-Mg}$ is determined by the chosen olivine model. A partition coefficient of $K_D = 0.30 \pm 0.02$ is often assumed for Fe-Mg exchange between olivine and liquid (Roedder and Weiblen, 1970; Ford et al., 1983; Danyushevsky, 2001; Putirka, 2008b). XANES analysis of the melt inclusions suggests that ~ 17% FeO* exists as Fe₂O₃, which gives a partition coefficient comparable to those measured from previous experiments, $K_D = 0.34$ (Matzen et al., 2011). This value is equivalent to a $K_D = 0.28$ if FeO* is assumed to all be FeO, which again is within the accepted range of K_D from MORB observations. The simple olivine addition model uses a $K_D = 0.34$ to match the Petrolog assumption that some of the total iron is Fe³⁺.

Figure 5.5d compares Miðfell PEC-uncorrected melt inclusion compositions to PECcorrected values. All three methods show very similar corrected compositions because they have been calculated using very similar $K_{D \ Fe-Mg}^{Ol-Liq}$ values. The average amount of olivine required for each PEC correction method, with interquartile ranges in parentheses, is: simple olivine addition 4.9% (3.6%–6.1%); Petrolog (Ford et al., 1983, olivine model) 4.7% (1.9%– 9.5%); Petrolog (Danyushevsky et al., 2002, olivine model) 5.3% (2.0%–9.7%). All averages and ranges are similar, and are comparable to those seen in other melt inclusion suites (Neave et al., 2014). Figure 5.5e shows PEC-corrected melt inclusion compositions from the Kistufell eruption, only the simple olivine addition method has been plotted. The average PEC correction required, with interquartile ranges in parentheses, is: simple olivine addition 5.7% (0.0%–14.4%); Petrolog (Ford et al., 1983, olivine model) 6.4% (0.2%–12.1%); Petrolog (Danyushevsky et al., 2002, olivine model) 6.7% (2.3%–12.3%).

The main result from PEC correction is that the MgO content of both Kistufell and Miðfell melt inclusion suites has been corrected back to concentrations higher than that of the matrix glass. This matches the prediction that melt inclusions are trapped prior to the full evolution of the carrier liquid, and therefore record a more primitive melt chemistry. Trace element concentrations have not been corrected for PEC as the calculated PEC corrections are small and the majority of discussion in the following chapters relies upon incompatible element ratios.

5.4 Melt modification in the crust

During melt migration through the crust, pressure and temperature changes help to drive melt evolution processes, which modify the chemistry of melts within the magmatic sys-



Figure 5.5 Major element oxide plots comparing different post-entrapment crystallisation correction models; red, olivine addition; orange, Petrolog Ford et al. (1983) olivine model; blue, Petrolog Danyushevsky (2001) olivine model; Petrolog symbols at 50% opacity. (a) Al₂O₃, (b) FeO, and (c) SiO₂ against MgO for melt inclusions (circles) and matrix glass (diamonds); Rhodes plot (glass Mg# against host olivine forsterite content) for (d) Miðfell, and (e) Kistufell. Percentage PEC lines in grey.

tem. Volatile elements have solubility in silicate melt that is strongly pressure dependent, therefore volatile modification relating to degassing occurs mainly in the mid to shallow crust. Lithophile element modification processes are mainly driven by temperature, with melt mixing, crystallisation and crustal assimilation all linked to temperature changes. This section discusses the crustal processes that could be responsible for compositional heterogeneity of olivine-hosted melt inclusions. It starts with volatile processing in the upper crust, before looking at melt mixing and crustal assimilation, to give an overview of crustal processing before proceeding into the mantle.

5.4.1 Diffusive re-equilibration of water (H₂O)

Measured H_2O concentrations have likely been modified by PEC, however another postentrapment process has the potential to more significantly affect the H_2O content of olivinehosted melt inclusions; diffusive re-equilibration of H_2O .

Several studies have demonstrated that H₂O content and oxygen fugacity of olivinehosted melt inclusions can be independently re-equilibrated with the external liquid by the diffusion of protons (H⁺) and oxygen through the host olivine (Gaetani et al., 2012; Bucholz et al., 2013; Chen et al., 2013, and references therein). External magmas tend to be more degassed than olivine-hosted melt inclusions, therefore the common transfer of H₂O is from the melt inclusion into the external melt. Proton diffusion through olivine can occur via several mechanisms, protons can exchange with (from slowest to fastest): Mg vacancies, Si/Ti vacancies, trivalent cations, electronic defects (polarons), or a combination (Padrón-Navarta et al., 2014; Peslier et al., 2015). Oxygen diffusion requires the creation of isolated metal vacancies, which oxidises Fe^{2+} in olivine octahedral metal sites. Together these mechanisms allow diffusion of H_2O through olivine, which can be independent of Fe^{3+} , and therefore oxygen fugacity (Gaetani et al., 2012; Bucholz et al., 2013). Oxygen fugacity can be re-equilibrated through the creation, diffusion and destruction of isolated point defects associated with the proton diffusion mechanisms detailed above. These processes allow for rapid diffusive re-equilibration of H₂O and oxygen fugacity between melt inclusion and external magma.

However, the reverse process, where melt inclusions become hydrated through diffusion, has been observed in rarer melt inclusion suites (Hartley et al., 2015). Hydration has been determined the consequence of concurrent mixing and crystallisation of compositionally diverse melts. The juxtaposition of melt inclusions low in H₂O next to more hydrous melts during mixing sets up the concentration gradient required to diffuse H₂O into the melt inclusions. Melt inclusions from Laki, Iceland, show evidence of both H⁺-loss and H⁺-gain, suggesting considerable H₂O content heterogeneity during melt mixing (Hartley et al., 2015).



Figure 5.6 Melt inclusion and matrix glass H_2O and CO_2 content for Kistufell, purple; and Miðfell, red. (a) CO_2 against H_2O ; (b) H_2O against Ce, the black line has $H_2O/Ce = 200$; and (c) CO_2 saturation pressure against Ba.

Low H_2O content measured within both the melt inclusions and the matrix glasses of Kistufell and Miðfell are comparable to Laki lava melt inclusions. However, they are 0.5 wt% lower than Laki tephra melt inclusions, and 0.3 wt% lower that Skuggafjöll melt inclusions (Hartley et al., 2015), suggesting that both eruptions have low H_2O content; Miðfell is very low. In both eruptions the H_2O content of the matrix glasses match those of the melt inclusions (Figure 5.6a).

Figure 5.6b shows that in H₂O-Ce space, the matrix glasses, the Kistufell melt inclusions, and the depleted Miðfell melt inclusions are all close to H₂O/Ce = 200, which is in the expected range of undegassed MORBs; 150–280 (Michael, 1995; Saal et al., 2002; Hartley et al., 2015). Only the enriched melt inclusions deviate from this ratio range, being lower in H₂O than would be expected from a ratio of 200. This deviation could suggest one of the following: (i) there was H₂O degassing before enriched melt entrapment, (ii) diffusive re-equilibration of H⁺ through the host olivine, or (iii) H₂O content was initially homogeneous.

5.4.2 Volatile degassing

Decompression during melt transport reduces the solubility of dissolved volatile species within silicate melt (Dixon and Stolper, 1995; Shishkina et al., 2010, 2014). CO_2 is the least soluble of the main volatiles dissolved within silicate melts, making it the most sensitive to pressure changes.

 CO_2 exsolution can occur after melt inclusion entrapment within an olivine as a result of decompression, cooling and PEC; this forms a vapour bubble trapped within the melt inclusion. Olivine acts as a pressure vessel for the melt inclusion, resulting in the melt inclusion experiencing a smaller pressure decrease than the host olivine upon decompression, so the extent of exsolution in the melt inclusion is smaller than for the carrier liquid (Lowenstern, 1995; Aster et al., 2016; Maclennan, 2017). To get back to initial dissolved CO_2 content at the time of entrapment, the CO_2 content of the vapour bubble must be recombined with that of the melt inclusions glass. A method combining X-ray tomography, raman spectroscopy and SIMS has been developed by workers at WHOI (Gaetani, Le Roux and Klein) and is described in Chapter 2.

Olivine decrepitation can cause these vapour bubbles to be lost during olivine decompression and cooling, as the pressure difference between the melt inclusion and carrier liquid is large enough to cause the olivine to fracture (Maclennan, 2017). When this process occurs, information about the initial CO_2 content of the melt inclusion is irretrievably lost.

Volatile degassing can also occur before melt inclusion entrapment, providing that the melt reaches a pressure low enough for dissolved CO_2 to saturate and exsolve. This pressure relationship means that CO_2 concentrations dissolved in silicate melt can be used as a barometer, with the assumption that dissolved CO_2 content represents saturation values (Dixon and Stolper, 1995; Shishkina et al., 2010, 2014).

Figure 5.6c shows calculated values of CO_2 saturation pressure using melt inclusion CO_2 content, including reconstructed values from vapour bubble addition. Data collected for vapour bubble CO_2 reconstruction is detailed in Chapter 2 and was conducted by colleagues at WHOI. Both Miðfell and Kistufell melt inclusions record a maximum pressure estimate of ~ 1.5 kbar (~ 5 km), however Miðfell vapour bubble reconstructions preserve as high as 4.3 kbar (13 km). Assuming these melts were at saturation, then these values present a minimum pressure estimate of entrapment. The pressure ranges are similar to those calculated by igneous barometry techniques presented in Chapter 4.

The vertical trends in Figures 5.6a & c suggest that degassing has occurred to reduce the CO_2 content of some of the melt inclusions, though it is not clear whether this happened before or after entrapment. In the Miðfell melt inclusion suite, vapour bubbles are only seen in the most trace element-enriched melt inclusions, which is likely due to the enriched



Figure 5.7 Halogen (F, Cl) concentrations for Kistufell, purple; and Miðfell, red; melt inclusions, circles; and matrix glass, diamonds. Comparison of SIMS and EPMA measured concentrations of (a) Cl, and (c) F. (b) EPMA Cl against K, Miðfell samples coloured by SIMS session; black line is Cl/K = 0.08, dotted lines are 0.08 ± 0.02 . (d) SIMS F against Nd; black line is F/Nd = 21, dotted lines are 21 ± 10 .

melts having had high initial CO_2 content. Therefore, they saturated at higher pressures than more depleted melts, and were more likely to exsolve CO_2 upon olivine decompression (Maclennan, 2017).

Halogens (F, Cl)

Halogen (F, Cl) measurements were carried out by SIMS and EPMA on Kistufell and Miðfell samples, the results of which are shown in Figure 5.7. For Cl concentrations there are considerable discrepancies between the SIMS and EPMA recovered measurements, but also between SIMS measurements from different sessions. Figure 5.7a shows that Cl content measured by the two techniques do not plot on a 1:1 line. The Miðfell trend above 1:1 is

from the January SIMS session, and below from June. All Kistufell measurements sit below the 1:1, suggesting that if the SIMS calibration can be trusted for these points, then EPMA underestimates Cl concentrations. When plotted on an ordered correlation matrix the two methods of Cl and F measurement do not sit next to each other (Figure 5.8). EPMA Cl and F correlate most with the highly incompatible trace elements, while the SIMS measurements for Cl and F sit among the moderately incompatible elements. The two methods should produce measurements that correlate with the other elements in exactly the same way.

Due to the discrepancy between SIMS sessions, EPMA Cl data has been used to plot Figure 5.7b, which shows the correlation between Cl and K. Miðfell melt inclusions show a Cl/K ratio of ~ 0.08 \pm 0.02. This ratio is comparable to MORB glasses observed by Saal et al. (2002), however they attributed the high ratio to Cl contamination from sea water. Given the subglacial nature of Miðfell, sea water contamination cannot be the reason for a similar Cl/K value. Halldórsson et al. (2016) measured the Cl/K ratio of Miðfell matrix glass and found it to be 0.31, suggesting that EPMA has underestimated Cl for Miðfell and that there is more Cl in the Icelandic source than compared to DMM. Cl isotopes suggest that the variation in δ^{37} Cl for Icelandic basalts can be solely attributed to mantle heterogeneity. Positive correlations between δ^{37} Cl, incompatible trace element ratios, and long-lived radiogenic Pb isotope ratios suggest the incorporation of altered lithosphere into the Icelandic mantle plume source (Halldórsson et al., 2016).

F measurements by SIMS and EPMA are vastly different (Figure 5.7c), with EPMA measurements ranging 0–300 ppm compared to 50–150 ppm for SIMS. The repeat matrix glass measurements are expected to give the same concentrations of F, due to the relatively homogeneous nature of the glass with respect to major and trace elements. This observation is close to being true for SIMS F analyses, but not for EPMA, where the matrix glass appears to vary as much as the melt inclusion populations. Therefore, it is inappropriate to draw conclusions from the EPMA F data.

The correlation between F and Nd is fairly strong in the Kistufell sample suite, with melt inclusions and matrix glasses showing $F/Nd = \sim 21$, which is the expected value for MORB (Workman et al., 2006). Depleted Miðfell melt inclusions and matrix glass plots at a higher F/Nd ratio, due to lower Nd concentrations than observed in Kistufell. However, the more enriched inclusions have a similar average F content as the rest of the suite, resulting in lower F/Nd. Due to the high solubility of F in silicate melt, it seems unlikely that this trend has been caused by degassing (Manning, 1981; Mysen et al., 2004).



Figure 5.8 Correlation matrix for trace element and volatile concentrations of Miðfell melt inclusions; positive correlations in blue, negative in red. Element ordering by angle order of eigenvectors. Subscript E denotes EPMA data, subscript S denotes SIMS data, subscript 2 denotes second isotope data.

5.4.3 Concurrent melt mixing and crystallisation

The supply of basalt to mid-ocean ridges (MORs), and subsequent low pressure evolution moves heat and elements to the surface. Processes controlling melt evolution drive the loss of heat from primitive mantle-derived magmas, reducing the liquid mass of the system and modifying the chemistry of the melt population (Shorttle et al., 2016). Olivine-hosted melt inclusions have the potential to provide a snapshot of a melt population before evolution processes have completely homogenised melt variability into the composition erupted at the surface.

As melts move through the crust, heat loss to the surrounding rock drives crystallisation of mafic phases. If the melt ponds in a magma chamber, convection currents driven by the loss of heat to the surroundings can cause compositionally different melts to mix within the magma body. When these two processes occur on similar time scales, then it can be expected that progressive crystallisation would be coupled to an increased extent of melt mixing. A proxy for melt crystallisation within the erupted products of Iceland is the forsterite content of olivine. Increasingly fayalitic olivines are crystallised as melt temperature and MgO content decrease. Melt mixing is assumed to reduce geochemical heterogeneity, therefore as mixing progresses trace element ratios, such as La/Yb, Ce/Y, and Nb/Zr are predicted to decrease in variability. Trace element ratios are used rather than absolute concentration because they are less sensitive to changes during crystallisation. Mixed trace element ratios will converge on the weighted mean composition of the melts in the system. If an olivine-hosted melt inclusion suite is representative of the population of melts within the magmatic system, then the average composition of the melt inclusions is expected to match that of the carrier liquid composition (Maclennan, 2008a; Hartley et al., 2015).

Figure 5.9 shows trace element ratios of melt inclusions from Kistufell and Miðfell against host olivine forsterite content. Kistufell melt inclusions show very little trace element ratio variability, with average melt inclusion ratios approximately equal to the matrix glass composition. This observation suggests that either melts entering the Kistufell magmatic system have very little primary heterogeneity, or significant mixing has occurred prior to melt entrapment. No melt inclusions were observed in olivine crystals with $X_{Fo} > 0.90$, suggesting that some melt evolution may have occurred prior to entrapment of the observed melt inclusions.

Miðfell melt inclusions show considerable trace element variability, with the mean ratios of the system weighted towards more depleted compositions, lower than that of the Kistufell matrix glass. The mean compositions of the melt inclusion populations have been calculated by $\sum X_i / \sum X_j$, which is equivalent to \bar{X}_i / \bar{X}_j assuming equal volume contribution of each melt in the population. An uncertainty on this average population value can been estimated using the standard error of the mean, σ / \sqrt{n} , where *n* is the number of melt inclusions in the population (Maclennan, 2008a). Melt inclusion averages for Miðfell are: La/Yb = 1.165 ± 0.171, Ce/Y = 0.329 ± 0.040, and Nb/Zr = 0.074 ± 0.014. All three ratios for melt inclusion population are greater than the matrix glass, even when the error on the population estimate is considered, suggesting a slight over-representation of the more enriched compositions or modification of the matrix glass post-entrapment.

The most enriched melt inclusions are distributed across the full range of measured forsterite content, rather than being lost due to mixing at lower forsterite content. This observation could suggest that mixing at the onset of crystallisation was ineffective, allowing melt heterogeneity to be preserved even as the melts cooled and crystallised, perhaps due to reduced convection within the magma body. Alternatively the melt inclusions might not



Figure 5.9 Trace element ratios (a) La/Yb, (b) Ce/Y, and (c) Nb/Zr of melt inclusions plotted against host olivine forsterite content for Kistufell, purple; and Miðfell, red. Matrix glass as solid line, average melt inclusion composition as dashed line, matrix glass equilibrium forsterite composition as vertical lines.

represent one generation of olivine crystals, but rather multiple crystallisation events with the final melt injection carrying the olivines to the surface as a hybrid magma containing more olivine than could have crystallised from the matrix glass. The slight disagreement between matrix La/Yb and Ce/Y with average melt inclusion ratios supported this, suggesting that the distribution of melt inclusions could be a time-integrated signal.

Neither Kistufell nor Miðfell display clear evidence for concurrent melt mixing and crystallisation. In the case of Kistufell it is likely that melt mixing occurred before melt inclusion entrapment, but for Miðfell mixing must have happened later in the magmatic system to produce the homogeneous matrix glass.

5.4.4 Plagioclase assimilation

There is evidence from across Iceland that melts have interacted with crustal material, causing modification of melt compositions (Nicholson et al., 1991; Eason and Sinton, 2009; Brounce

et al., 2012). Globally the interaction between melt and crustal material has been observed on various lenghtscales; from large intrusive bodies assimilating blocks of crustal material, to the expression of simple plagioclase addition within a single olivine-hosted melt inclusion (Bédard, 1993; Hémond et al., 1993; Danyushevsky et al., 2003). It has been suggested that several Icelandic lavas have been modified by crustal assimilation, therefore it is important to constrain the extent of crustal interaction within the Kistufell and Miðfell melt inclusion suites.

The interaction of melts with plagioclase in the crust is thought to be well-characterised due to the distinct trace element chemistry of plagioclase. Unlike the majority of trace elements, Sr is compatible in plagioclase, so interactions with plagioclase-bearing lithologies result in large Sr enrichments. To a much lesser extent, enrichments in Eu are also seen from plagioclase assimilation. These enrichments are quantified as Sr and Eu anomalies, which are ratios of observed Sr and Eu concentrations against expected concentrations, Sr* and Eu*, calculated from elements with similar expected compatibility (Danyushevsky et al., 2003; Workman and Hart, 2005):

$$\left(\frac{Sr}{Sr^*}\right)_n = \frac{[Sr_n]}{([Ce_n]^2 + [Nd_n]^2)^{0.5}} \ , \label{eq:sr_n}$$

where subscript n refers to normalised values.

Figure 5.10a shows the primitive mantle normalised trace element concentrations for Miðfell melt inclusions and matrix glass. Key features shown by this spider diagram are the positive Sr and negative Zr anomalies in the matrix glass and some of the more depleted melt inclusions. Also, twelve of the depleted melt inclusions show Ba > Nb anomalies, as highlighted in Figure 5.10c. Figure 5.10b shows five melt inclusions that have high Sr anomalies (Sr/Sr* > 10), which have been interpreted as having interacted with plagioclase (Danyushevsky et al., 2003). One of the five melt inclusions shows a Eu anomaly, supporting the notion of plagioclase interaction, however the other four and the rest of the melt inclusion suite does not show a Eu anomaly. The five melt inclusions in Figure 5.10b do show a clear depletion in heavy rare Earth elements (HREEs), due to the very low HREE concentrations in plagioclase.

Figure 5.10d shows the primitive mantle normalised trace element content of Kistufell melt inclusions, one of the inclusions shows a Sr anomaly (Figure 5.10e). It appears significant with respect to the Kistufell melt inclusion suite, but in comparison to Miðfell melt inclusions it is a small anomaly.

The five melt inclusions highlighted in Figure 5.10b have been plotted in Figure 5.11 as red stars for comparison with the rest of the Miðfell melt inclusion suite. Figure 5.11a shows primitive mantle normalised ratio of Sr/Ce, with high ratios indicating that the behaviours of



Figure 5.10 Trace element spider diagrams of Miðfell (a) melt inclusions, red, and matrix glass, dark red; (b) plagioclase-contaminated melt inclusions; and (c) melt inclusions with Ba > Nb anomalies. Kistufell (d) melt inclusions, purple, and matrix glass, dark purple; and (e) high Sr anomaly melt inclusions. All elements normalised against primitive mantle composition (Palme and O'Neill, 2014).



Figure 5.11 Major element oxide and trace element ratios for Kistufell (purple) and Miðfell (red) melt inclusions (circles) and matrix glass (diamonds); (a) Al₂O₃/CaO, (b) Al₂O₃/TiO₂, and (c) Sr/Ce (primitive mantle normalised) against MgO; (d) Sr/Sr^{*} (primitive mantle normalised) against La/Yb.

Sr and Ce are quite different, i.e. plagioclase has enriched the Sr concentration of the melt. Figure 5.11b clearly shows the high Sr anomalies of the highlighted melt inclusions, with higher Sr anomalies causing a greater dilution of trace elements and low La/Yb ratios. In both plots plagioclase addition has created a compositional trend subparallel to the main melt inclusion suite, defining the vector for plagioclase addition.

Plagioclase mixing model

Plagioclase mixing with Miðfell melts has been modelled to test the theory that simple plagioclase addition has created trace element variability within the Miðfell melt inclusions. Trace element plagioclase compositions have been taken from the Borgarhraun eruption, Iceland for comparison with plagioclase-silicate melt partition coefficients calculated from experiments and natural samples (Bindeman et al., 1998; Winpenny and Maclennan, 2011; Jenner and O'Neill, 2012). Figure 5.12a compares natural plagioclase compositions to those in equilibrium with Miðfell melts, based on partition coefficients. The natural and fictive plagioclase compositions match reasonably well, with the key anomalies being expressed correctly. The simple mixing model described below uses partition coefficients from Jenner and O'Neill (2012) to give plagioclase compositions for addition to Miðfell melts. These compositions have been used because they contain concentration estimates for more trace elements than the natural plagioclase samples (Winpenny and Maclennan, 2011).

Two Miðfell melt compositions have been selected from the melt inclusion suite to model plagioclase addition. Figures 5.12c & e use a smooth trace element depleted composition,



Figure 5.12 Primitive mantle normalised trace element concentrations for (a) natural plagioclase (Winpenny and Maclennan, 2014), and plagioclase in equilibrium with Miðfell melt based on partition coefficients (Bindeman et al., 1998; Jenner and O'Neill, 2012), (b) mixing of plagioclase from Bindeman et al. (1998) partition coefficients with depleted Miðfell melt inclusions; Jenner and O'Neill (2012) partition coefficients used to mix (c) depleted plagioclase with depleted Miðfell melt, (d) depleted plagioclase with enriched Miðfell melt, (e) enriched plagioclase with depleted Miðfell melt, (f) enriched plagioclase with enriched Miðfell melt.

while Figures 5.12d & f use a smooth trace element enriched composition. The Miðfell matrix glass was not used as it clearly contains a Sr anomaly, which would only been accentuated by mixing with plagioclase.

Coloured lines in Figures 5.12c–f are 10% steps of plagioclase fraction in the meltplagioclase mixture. Trace element concentrations have been calculated based on the bulk addition of plagioclase to the melt using the equation:

$$X_i^{\text{mixed melt}} = FX_i^{\text{pl}} + (1 - F)X_i^{\text{initial melt}}$$

where *F* is the fraction of plagioclase in the mixture, X_i^{pl} is the concentration of element *i* in plagioclase, and $X_i^{\text{initial melt}}$ is the concentration of element *i* in the initial melt.

Mixing of plagioclase with Miðfell melt inclusion compositions produces two key observations: (i) clear Sr anomalies, but no Eu anomaly, and (ii) dilution of trace element concentrations below that of the primitive mantle. Comparison of these plots to Figure 5.10b suggests that the scenario in Figure 5.12e best describes the high Sr anomaly melt inclusions, with an enriched plagioclase composition interacting with a depleted Miðfell melt, although to reproduce the size of the observed Sr anomaly, the plagioclase Sr content needs to be higher. Large macrocrysts of plagioclase exhibiting sieve textures have been observed in Miðfell pillow basalts and xenoliths (Figure 5.3), suggesting that Miðfell plagioclase has melted and contributed to the matrix glass composition. It is not possible to infer from just this observation that the melt inclusions were affected in the same way.

Simple plagioclase addition might be able to describe the trace element patterns for those five melt inclusions, but it cannot describe the trace element patterns for the rest of the melt inclusion suite. An alternative crustal contaminant is gabbroic material, which is addressed in the next section (Sobolev et al., 2000; Gurenko and Sobolev, 2006).

5.4.5 Gabbro assimilation

Fractional crystallisation of mafic melts within the crust will produce gabbroic cumulates, therefore later melts ascending through the magmatic system have the potential to interact with this material (Langmuir, 1989). Miðfell melts have previously been shown to contain higher CaO and Al₂O₃ than MORB, which has been interpreted as the result of clinopyroxene dissolution, although relative enrichments in major elements could be the result of fractional mantle melting (Trønnes, 1990; Langmuir et al., 1992; Sobolev et al., 2000). Gurenko and Sobolev (2006) tested this theory by looking at the interaction between primitive Miðfell melts and gabbroic xenoliths erupted at Miðfell. The trace element anomalies seen in Miðfell melt inclusions are more subtle than those expected by simple plagioclase addition, and

could therefore be caused by interaction with a plagioclase-bearing assemblage instead, i.e. a gabbro.

Gurenko and Sobolev (2006) modelled the interaction of a depleted melt composition, selected from a database of Icelandic melt compositions, with Miðfell gabbroic melt. They concluded that Miðfell melts did interact with Miðfell gabbro, giving rise to a positive Sr anomaly, a negative Zr anomaly, and a Ba > Nb anomaly. Reactive infiltration of the primitive melt with gabbro seemed to fit the melt inclusion suite better than mixing with gabbroic partial melt. Danyushevsky et al. (2003) agreed that Miðfell melts interacted with gabbroic crustal material. They attributed the smaller Sr and Eu anomalies described in high forsterite content olivines in Iceland to assimilation of gabbroic material into hot primitive magma.

Gabbro melting and mixing model

The more extensive suite of melt inclusions measured for this study provides an improvement to previous melt-gabbro interaction modelling with the composition of depleted, uncontaminated melt (Gurenko and Sobolev, 2006). The most depleted melt composition with a smooth trace element pattern, i.e. no anomalies, has been selected for modelling the interaction of Miðfell gabbro with primitive melt.

Mineral trace element concentrations and modal proportions for Miðfell gabbro have been taken from Gurenko and Sobolev (2006). Using batch and fractional melting equations, the compositions of 1%, 5%, 10%, 20%, 30%, and 40% melt fractions have been calculated (Figures 5.13a-b). These melts have been mixed with the assumed-primitive, depleted Miðfell melt composition in 10% fraction steps to give mixed melt compositions. Figure 5.13c shows the mixing between the uncontaminated depleted melt with 10% fractional gabbroic melt. This gabbroic melt composition best recreates the size of the anomalies seen in some of the depleted Miðfell melt inclusions and the matrix glass (Figure 5.10), supporting previous conclusions that gabbroic material has influenced the composition of Miðfell melts (Gurenko and Sobolev, 2006). Up to 10% of this gabbroic melt can be mixed with the depleted composition to recreate anomalies seen in more enriched inclusions. Mixing of gabbroic melt with an enriched melt inclusion shows that up to 50% gabbroic melt can be mixed with this composition and remain similar in trace element composition to the melt inclusions observed. However, the range of variability seen within the Miðfell melt inclusions cannot be explained purely by gabbroic melt addition as gabbroic melts cannot explain: (i) the most trace element enriched melt inclusions, (ii) the smooth trace element patterns of some of the depleted melt inclusions, and (iii) the lack of Eu anomalies in most of the inclusions.

Large positive Sr anomalies are only observed in five of the Miðfell melt inclusions (Figure 5.10b), with 12 more depleted compositions showing negative Zr and Ba > Nb



Figure 5.13 Primitive mantle normalised compositions of melts (blue) formed by the partial (a) batch melting and (b) fractional melting of Miðfell gabbro by 1%, 5%, 10%, 20%, 30% and 40%. Gabbro composition in green. Compositions of melts formed by the mixing of an initially (c) depleted and (d) enriched Miðfell melt compositions (red) and a 10% gabbroic fractional melt (blue), with Miðfell melt inclusion compositions plotted in grey.

anomalies (Figure 5.10c). Removal of these depleted inclusions would suggest that the other depleted melt inclusions are free from crustal assimilation or interaction with crustal material, however there is the potential that the more enriched inclusions contain a 'hidden' crustal component. However, pure gabbroic melts cannot explain the range of melt variability seen in the Miðfell dataset. If gabbroic addition was significant in the dataset, then it could be expected to be shown by principal component analysis as being a key control on dataset variance (see next section). However, as it is not, melt heterogeneity is likely from deeper within the magmatic system, potentially as a result of primary mantle melt variability.

5.5 Melt variability in the mantle

A major control on the extent of compositional variability within the Miðfell melt inclusions is from the mantle. Fractional melting and source heterogeneity are likely the two main factors influencing the diversity of melts produced in the mantle, however these two processes are intimately linked by recycling. Source heterogeneity within the mantle is in part controlled by the recycling of lithospheric material back into the mantle at subduction zones. Recycled oceanic crust was initially formed by the fractional melting of older mantle, concentrating it in incompatible elements, while recycled refractory lithologies, such as harzburgite, represent the depleted residue of a mantle melting process. Therefore, recycled lithologies can provide both an enriched or depleted signature to the more recent mantle assemblages. The formation of source mantle heterogeneities by fractional melting and subsequent subduction recycling makes it difficult to distinguish between an enriched source melted by high melt fraction and a normal to depleted source melted by a low melt fraction. Ultimately, fractional melting is responsible for the level of lithophile enrichment seen within a population of mantle melts.

Once melts have been formed within the mantle they ascend towards the crust, where it is assumed the majority of melt evolution takes place. However, during ascent it is possible that migrating melts interact with each other, mixing en route to the MOHO. This process of melt mixing during transport could be responsible for reducing the range of melt compositions entering the crustal magmatic system.

5.5.1 Major element enrichment

Source heterogeneity and extent of fractional melting have previously been used to explain observations of major element variability from MORB suites (Langmuir et al., 1980, 1992; Shen and Forsyth, 1995; Hirschmann et al., 1999). Figure 5.14 shows some of the major element concentrations and ratios used to describe enrichment. In general, the Miðfell melt inclusions show more compositional variability than those from Kistufell, with the Kistufell melt inclusion population closely matching that of the matrix glass. PEC corrections have slightly modified the absolute concentrations, however ratios of major elements incompatible in olivine remain fairly invariant to crystallisation. Assuming that melt heterogeneity was originally present within the Kistufell magmatic system, then the melt inclusions suggest considerable melt mixing has occurred prior to entrapment to destroy this compositional variability. Comparing these eruption to other melt inclusion suites from Iceland, suggests that Miðfell is one of the most compositionally variable datasets spanning nearly the full range of major element compositions recorded by other suites (Figure 5.14). Despite having lower variability, Kistufell melt inclusions are on average more enriched than Miðfell; this is also reflected in the composition of the matrix glass. Compared to the wider Iceland dataset, Miðfell matrix glass and the majority of the melt inclusions show more depleted compositions.

Figure 5.14a shows that Miðfell matrix glass and several of the melt inclusions have higher CaO content than equivalent melts from Kistufell; they also show a higher CaO/Al₂O₃ ratio



Figure 5.14 Major element oxides and ratios against MgO content for Kistufell, purple; and Miðfell, red; melt inclusions, circles; and matrix glass, diamonds; compared against Icelandic melt inclusions, blue. (a) CaO, (b) FeO, (c) CaO/Al₂O₃, and (d) K₂O/TiO₂.

(Figure 5.14c). FeO content are similar for both eruptions, with both average compositions sitting in the centre of the Icelandic dataset, but Miðfell melt inclusions showing considerable variability (Figure 5.14b). Miðfell melt inclusions span a large range of K_2O/TiO_2 ratios, showing a similar level of variability as wider Iceland melt inclusions (Figure 5.14d). The Miðfell matrix glass has a lower K_2O/TiO_2 ratio than Kistufell matrix glass, suggesting a greater contribution from high K_2O/TiO_2 melts in Kistufell compared to Miðfell.

The high MgO content of Kistufell and Miðfell melt inclusions suggests that their compositions are close to primary mantle melts, and if Fe-loss has affected these inclusions during PEC then their equilibrium MgO content would have been higher (Klein and Langmuir, 1987; Danyushevsky et al., 2002). The high forsterite content of host olivines, up to $X_{Fo} = 0.91$, suggests that these melts are the most primitive that could be trapped at Miðfell. Some melt evolution may have occurred before entrapment, however it is difficult to quantify due to the uncertainty in the exact Mg# of primary mantle melts (Maclennan et al., 2003a). It has been suggested that primary mantle melts have Mg# > 70, a condition met by melt inclusions within the most forsteritic Miðfell olivines (Green, 1970; Gurenko and Chaussidon, 1995).

The presence of clinopyroxene within the melting region has been shown by experiments to influence melt CaO content and CaO/Al₂O₃ ratios (Hirschmann et al., 1999). Higher CaO concentrations and CaO/Al₂O₃ ratios are expected from more depleted sources, providing that clinopyroxene is still present within the source residue. This is due to the compatible nature of Ca within clinopyroxene, resulting in equilibrium melts with low CaO content. Miðfell melt inclusion CaO content suggests that the mean source of melts beneath Miðfell is more depleted than Kistufell, either from low clinopyroxene proportions at the base of the melting region or due to a higher melt fraction of weighted melts (Figure 5.14a & c). Low CaO/Al₂O₃ can be produced by low melt fractions, or high melt fractions where clinopyroxene is exhausted in the solid residue, therefore there is not complete certainty that different extents of melting have not caused the variation in CaO and CaO/Al₂O₃. If the Kistufell source was exhausted in clinopyroxene and the Miðfell source was not, then the difference in CaO/Al₂O₃ ratios could indicate that Kistufell has undergone a higher fraction of melting to cause a significant dilution of CaO.

High K₂O/TiO₂ ratios have previously been interpreted as an enriched mantle source signature, which assumes that both elements are similarly incompatible during mantle melting (Langmuir et al., 1992; Shen and Forsyth, 1995; Gurenko and Chaussidon, 1995). However, peridotite melting experiments suggest that at low melt fractions the two elements are significantly different in compatibility ($D_K > D_{Ti}$), meaning that high K₂O/TiO₂ ratios are the result of low melt fractions. This difference in compatibility is due to increased SiO₂ content of near solidus liquids at low to moderate pressures, and due to more Na- and Al-rich clinopyroxene compositions stabilising higher valency ions at high pressures (Hirschmann et al., 1999). It has been suggested that the extent of melting has a stronger influence on melt K₂O/TiO₂ ratios then source enrichment. A correlation matrix from Miðfell melt inclusion data supports the difference in compatibility between K and Ti, as K correlates best with the most incompatible elements, while Ti clusters with the moderately incompatible elements. Miðfell melt inclusions do not show two clear populations based on K₂O/TiO₂ as suggested by others (Figure 5.14d), instead the continuous distribution of melt inclusion compositions implies that the simplified view of enriched and depleted populations is not quite correct (Gurenko and Chaussidon, 1995).



Figure 5.15 Nb/Zr ratio against MgO content for (a) Kistufell, purple; Miðfell, red; melt inclusions, circles; matrix glass, diamonds; and Icelandic melt inclusions, blue; (b) Nb/Zr ratio against MgO content for Miðfell melt inclusions and matrix glass (black outlines) and Icelandic melt inclusions coloured by K_2O/TiO_2 content.

FeO content of basalts that have undergone olivine crystallisation are dependent on melting pressure, and the extent of melting, though source composition must also be taken into account (Langmuir et al., 1980; Shen and Forsyth, 1995). Overall, the major element compositions of Miðfell melt inclusions, and to a lesser extent Kistufell, can be used to infer mantle source heterogeneity. However, trade-offs between extent of melting and source enrichment mean that conclusions are by no means certain (Hirschmann et al., 1999). Other geochemical indicators of source heterogeneity and extent of melting are required to resolve this uncertainty.

5.5.2 Trace element variability

Trace element ratios involving two incompatible elements can be less sensitive to fractionation processes than absolute concentrations, providing the fraction of melting is neither a lot smaller than the element partition coefficients ($F << D_i, D_j$) nor the extent of melting is smaller than one but larger than the other partition coefficient ($D_i < F < D_j$). Nb/Zr has been highlighted as one such trace element ratio useful for discerning source enrichment. This assumption is based on observations of covariance between trace elements and isotopic enrichment in Icelandic basalts, and coupled to the finding that Nb/Zr is resistant to change through fractional crystallisation (Shorttle and Maclennan, 2011).

Figure 5.15a presents the Nb/Zr ratios from Kistufell and Miðfell, along with wider Icelandic melt inclusions suites. Kistufell shows very little variability, while Miðfell nearly spans the whole Icelandic melt inclusions range. The Miðfell melt inclusion suite is compared



Figure 5.16 Principal component loadings on trace elements and volatiles from (a) Miðfell melt inclusions for the first three principal components. PC1 in green, $\sigma_1 = 4.07$; PC2 in blue, $\sigma_2 = 1.76$, PC3 in red, $\sigma_3 = 1.19$; and (b) Kistufell melt inclusions; $\sigma_1 = 4.11$; $\sigma_2 = 1.46$; $\sigma_3 = 0.97$

to the Icelandic melt inclusion dataset in Figure 5.15b, with Nb/Zr plotted against MgO content and coloured by K_2O/TiO_2 . Both the Miðfell and Icelandic melt inclusions show higher K_2O/TiO_2 for more enriched Nb/Zr compositions. This supports the inference by Shen and Forsyth (1995) that K_2O/TiO_2 is a measure of enrichment, rather than extent of melting. The conclusion by Hirschmann et al. (1999) may be valid for lower melt fractions, but in the case of Miðfell, melt fractions might be high enough to negate this effect.

The use of major element compositions to define enrichment and suggest source heterogeneity must be cautioned, particularly when using published melt inclusion datasets. It is likely that some of the melt inclusions from the Icelandic dataset have been strongly influenced by natural PEC and diffusion, as well as experimental homogenisation and different PEC corrections. The variability introduced by these modification processes and correction procedures can obscure the true relationship between major element chemistry and source characteristics. Trace elements and ratios can be more reliable indicators of enrichment.

Source enrichment can be quantified by calculating the pyroxenite melt contribution to an eruption from trace element concentrations (Shorttle et al., 2014). Enriched and depleted trace element end-members have previously been defined for Iceland's rift zones, which are used with observed trace element concentrations to calculate a pyroxenite melt component (Shorttle et al., 2014, Equation 4). The Nb and Zr content of Kistufell and Miðfell matrix glasses have been used to give the following pyroxenite melt proportion estimates: Kistufell 14.6% (NVZ end-members); Miðfell 1.8% (WVZ end-members). The higher average Nb/Zr content of Kistufell compared to Miðfell translates as $\sim 12\%$ more pyroxenite melt, however given the higher variability in Miðfell melt inclusions it can be inferred that the melt inclusions with higher Nb/Zr contain a more significant proportion of pyroxenite melt.



Figure 5.17 Modelled trace element content, normalised to initial solid composition, for garnet peridotite melting in fractional steps of 1% using the melting equations of Walter (1998). Gt-present reaction: 7 Ol + 68 Cpx + 25 Gt = 84 Liq + 16 Opx, Gt-absent reaction: 6 Ol + 94 Cpx = 64 Liq + 36 Opx. Partition coefficients from McKenzie and O'Nions (1991).

Principal component analysis

Principal component analysis on the Kistufell and Miðfell datasets resolves the covariance of the data into orthogonal vectors, which reflect different processes or combination of processes acting in the magmatic systems. Figure 5.16 shows the principal component (PC) loadings on each of the trace and volatile elements measured in the Miðfell and Kistufell melt inclusions. The first PC for the Miðfell dataset is positive for all elements, showing that all these components are varying along PC1 together in the same sense (Figure 5.16a). The processes responsible for this are likely fractionation processes, with all incompatible elements increasing or decreasing together. The volatile elements appear to be decoupled from PC1, showing very little weighting for CO_2 and H_2O , but more for F and C1. This likely reflects the fact that F and C1 are relatively soluble, behaving like a trace element, while CO_2 has undergone degassing and H_2O has been partially reset by diffusive re-equilibration, decoupling them from trace element fractionation.

PC2 shows a strong positive Sr peak, and negative Y value similar to the heavy rare Earth elements (HREEs). The positive PC2 values for light REEs (LREEs) transition to negative values going across into HREEs. These observations suggest that PC2 reflects the fractionation of light and heavy REEs during mantle melting in the presence of garnet. Figure 5.17 shows the results of a simple fractional melting model for a garnet peridotite composition. While garnet is present within the solid assemblage, light and heavy REEs are strongly fractionated. This is due to the HREEs being more compatible within garnet than the LREEs. Once the garnet is exhausted, or converted to spinel at lower pressures, the trace element pattern flattens and all REEs are expected to show the same polarity of PC related



Figure 5.18 Sr-Nd and Pb isotopes for Kistufell, purple; and Miðfell, red, matrix glass compared to other Icelandic glasses; WVZ and RP, pink; SNVZ, green; NNVZ, blue. Known Miðfell matrix glass ²⁰⁶Pb/²⁰⁴Pb and Nb/Zr as dark red lines, ratios estimated from WVZ trends are dotted and dashed. Data from Brandon et al. (2007); Debaille et al. (2009); Peate et al. (2010); Halldórsson et al. (2016).

to melting. The strong Sr signal in PC2 suggests that interaction with a plagioclase-bearing lithology is incorporated into this principal component. PC3 appears more of a random signature. Kistufell principal components show a very similar structure to the Miðfell values, with PC1 showing some sort of bulk fractionation process, PC2 reflecting garnet zone melting but without an added Sr signal, and PC3 has a large Gd signal but fairly invariant with the other elements (Figure 5.16b).

5.5.3 Isotope signatures

Variability of radiogenic isotopes is caused by processes that fractionate parent and daughter isotopes, and then time to allow isotopic evolution of the new reservoirs. In the case of

OIB and MORB, these reservoirs are mantle components with different isotopic signatures (Stracke et al., 2005; Brandon et al., 2007). Several radiogenic decay systems are used to define different isotopic reservoirs within the mantle, namely Sr-Nd, Hf-Nb, U-Pb. Pb isotopes provide a way of discriminating mantle components based on ratios of ²⁰⁴Pb, ²⁰⁶Pb, ²⁰⁷Pb, and ²⁰⁸Pb, with end-members such as HIMU, FOZO, EMI, and EMII (Stracke et al., 2005). Isotopic ratios are unaffected by mantle melting and melt crystallisation, therefore providing that the isotopic signature has not been modified by crustal assimilation, then isotopic ratios relate directly to source mantle components.

Kistufell and Miðfell melt inclusions have not been measured for Sr, Nd, or Pb isotopes, however several studies have measured their matrix glasses to characterise isotopic variability in Iceland (Brandon et al., 2007; Debaille et al., 2009; Peate et al., 2010; Halldórsson et al., 2016). Figure 5.18 shows the Sr-Nd and Pb isotopic nature of these two eruptions. For Sr-Nd measurements it can be seen that the two eruptions have very similar compositions, in the middle of the Icelandic range. Pb isotopes are similarly in the middle of the Icelandic range for Kistufell. The ²⁰⁶Pb/²⁰⁴Pb ratio of Miðfell glass has been measured, along with the Nb/Zr ratio (Halldórsson et al., 2016). The correlation between Pb isotopes and Nb/Zr for WVZ glasses can be used to estimate the other Pb isotope ratios for Miðfell, placing it a slightly higher ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb, and ²⁰⁸Pb/²⁰⁴Pb ratios than Kistufell.

5.6 Compositional enrichment and variability

The Miðfell and Kistufell eruptions provide two melt inclusion datasets for comparison with Icelandic melt inclusions to help explore the controls on trace element variability within olivine-hosted melt inclusions. Observations of trace element concentrations from these suites show that in general, lower melt variability is preserved within low forsterite olivines (Figure 5.19a). It is thought that melt mixing within the crust is able to homogenise melt compositions, with melt inclusions trapped within lower forsterite olivine representing the compositions of more evolved melt; long crustal residence times allow for more melt mixing (Maclennan et al., 2003a; Maclennan, 2008a; Neave et al., 2014). As a result of melt mixing the extent of melt variability is reduced, with compositions converging on the mean composition of the system.

Melt inclusions from Borgarhraun, NVZ, show the clearest evidence for concurrent melt mixing and crystallisation, as trace element ratio variability is seen to decrease as forsterite content decreases, i.e. as crystallisation progresses (Figure 5.19c). Borgarhraun is essentially unique among other Icelandic melt inclusion suites, as the other eruptions shown in Figure 5.19 do not show this clear evolution on a individual eruption basis (Maclennan,



Figure 5.19 La/Yb content of Icelandic olivine-hosted melt inclusions, normalised to the mean La/Yb ratio of each melt inclusion suite and plotted against host olivine forsterite content; (a) all Iceland; (b) WVZ; (c) NVZ; (d) EVZ.

2008a). This relationship is either because the melts have already mixed before olivine entrapment, with crystallised olivine showing low forsterite content (Figure 5.19d), or olivine compositions do not extend down to low enough forsterite content to show complete mixing of the melt population (Figure 5.19b).

Comparing the La/Yb ratios (normalised by the mean La/Yb of each melt inclusion suite) of melt inclusions trapped within olivines with $X_{Fo} > 0.85$ shows that some eruptions are more variable than others (Figure 5.19). Miðfell has the largest normalised variability of all the Icelandic melt inclusion suites, showing more variability than Borgarhraun. At the other end of the spectrum, Kistufell shows very little compositional heterogeneity.

Eruption	νz	\bar{x}_{La} (ppm)	σ_{La} (%)	$C_0 (\text{ppm})$	F	D	σ_{model} (%)	Μ	IMu	Reference
Fimmvörðuháls	EVZ	17.5	19.7	0.65	0.04	0.012	124	0.975	14	Moune et al. (2012)
Laki	EVZ	11.7	28.8	0.65	0.06	0.012	152	0.964	136	Hartley et al. (2014)
Laki	EVZ	11.7	27.8	0.65	0.06	0.012	152	0.966	34	Neave et al. (2013)
Saksunarvatn	EVZ	4.4	60.8	0.65	0.15	0.012	248	0.940	117	Neave et al. (2015)
Skuggafjöll	EVZ	3.6	28.0	0.65	0.18	0.012	275	0.990	112	Neave et al. (2014)
Surtsey	EVZ	14.1	30.4	0.65	0.05	0.012	139	0.952	63	Schipper et al. (2016)
Bóndhólshraun	NVZ	2.0	27.9	0.65	0.32	0.012	364	0.994	11	Slater et al. (2001)
Borgarhraun	ZVN	1.4	49.8	0.65	0.47	0.012	443	0.987	36	Maclennan et al. (2003b)
Borgarhraun	NVZ	1.3	70.3	0.65	0.50	0.012	457	0.976	29	Slater et al. (2001)
Gæsafjöll	NVZ	7.7	15.8	0.65	0.08	0.012	188	0.993	67	Maclennan (2008a)
Kistufell	NVZ	2.8	23.7	0.65	0.24	0.012	314	0.994	122	This study
Krafia	ZVN	5.9	29.4	0.65	0.11	0.012	215	0.981	8	Maclennan (2008a)
Langavíti	NVZ	1.2	60.1	0.65	0.54	0.012	473	0.984	12	Slater et al. (2001)
beistareykir pic.	NVZ	1.6	70.2	0.65	0.41	0.012	411	0.971	7	Slater et al. (2001)
Háleyjabunga	ZVW	0.7	94.9	0.65	0.93	0.012	623	0.977	6	Maclennan (2008b)
Lagafell	ZVW	3.1	16.9	0.65	0.21	0.012	294	0.997	б	Gurenko and Chaussidon (1995)
Mælifell	ZVW	3.7	139.4	0.65	0.18	0.012	271	0.735	6	Gurenko and Chaussidon (1995)
Miðfell	ZVW	2.5	117.6	0.65	0.26	0.012	329	0.872	16	Gurenko and Chaussidon (1995)
Miðfell	ZVW	2.0	144.4	0.65	0.33	0.012	372	0.849	98	This study
Stanafell	WVZ	7.6	34.3	0.65	0.09	0.012	189	0.967	6	Maclennan (2008h)

Table 5.3 Observed and calculated La variability using the equations of Rudge et al. (2013) for Icelandic melt inclusion suites. These statistics can be used to provide an estimate of melt mixing prior to melt inclusion entrapment.

modelled primary mantle melt variability, as a percentage of the mean, $\sigma_{model}^2 = (C_0)^2/2FD$; *M* is the mixing parameter, where $M = 1 - (\sigma_{La}^2/\sigma_{model}^2)$, M = 1 is completely homogeneous, and M = 0 is a population of fully unmixed incremental melts; n_{MI} is the number of melt inclusion measurements in the dataset source composition (McDonough and Sun, 1995); melt fractions, F, have been estimated from $F \approx C_0/C_1$; D is the peridotite-melt partition coefficient for La; \bar{x}_{La} is the mean La composition of the melt inclusion suite; σ_{La} is the standard deviation of La compositions as a percentage of the mean; C_o is the assumed initial (Jennings et al., 2017).



Figure 5.20 Relative standard deviation against mean for (a) La, and (b) Nb concentrations measured from Icelandic melt inclusion suites. Relative standard deviation error bars estimated from the χ^2 distribution. Grey points are expected mantle melt variability using the model of Rudge et al. (2013). Blue line is predicted variability using the model presented in this study. Suites are filtered such that numbers of melt inclusions > 30 (c) La, (d) Nb.

The previous sections explored the processes responsible for the lithophile and volatile element variability seen within the Miðfell and Kistufell melt inclusion suites. Some of the Miðfell melt inclusions have been modified by interaction with gabbroic material, giving them positive Sr, negative Zr, and Ba > Nb anomalies. The majority of these inclusions have depleted trace element concentrations, however some also show smooth trace element profiles suggesting no gabbro assimilation. Trace element patterns suggest that the full range of trace element variability is not the result of crustal processing, but rather inherited

from the mantle, through fractional melting of mantle material. In comparison, the Kistufell melt inclusion suite shows very little compositional variability, which raises the question of why Kistufell would show minimal trace element variation and yet Miðfell produces such a diverse array of melt compositions. Melt heterogeneity originates in the formation of primary mantle melts, so investigation of this process can provide an indication of the possible melt variability generated by melting.

The variability of primary mantle melts can be estimated using mantle melting models (Rudge et al., 2013; Jennings et al., 2017). Table 5.3 shows compositional variability statistics for several melt inclusion suites from across Iceland, using the equations of Rudge et al. (2013) as implemented by Jennings et al. (2017). These calculations assume an initial source La concentration $C_0^{La} = 0.65$, an average melt fraction $F \approx C_0/C_1$, and a partition coefficient $D_{La} = 0.012$, to give an estimate of compositional variability for a population of primary mantle melts (McDonough and Sun, 1995). The extent of mixing, M, required to go from expected primary mantle melt variability to observed variability is calculated using $M = 1 - (\sigma_{La}^2/\sigma_{model}^2)$, with melts assumed to be produced by perfectly fractional melting. As suggested by the observed variability in Figure 5.19, Miðfell has experienced the lowest fraction of mixing, while Kistufell has seen the highest, hence their contrasting extents of variability.

Figure 5.20 compares the Rudge et al. (2013); Jennings et al. (2017) model predictions for primary mantle melt variability to observed La and Nb variability. All the melt inclusion suites show lower variability than is predicted by fractional primary mantle melts. The melt inclusion suite observations seem to show a trend of greater trace element variability, relative to the mean of the melt population, at lower mean trace element concentrations. The form of the Rudge et al. (2013); Jennings et al. (2017) expected variability loosely matches the shape of observed data, but variability has been lost likely due to mixing. Uncertainties of variance are based on the 95% confidence limits of a χ^2 distribution.

The blue line on Figure 5.20 is another simple model for calculating the expected primary mantle melt average composition and variability for different extents of melting. It has be derived from the fractional melting equation:

$$C_i = \frac{C_0}{D} (1 - F)^{\frac{1}{D} - 1},$$

where C_0 is the initial solid concentration, D is the peridotite-silicate melt partitioning coefficient, and F is the fraction of melting. From this the mean concentration of a population of melts, \bar{C}_i , can be calculated by integrating over the melt fraction range and dividing by the cumulate melt fraction:

$$\bar{C}_i = \frac{C_0}{F_{max} - F_{min}} \int_{F_{min}}^{F_{max}} C_i(F) dF,$$

when $F_{min} = 0$, then:

$$\bar{C}_i = \frac{C_o}{F_{max}} \left[1 - (1 - F_{max})^{\frac{1}{D}} \right].$$

The variance of a function can be calculated by:

$$(\boldsymbol{\sigma}_i)^2 = E\left((C_i)^2\right) - (\bar{C}_i)^2,$$

where $E((C_i)^2)$ is the expected value of the melting function squared for the melt fraction range of interest. When $F_{min} = 0$, then the variance can be calculated from:

$$(\boldsymbol{\sigma}_i)^2 = \frac{(C_0)^2}{F_{max}} \left[\frac{1}{D(2-D)} \left((1-F_{max})^{\frac{2}{D}-1} - 1 \right) \right] - \frac{(C_0)^2}{(F_{max})^2} \left[1 - (1-F_{max})^{\frac{1}{D}} \right]^2.$$

Figure 5.21 shows the predicted variability and enrichment of a population of primary mantle melts for different extents of melting, i.e. from $F_{min} = 0$ to F_{max} , where $0.01 \le F_{max} \le 0.50$. The effect of changing initial source composition, C_0 , can be seen in Figure 5.21a, which shows greater variability at equivalent levels of enrichment for higher initial source concentrations. Figure 5.21b shows that increasing the incompatibility of the trace element, i.e. smaller *D*, increases the expected relative variability.

Melt mixing is required to reduce primary mantle melt variability, therefore this must be included in the simple melt model to bring predicted variability closer to the values observed. Figures 5.21c–d test two melt mixing scenarios, which have been modelled using a Monte Carlo simulation of 50000 melts created by melt fractions selected randomly from a specified melting range and using the fractional melting equation.

In Figure 5.21c the highest X fraction of melts are mixed and homogenised, simulating the mixing of the first, and most enriched melts, to form in the mantle. This model does not fit the observed data, with variability decreasing more for the depleted compositions than the enriched. A different mixing scenario is presented in Figure 5.21d, where the melt mixing parameter (X) is a function of melt fraction (F), X = f(F) = (1 - aF). In this scenario, the extent of melt mixing decreases as melt fraction increases. The reason for this is that as melt fraction increases, melt volume increases, and it is assumed that larger melt volumes are more difficult to homogenise as mixing over longer lengthscales is less efficient. Therefore, heterogeneity within a larger melt volume is more easily preserved than for a



Figure 5.21 Simple primary mantle melt variability model derived from the fractional melting equation. Lines show model results when changing (a) C_o ; (b) D_i ; (c) fraction of highest La melts that are homogenised; (d) *a* in the mixing coefficient equation, X = (1 - aF). Observed La melt inclusion variability plotted in grey.

small melt volume. This is not true once the melt enters a magma chamber and starts mixing by convection currents, so this melt mixing model would only apply to melt in transport from its mantle source. This simple model matches the observed data more successfully than the model in Figure 5.21c, but it does not fit the data perfectly and could not be used as a predictive tool. A more rigorous mathematical basis is required to effectively describe the mixing of primary mantle melts during melt transport in the mantle.

The model presented above and shown in Figure 5.21d suggests that a certain extent of melt mixing in the mantle or lower crust is required before melt inclusion entrapment to reduce the variability of a primary mantle melt population. Eruptions that have trace element enriched compositions are inferred to have undergone a lower extent of fractional melting, suggesting lower melt volumes. If melt mixing in the mantle becomes less efficient as melt volume increases, then enriched melts should be more thoroughly mixed than depleted average melt populations (Maclennan, 2008b; Stracke and Bourdon, 2009; Rudge et al., 2013).
5.7 Conclusions

The observed geochemical variability within the Kistufell and Miðfell melt inclusion suites is believed to be greater than variability produced by analytical error for the majority of major, trace and volatile elements to a high degree of confidence (> 99%). This chapter assessed the crustal and mantle processes that could have been responsible for creating and modifying the geochemical variability preserved in these melt inclusion suites.

Melt inclusion H_2O content has likely been reset by diffusive re-equilibration through the olivine host, which has resulted in low H_2O concentrations and low variability across both melt inclusion suites. The low solubility of CO_2 in silicate melt has resulted in many of the melt inclusions experiencing degassing. This process has either occurred prior to entrapment, which is more likely for the trace element enriched inclusions as their parent melt would have saturated in CO_2 at the highest pressure. Additionally, degassing may have occurred post-entrapment during decompression, cooling and crystallisation, which in some cases led to the formation of a vapour bubble and its loss due to olivine decrepitation. Some of the melt inclusions have low trace element content and CO_2 concentrations lower than the degassed maximum seen at higher trace element content, suggesting that these melt inclusions are undegassed; therefore preserving their primary carbon content.

Both melt inclusion suites do not show trace element ratios that vary systematically with host olivine forsterite content, which suggests that concurrent melt mixing and olivine crystallisation was not responsible for the formation and trace variability of the two melt inclusion suites. Kistufell melt inclusions have compositions very similar to the matrix glass, suggesting that the majority of any primary melt heterogeneity has been eliminated by mixing prior to entrapment. Miðfell melt inclusions show variability at all olivine forsterite values, suggesting that mixing had been inefficient up until the point of melt inclusion entrapment in the magmatic system. These two eruptions add to the Icelandic melt inclusion dataset, which overall shows lower compositional variability at lower olivine forsterite content.

Melts ascending through the crust have the opportunity to interact with crustal lithologies, such as gabbro, which imprints a geochemical signature onto their trace element patterns. Five melt inclusions from Miðfell show strong Sr anomalies and REE depletion, suggesting that they have been modified by simple plagioclase addition. Twelve Miðfell melt inclusions show more subtle positive Sr anomalies, negative Zr anomalies, and Ba > Nb anomalies. These inclusions have interacted with gabbroic material, as observed from xenoliths in the eruption, to create this signal, which is also observed within the matrix glass. The full range of lithophile element variability is preserved within melt inclusions that have smooth trace element patterns, i.e. melt inclusions that have not been modified by interaction with gabbro. This observation suggests that melt heterogeneity is coming from the mantle.

Source heterogeneity and extent of fractional melting are intricately linked in their control over lithophile enrichment, as enriched lithologies melted by high fractions will produce similar composition melts to a more depleted lithology melted by lower extents. Major and trace element concentrations and ratios can be used to infer source enrichment, however the trade-off with fractional melting cannot be ignored. Miðfell and Kistufell major element content show some variability and suggest that on average Kistufell is a more enriched eruption, however firm conclusions on melt heterogeneity cannot be drawn from these observations alone.

Observations of Nb/Zr ratios from Icelandic melts suggest that this ratio is invariant to fractional melting and crystallisation, therefore the high Nb/Zr ratios seen in Miðfell melt inclusions reflect source enrichment, but the higher average Nb/Zr ratio for Kistufell suggests that its source region is more enriched than for Miðfell. Nb/Zr is found to correlate with CaO and K_2O/TiO_2 melt inclusion content, and to a lesser extent with CaO/Al₂O₃ and FeO, which supports the notion that major element variability is in part the result of source heterogeneity. The matrix glass Nb/Zr ratios of Kistufell and Miðfell can be converted into an estimate of pyroxenite melt contribution to the eruption. Calculations suggest that pyroxenite melt contributes to 14.6% of the Kistufell melt and to 1.8% of Miðfell, further supporting the idea that Kistufell is a more enriched eruption than Miðfell.

Comparison of several Icelandic melt inclusion datasets suggests that populations with a more depleted average trace element composition preserve a higher degree of relative variability compared to more enriched populations. The variability observed within these melt inclusions is much lower than predicted by a simple fractional melting model for primary mantle melts. Modification of this model, which is derived from the fractional melting equation, by incorporation of a melt mixing component improves the match between observed and predicted variability, however the mathematical basis for mixing is not particularly rigorous. The model does suggest that reduced variability in enriched melts is due to efficient melt mixing at low melt fractions within the mantle. Once depleted melt compositions are produced at higher melt fractions and volumes, mixing appears to be less efficient to preserve compositional heterogeneity as the melts ascend into the crust.

The inference that melt heterogeneity from the mantle is preserved within the Miðfell melt inclusions is key for the discussions in the next chapter, which explores the carbon systematics of the Miðfell eruption and the mantle source.

Chapter 6

Miðfell: A gateway to Earth's primordial mantle

6.1 Introduction

In recent decades there has been increased interest in the carbon content of Earth's geochemical reservoirs due to the impact of atmospheric carbon on the habitability of our planet. This interest extends down into the solid Earth, as the planet's interior likely hosts a greater mass of carbon than that of the oceans, atmosphere, and crust combined. Exchange of carbon between the deep Earth and Earth's surface environment is controlled by geological processes acting over millions of years. These processes are central to our understanding of the geologic carbon cycle, which through the buffering of atmospheric carbon has ultimately created surficial conditions ideal to foster life (Hayes and Waldbauer, 2006).

One of the key requirements for understanding Earth's geologic carbon cycle is having estimates for all of Earth's carbon reservoirs. For reservoirs at the Earth's surface, such as the atmosphere or oceans, carbon capacity can be estimated relatively easily by direct observation. However, carbon estimates for the crust, mantle, and core are more difficult to acquire, and are coupled to greater uncertainty. Carbon heterogeneity within these reservoirs can complicate estimation further as the limited observations made on solid Earth reservoirs require careful interpretation to ascertain the length scale over which they are representative of the solid Earth.

The carbon content of the bulk mantle and bulk Earth have been estimated indirectly using planetary mass balance arguments from isotopic and elemental ratios (Marty, 2012). However, observations at mid-ocean ridges (MORs) have been used to infer the carbon content of the mantle on much shorter length scales, i.e. the mantle region sampled by a

MOR segment. CO₂ fluxes at MORs can be calculated from C/He ratio fractionation models, vesicle distribution models, and vesicle compositions of mid-ocean ridge basalt (MORB; Marty and Jambon. 1987; Javoy and Pineau 1991; Marty and Tolstikhin 1998; Cartigny et al. 2008; Burnard et al. 2014; Chavrit et al. 2014). These fluxes of CO₂ are assumed to directly relate to the carbon content of the MORB mantle source; depleted MORB mantle (DMM), hence providing a direct observational carbon estimate for the upper mantle. A further key constraint on this estimate comes from CO₂ and trace element measurements of rare undegassed suites of MORB (Saal et al., 2002; Michael and Graham, 2015; Le Voyer et al., 2017). The observation that CO₂ correlates strongly with certain incompatible trace elements (ITEs), such as Nb, Ba, and Rb, in these suites suggests that they appear to have similar compatibility when viewed over the whole mantle melting process. Therefore, CO₂/ITE ratios preserved in these undegassed MORB suites are inferred to directly reflect the C/ITE ratios of the source mantle, giving an estimate for DMM carbon content.

Previous measurements of CO_2/ITE ratios from undegassed MORB suites have estimated DMM carbon to be 37–50 ppm. CO_2/Nb ratios have been observed to vary between different MOR segments, indicating carbon heterogeneity within the DMM (Saal et al., 2002; Cartigny et al., 2008; Helo et al., 2011). No such direct observation has been made on the lower mantle. However, CO_2 degassing observations coupled with probabilistic magma supply rates at Hawaii suggest that the Hawaiian mantle plume source is richer in carbon than DMM (Anderson and Poland, 2017).

The assumption that CO_2/ITE ratios measured in undegassed MORB suites are representative of their source mantle has been questioned by trace element partitioning experiments on carbonated peridotite. It has been argued that at the initiation of carbonated peridotite melting, i.e. at low melt fractions, C and Nb can be fractionated due to the slightly more compatible nature of Nb (Dasgupta and Hirschmann, 2006; Dasgupta et al., 2013). This process could be responsible for observed high CO_2/Nb ratios in Nb-rich MORB. However, C and Ba are not fractionated as easily, suggesting that Ba is a better tracer for estimating mantle carbon content from CO_2/ITE ratios (Rosenthal et al., 2015). Carbonated peridotite melting experiments seek to explain the mobilisation of carbon from the mantle, in which it is stored as a carbon-rich phase, e.g. graphite or carbonate, rather than as a silicate-hosted trace element. This process could have implications for the CO_2/ITE ratios expressed in MORB.

This chapter discusses the trace element and carbon chemistry of olivine-hosted melt inclusions from the Miðfell eruption, presented in Chapter 5, to estimate the carbon content of mantle within the Miðfell source region. Firstly, this chapter presents the current understanding of solid Earth carbon along with published carbon content estimates for DMM from undegassed melt inclusion and MORB suites. Secondly, it explains the use of CO_2/ITE ratios to estimate mantle carbon, and presents the Miðfell and Kistufell melt inclusion suites. Thirdly, this chapter compares the two melt inclusion suites to a global melt inclusion database, before outlining the arguments for an undegassed primordial reservoir from the lower mantle as the origin of carbon enrichment observed within the Miðfell melt inclusions. The chapter ends with a discussion of the wider impact of such a carbon-rich lower mantle domain on Earth's carbon reservoirs and deep Earth carbon cycling.

6.2 Carbon in the solid Earth

The behaviour of carbon within the solid Earth is dependent on pressure, temperature, oxidation state, and composition of the host assemblage (Dasgupta and Hirschmann, 2010). At crustal pressures, carbon behaves as a very volatile element, readily degassing from magma at MORs and contributing to the flux of carbon from the upper mantle to the atmosphere (Dixon et al., 1991). At higher pressures, carbon takes the form of a C-rich accessory phase, such as graphite or carbonate, as both elemental C and CO_3^{2-} ions are incompatible within silicate lattices (Keppler et al., 2003; Shcheka et al., 2006). Therefore, it is assumed that even at low carbon concentrations, the majority of mantle is carbon saturated, with the nature of the carbon-bearing phase being dependent on pressure and oxygen fugacity.

The observation that MORB tends to be relatively oxidised, as suggested by ferric to ferrous iron ratios, suggests that carbon exists as carbonate beneath MORs (> 140 km; Dasgupta and Hirschmann 2006; Stagno et al. 2013). Experiments on carbonated peridotite show that as pressure increases, stable carbonate changes from being Ca-rich to Mg-rich, highlighting that carbon can be coupled to different elements within its host lithology (Dasgupta et al., 2004, 2006). At more reducing conditions in the mantle, which is mainly controlled by the amount of ferric iron within the mantle assemblage, carbon will exist as graphite and diamond (Dasgupta and Hirschmann, 2010; Stagno et al., 2013).

The speciation of carbon within the upper mantle is important for processes such as mantle melting. Carbonate has a low solidus temperature, so will readily form low fraction melts at depths greater than the anhydrous peridotite solidus (Dasgupta et al., 2013). Graphite and diamond are much more refractory, but they are key to redox melting; carbon is oxidised by the reduction of ferric iron and carbonated silicate melt is formed. Observations of erupted melt oxygen fugacity can be used to suggest the process by which carbon is mobilised from the mantle (Stagno et al., 2013).

Experiments have shown that at high pressure within the lower mantle, carbon can form complexes with iron, stabilising iron carbide and dissolving within iron metal (Frost et al.,

2004; Frost and McCammon, 2008; Dasgupta et al., 2009; Lord et al., 2009). The exact speciation of carbon is coupled to the availability of iron and other siderophile elements, which makes carbon important for the fractionation of elements between the mantle and the core (Cottrell et al., 2009). Even at lower mantle pressures, it is thought that elemental iron and iron carbonate can coexist because of self oxidation reactions (Oganov et al., 2008; Cerantola et al., 2017).

The behaviour of carbon within the solid Earth has mainly been determined by experiments on carbonated peridotites and rare inclusions within diamond. However, direct observations of mantle carbon have been made at MORs to place constraints on carbon concentration within the MORB source (Cartigny et al., 2008; Le Voyer et al., 2017).

6.2.1 CO₂ in MORB glasses

CO₂ dissolved in MORB glass and exsolved in vesicles can be used to gain insight into the carbon content of the DMM. However, due to the low solubility of CO₂ within silicate melt, the majority of MORB is partially degassed, losing information about initial CO₂ content (Dixon et al., 1991). Certain MORB samples can be used to estimate DMM carbon despite being partially degassed by using the composition of exsolved gas and MORB vesicular volume (Chavrit et al., 2014). Such methods have coupled CO₂ flux observations with geochemical enrichment to estimate $18-21 \pm 11$ ppm for DMM carbon, which is in agreement with estimates from ³He flux and C/³He ratios (Marty and Jambon., 1987; Marty and Tolstikhin, 1998). One highly vesicular MORB sample, 'popping rock' ($2\pi D43$) from the mid-Atlantic ridge (MAR), has been used to suggest that DMM carbon could be as high as 300 ppm, although such a highly vesicular basalt could represent bubble accumulation at the top of a magma chamber (Javoy and Pineau, 1991). Isotopic fractionation of ⁴He/Ar* (radiogenic Ar) during CO₂ degassing has been modelled to infer the carbon content of DMM beneath the SE Indian ridge (Burnard et al., 2014). All of these methods require some sort of modelled chemical reconstruction of CO₂ degassing. Seemingly more reliable estimates can be achieved by using undegassed samples.

In some MORB suites and olivine-hosted melt inclusions, which have been inferred to be undegassed, it has been observed that CO_2 correlates well with ITEs (Ba, Nb, Rb) (Saal et al., 2002; Michael and Graham, 2015; Le Voyer et al., 2017). This correlation has been interpreted as evidence for carbon behaving like an incompatible trace element during the mantle melting process, therefore CO_2/ITE ratios measured in undegassed erupted products are proportional to the C/ITE ratio of their source mantle. These ratios can be converted into carbon concentration by assuming an estimates of source mantle trace element content.

CO₂-incompatible trace element ratios

If the assumption that CO_2 and ITEs are coupled during mantle melting holds true then the CO_2/ITE ratio measured in undegassed MORB equals that of the MORB mantle source, from which the C/ITE ratio can be estimated. It must also be assumed that no modification of the magmatic ratio has occurred during ascent of the melt by either CO_2 degassing or assimilation. As asserted in the previous section, undegassed samples are deemed more reliable than samples requiring CO_2 to be reconstructed from vesicles. Olivine-hosted melt inclusions provide the greatest probability of preserving undegassed melt compositions, due to the early crystallisation of olivine within a magmatic system. Estimates of CO_2/ITE ratios for key undegassed melt inclusion and MORB suites are detailed in Table 6.1.

Reference	Locality	Туре	CO ₂ /Nb	CO ₂ /Ba
Saal et al. (2002)	Siqueiros, East Pacific Rise	MI	$239\ \pm 46$	100 ± 22
Cartigny et al. (2008)	MAR	MORB	570-730	-
Michael and Graham (2015)	MOR	MORB	$607\ \pm 327$	105 ± 9
Le Voyer et al. (2017)	MAR	MI	$557\ \pm 79$	97 ± 10
Hauri et al. (2017)	Borgarhraun, Iceland	MI	$391\ \pm 16$	$48.3\ \pm 2.7$

Table 6.1 CO₂/ITE ratios of undegassed melt inclusions and MORB glasses.

Siqueiros, Borgarhraun, and MAR values have been measured from melt inclusion suites that show strong correlations between CO_2 and ITEs, indicating that they are undegassed and likely preserve CO_2 /ITE ratios representative of their source mantle (Saal et al., 2002; Hauri et al., 2002a; Le Voyer et al., 2017). The MAR MORB sample assumes a degassing history to reconstruct initial CO_2 content and estimate an initial CO_2/Nb (Cartigny et al., 2008). The values of CO_2/ITE for the general MORB suite are calculated averages of many samples, which may be useful when considering the carbon budget of the DMM, however they hide the heterogeneity suggested by observations from different MOR segments. The variability in measured CO_2/Nb and CO_2/Ba values for the MOR system could be suggesting carbon heterogeneity within the DMM (providing that the CO_2/ITE variability is not the result of constant mantle carbon and heterogeneous trace element concentrations). More estimates of CO_2/ITE ratios are required from different points along the MOR system to describe the extent of DMM carbon variability more thoroughly.

Table 6.1 contains one estimate of CO₂/Nb from Iceland. The suite of melt inclusions is from the Borgarhraun eruption, at the North end of the northern volcanic zone. He isotopes suggest that there is not a strong plume signature in the Borgarhraun eruption ($R/R_a \sim 8$), suggesting that upper mantle material is more significant within the melt region, which rationalises the observation that Borgarhraun CO₂/Nb is within the range of MORB estimates. Iceland is known to exhibit high trace element variability comparable to the variability seen along the length of the MAR, therefore it could be expected that Iceland would also show variability in CO_2/ITE ratios and mantle carbon estimates (Peate et al., 2010; Shorttle and Maclennan, 2011).

Here, Miðfell and Kistufell were targeted for study because they both were previously shown to have high 3 He/ 4 He ratios (both ~ 17 R/R_a), indicating a primordial plume component within their mantle source regions (Breddam et al., 2000). Miðfell also shows primordial Ne and Xe isotopes (Harrison et al., 1999; Mukhopadhyay, 2012). Both eruptions contain highly forsteritic olivines hosting melt inclusions, giving them the highest chance of trapping undegassed primitive melts (Gurenko and Chaussidon, 1995; Breddam, 2002; Gurenko and Sobolev, 2006). These two characteristics suggest that olivine-hosted melt inclusions from both Kistufell and Miðfell could preserve information about the carbon content of the Icelandic mantle plume, which is suggested to be rooted within the lower mantle (Jenkins et al., 2016; Yuan and Romanowicz, 2017).

6.2.2 CO₂ in Kistufell and Miðfell melt inclusions

The previous section outlined the reasons for measuring CO_2/ITE ratios within olivine-hosted melt inclusions. Here, the observations from Kistufell and Miðfell melt inclusions are presented for comparison with other melt inclusion datasets. The Miðfell melt inclusions show a range of CO_2 content from 20–1120 ppm, while the matrix glass is generally degassed with a maximum concentration of 200 ppm CO_2 . Ba concentrations show two orders of magnitude variation (0.37–115 ppm; matrix glass 6.3 ppm; Figure 6.1a), and Nb varies by almost the same amount (0.04–22.9 ppm; matrix glass 0.78 ppm; Figure 6.1b). Reconstructed melt inclusions measured at Woods Hole Oceanographic Institution (WHOI) have compositions towards the enriched end of the melt inclusion suite: Ba, 12.4–62.2 ppm; Nb, 2.47–12.2 ppm; unreconstructed CO_2 720–1160 ppm (glass only); reconstructed CO_2 1340–4550 ppm (glass and vapour bubble combined).

Figure 6.1 shows that, in general, the CO_2 content of the Miðfell melt inclusions is variable at any given ITE composition, however the maximum CO_2 concentrations behave more systematically. At low ITE concentrations, maximum CO_2 content can be seen to increase as ITE concentration increases, up until 1120 ppm (at Nb = 0.2 and Ba = 3). For all ITE concentrations higher than these values, the maximum CO_2 content observed is also 1120 ppm. Melt inclusion glass analyses from WHOI also have a maximum value of 1170 ppm, however when recombined with the CO_2 content of exsolved vapour bubbles, the CO_2 content of these inclusions is higher than the rest of the dataset.



Figure 6.1 CO_2 content of Kistufell (purple) and Miðfell (red) melt inclusions against (a) Nb, and (b) Ba. Vapour bubble-melt inclusion CO_2 reconstructed melt inclusions from Miðfell plotted as diamonds.

At ITE concentrations equivalent to those of the Miðfell matrix glass, several melt inclusions plot in a vertical trend from the maximum measured CO₂ concentration down to ~ 30 ppm. Kistufell melt inclusions show low trace element variability, with very little difference in concentration between the melt inclusions and the matrix glass. This low variability has resulted in a vertical cluster of melt inclusion points at approximately the ITE concentration of the matrix glass from the maximum observed CO₂ concentration (1170 ppm) down to < 10 ppm (Nb = 2.85 ppm, Ba = 14.0 ppm; Figure 6.1).

The positive correlation between CO_2 and ITEs at low ITE concentrations, suggest that the three are behaving similarly during mantle processes, i.e. as incompatible elements. This region can be used to describe the maximum CO_2/Nb and CO_2/Ba ratios of Miðfell melt inclusions (Figure 6.1), which are summarised in Table 6.2 along with the maximum ratios for CO_2 reconstructed melt inclusions.

Table 6.2 CO₂/ITE ratios from Miðfell olivine-hosted melt inclusions.

Estimate type	CO ₂ /Nb	CO ₂ /Ba
Highest	5737 ± 987	566 ± 68
Reconstructed	1186 ± 127	236 ± 25
Smooth trace	1832 ± 316	396 ± 48

The Miðfell melt inclusions are hosted within highly forsteritic olivines ($X_{Fo}^{Ol} = 0.852-0.909$), with the highest trapping the most primitive melt and decreasing forsterite being a proxy for olivine crystallisation from the host magma. Figure 6.2 shows the variation of melt inclusion La/Yb and CO₂/Ba with MgO content, and melt inclusion CO₂/Ba with host olivine forsterite content. The highest CO₂/Ba (and CO₂/Nb) ratios are found within the



Figure 6.2 (a) La/Ba against melt inclusion MgO coloured by CO_2/Ba , (b) CO_2/Ba against melt inclusion MgO content and (c) host olivine forsterite content, both coloured by La/Yb ratio. The highest CO_2/Ba is found within the most depleted melt inclusions.

most MgO-rich melt inclusions, which are also hosted within the most forsteritic olivines. This observation suggests that the high CO_2/ITE ratios were present in the most primitive melts observed within the Miðfell suite. The melt inclusions with the highest CO_2/ITE ratios are also the most depleted, as shown by the low primitive mantle normalised trace element concentrations in Figure 6.3.

Figure 6.2c shows that CO_2/Ba variability decreases as host olivine forsterite content decreases, suggesting that as the Miðfell melts evolved and crystallised olivine, the high CO_2/ITE ratios were diluted by melt mixing and/or CO_2 degassing. In Chapter 5 it was concluded that melt heterogeneity was being preserved during olivine crystallisation, as trace element ratio variability was maintained as host olivine forsterite decreased. If CO_2/ITE ratios are seen to decrease during the same period of melt mixing and crystallisation, then perhaps CO_2 is able to diffuse and mix between melts faster than trace elements, reducing CO_2/ITE variability more quickly than trace element ratios. Alternatively, degassing is occurring as the melts cool, mix, and crystallise, again causing CO_2/ITE ratios to drop. Olivine decrepitation could also be contributing to CO_2/ITE decrease through the loss of CO_2 vapour bubbles from melt inclusions (Maclennan, 2017).

Comparing the Miðfell CO_2/ITE ratios to the melt inclusion suites detailed in the above sections, it can be seen that Miðfell has preserved considerably higher CO_2/Nb and CO_2/Ba ratios (Table 6.1). These high ratios require the context of global melt inclusions and MORB glasses to ascertain the level of carbon enrichment recorded in the Miðfell melt inclusions.



Figure 6.3 Trace element concentrations for melt inclusions by CO_2/Nb ratios for (a) Miðfell and (b) Kistufell, and coloured by CO_2/Ba ratios for (c) Miðfell and (d) Kistufell. Average matrix glass compositions plotted as dotted lines. Undegassed basaltic glass and melt inclusions suites are plotted as grey fields; dark grey, MORB glasses (Michael and Graham, 2015); medium grey, Mid-Atlantic Ridge melt inclusions (Le Voyer et al., 2017); light grey, Siqueiros melt inclusions (Saal et al., 2002).

6.2.3 Global melt inclusion suite

Trace and volatile element concentrations of melt inclusions and MORB glasses have been collated from several published sources (see Appendix B for references). Sample localities cover a wide range of magmatic settings, covering MORB, ocean island basalt (OIB), and continental rifts. MORB plots at the lowest ITE concentrations, with OIB plotting slightly higher, and continental rift basalts at the highest ITE concentrations. This distribution of magmatic settings in ITE space is due to the effect of source composition and extent of melting on basalt enrichment. The extent of mantle melting is further controlled by mantle potential temperature, lithospheric thickness, and source composition. ITE-enriched basalts are formed by low fractions of melting and/or enriched source compositions.

Melt inclusions from Iceland show high trace element variability, Nb = 0.5-50 ppm and Ba = 10-500 ppm, which cover the full range of OIB samples and extend down across the MORB suite. This variability clearly highlights that melts are generated beneath Iceland due to the interaction between a mantle plume and a MOR, tapping more primitive mantle as well as DMM.

Miðfell melt inclusions plot at the depleted end of the global range, but show considerable variability in ITE concentrations (as seen with the Iceland suite). Miðfell covers the MORB



Figure 6.4 CO_2 content against (a) Nb and (b) Ba abundances, and CO_2 trace element ratios against (c) Nb and (d) Ba abundances, for Miðfell melt inclusions compared with a global compilation of melt inclusions and basalt glasses. One sigma error ellipses are at the top of each plot, with relative error dependent on the trace element concentration. DMM estimates of CO_2/ITE ratios are plotted as a grey field; high Miðfell CO_2/ITE ratios are plotted as a pink field; Borgarhraun, Iceland CO_2/Nb ratio is plotted as a blue field.

region and extends up into OIB. Figures 6.4c–d again show that the highest CO_2/ITE ratios are found in the most ITE depleted inclusions, with several of these inclusions from Miðfell plotting higher than the global dataset. These high ratios are due to low ITE concentrations, and is not attributed to extremely high CO_2 measurements (cf. Helo et al. 2011). Again the Kistufell samples show a narrow range of ITE concentrations, sitting at the lower end of the OIB region, but showing similar maximum CO_2/ITE ratios as Miðfell melt inclusions with the same ITE concentration. Figure 6.4c shows that the maximum CO_2/Nb preserved in Kistufell melt inclusions is comparable to that of the Borgarhraun melt inclusion suite (Hauri et al., 2017).

Several of the Miðfell melt inclusions plot at higher CO_2/ITE ratios than the majority of the global dataset, however Figures 6.4c–d show that there are three data points from the global dataset that are comparable to Miðfell. The blue circle at Ba = 6 ppm (Nb = 3 ppm) is a plagioclase-hosted melt inclusion from Axial Seamount, Juan de Fuca (Helo et al., 2011), which has been suggested to have decoupled CO_2 and Nb (Le Voyer et al., 2017). The orange square at Ba = 34 ppm (Nb = 9 ppm) is a vapour bubble-melt inclusion reconstruction data point, which includes a vapour bubble from the trapping of an exsolved fluid phase, rather than CO_2 exsolution from the melt inclusion after entrapment (Hartley et al., 2014). The blue circle at Ba = 76 ppm (Nb = 10 ppm) has been calculated by reconstructing the CO_2 concentration of a partially degassed, vesicular, enriched MORB (Michael and Graham, 2015). The significance of these three data points is questionable as they suffer from substantial uncertainty or are not olivine-hosted melt inclusions.

Figures 6.4c–d define a negative trend in the global dataset, with the maximum CO_2/ITE ratio measured at any ITE concentration decreasing as ITE concentration increases. This trend could be formed by a suite of melts that have the same CO_2 concentration, but variable ITE concentrations. Figures 6.4a–b show that within each magmatic setting there is some CO_2 maximum variability, however across the global dataset the trace element concentrations are more variable. The reduced CO_2 variability could be due to degassing or olivines reaching a decrepitation limit, which would have the potential to modify a suite of melts down to a constant CO_2 concentration or lower.

6.2.4 CO₂ solubility and degassing

 CO_2 will readily degas before other volatile elements upon decompression of a magma due to its low solubility within silicate melt (Dixon et al., 1991). Depending on the CO_2 content of a magma, degassing can start at mid-crustal pressures (> 5 kbar) and deeper. Once melts start degassing they ascend through the magmatic system close to CO_2 saturation, exsolving CO_2 into vapour bubbles. This process will cause CO_2 and ITEs to become decoupled, reducing measured CO_2/ITE ratios in melt inclusions and predicting mantle carbon content lower than undegassed estimates. Figures 6.4 a and b show degassing vectors, indicating that vertical arrays of melt inclusions suggest degassing and that undegassed melts lose their strong correlation with ITEs.

The global trend highlighted in the previous section, with ITE-enriched melts having low CO_2/ITE ratios, can be explained by degassing. For a suite of melts all with the same CO_2/ITE ratio but variable ITE concentrations, melts with high ITEs will also have high CO_2 . As melts decompress in a magmatic system, melts with higher dissolved CO_2 content (enriched melts) will reach CO_2 saturation before melts with lower CO_2 . Upon CO_2 saturation the high- CO_2

(enriched) melt will exsolve CO_2 and partially degas until the dissolved CO_2 content goes below CO_2 saturation. If several melts degas at the same pressure, then providing they have the same major element chemistry they will have the same dissolved CO_2 content (Shishkina et al., 2010). Though melts that experience high amounts of degassing can be identified by high ⁴He/⁴⁰Ar* ratios (Burnard et al., 2014). If this hypothetical melt population has variable ITE concentrations then it will produce a negative trend in CO_2/ITE -ITE space, as shown by the global dataset in Figures 6.4c–d.

On a single eruption locality basis it is conceivable that melts of variable composition could degas at the same pressure, i.e. within a magma chamber, producing a linear negative trend in CO_2/ITE -ITE space. However, globally this is unlikely as the depths of magma chambers are variable in pressure (depth) due to crustal heterogeneity and density variability of melts entering magmatic systems. Hence the global dataset does not lie on a perfectly linear array in Figures 6.4c–d.

 CO_2 solubility is dependent on the major element chemistry of the melt and can be shown by a simple solubility model (Shishkina et al., 2010). Figure 6.5a shows lines of saturation composition at different pressures for melts with average MORB major element chemistry (Gale et al., 2013). A more enriched composition has also been calculated at various pressures, indicating that increased concentrations of alkalis increase CO_2 solubility (Hudgins et al., 2015). Saturation CO_2 content was calculated at each pressure and this value was used to calculate the maximum CO_2/Ba ratio for melts of this CO_2 content and Ba varying 0.1–2000 ppm. Figure 6.5b shows the effect of varying major element chemistry between average MORB and a more alkaline melt with Ba content; i.e. 0.1 ppm Ba = 100% average MORB major element chemistry, 2000 ppm Ba = 100% alkali basalt major element chemistry.

The CO₂ solubility model detailed above and observations of the global dataset suggest that the majority of samples have been partially degassed. Melts most likely to have been unaffected by degassing are depleted in trace elements, suggesting that their initial CO₂ content was also low. Therefore CO₂/ITE ratios from depleted melts are required to estimate the carbon content of the mantle.

Miðfell melt inclusions

The solubility model in the previous section suggested that depleted melts are required to measure undegassed CO_2 concentrations. The same model can be applied to Miðfell melt inclusions to constrain the effect of degassing on their CO_2 content. CO_2 concentrations can be converted into saturation pressures with knowledge of the major element chemistry of the melt (Shishkina et al., 2010). The Miðfell melt inclusions show a range of CO_2 content,



Figure 6.5 CO₂ saturation curves based on the Shishkina et al. (2010) CO₂ solubility model for silicate melt calculated at pressures of 1, 2, 5 and 10 kbar. (a) Separate CO₂ saturation trends for depleted (Gale et al., 2013, pink) and enriched (Hudgins et al., 2015, green) melt compositions, both match relevant global data at ~ 2 kbar. (b) Mixed melt compositions with major elements changing linearly with Ba enrichment. Global melt inclusion and MORB glass dataset plotted for comparison.

but the maximum saturation pressure recorded by CO_2 concentration is around 1.8 kbar (equivalent to ~ 5 km depth). Assuming that melts were saturated in CO_2 at the time of entrapment, this value would suggest that melts had ascended to reasonably shallow levels in the crust beneath Iceland before significant melt evolution had taken place.

The vapour bubble-melt inclusion reconstructions provide more information to suggest that Miðfell melt inclusions were trapped deeper. Reconstructed melt inclusions have considerably higher CO₂ content than glass only melt inclusions, with a maximum saturation pressure of around 4.3 kbar (equivalent to ~ 13 km depth; Figure 5.6c). This observation revises the previous estimate of magma chamber depth to the mid-crust beneath Miðfell. One of the olivines measured at WHOI ($X_{Fo}^{Ol} > 0.90$) contained numerous fluid inclusions, along with a melt inclusion containing a vapour bubble. The presence of fluid inclusions suggests that a fluid phase was present during olivine crystallisation, implying that degassing had already started and was likely from the most CO₂ and trace element enriched melts.

The two maximum saturation pressures recorded in the Miðfell melt inclusions (before and after vapour bubble reconstruction) are very different, but can be explained by olivine integrity during magma ascent (Steele-Macinnis et al., 2011). The pressure drop on the melt inclusion related to thermal contraction during cooling and decompression during melt transport leads to the exsolution of CO_2 out of the melt inclusion and formation of a vapour bubble (Anderson et al., 1989; Lowenstern, 1995, 2015). This process is also sensitive to increases in dissolved CO_2 melt content and changes in major element chemistry due to postentrapment crystallisation (PEC), as this will affect the CO_2 solubility in the melt inclusion. If the host olivine keeps its structural integrity during this process of CO_2 exsolution then the vapour bubble will be preserved in the melt inclusion. If, however, the olivine decrepitates then the vapour bubble will be lost through cracks in the olivine (Maclennan, 2017).

Olivine decrepitation can be avoided if magmas undergo substantial pre-eruptive cooling, as observed for some melt inclusions from Laki EVZ, which record reconstructed vapour bubble-melt inclusion CO_2 content equivalent to > 5 kbar. The CO_2 content of the global melt inclusion suite suggests that either the majority of melts degassed prior to entrapment or the majority of host olivines underwent decrepitation at the predicted threshold of ~ 2.5 kbar (Maclennan, 2017). The Miðfell melt inclusions show vapour bubble formation in the most enriched melts, which are expected to contain the most CO_2 . Therefore, the more depleted melt inclusions are not expected to have formed vapour bubbles due to their predicted lower initial CO_2 content. This assertion is important for CO_2/ITE ratios, as it again suggests that the most depleted melt inclusions are the most likely to have preserved undegassed CO_2 concentrations, and therefore record unmodified CO_2/ITE ratios.

6.3 Miðfell carbon enrichment

The previous sections have concluded that undegassed CO_2 concentrations are most likely found within the most depleted melt inclusions, This section compares the most depleted, and therefore undegassed, melt inclusions from Miðfell to assumed undegassed melt inclusion and MORB suites. Globally there are three main datasets that have identified undegassed melt inclusions and MORB glasses from the MOR system. Saal et al. (2002) measured olivine-hosted melt inclusions from Siqueiros, EPR; Michael and Graham (2015) compiled data from all around the MOR system, including data from Cartigny et al. (2008); and Le Voyer et al. (2017) measured olivine-hosted melt inclusions from the MAR. Hauri et al. (2017) also presented a CO_2/Nb ratio = 391 ± 16 and $CO_2/Ba = 48.3 \pm 2.7$ for olivine-hosted melt inclusions from Borgarhraun NNVZ Iceland.

Figure 6.6 compares Miðfell melt inclusions to these suites of undegassed melts and the CO_2/Nb ratio for Borgarhraun. Melt inclusions from the MAR dataset produce the strongest correlation between CO_2 and ITEs, giving the clearest evidence for undegassed behaviour (Le Voyer et al., 2017). The other two datasets define a trend of maximum CO_2/ITE ratio for variable ITE concentration, however there is some scatter within each sample set down to lower CO_2/ITE at the same ITE concentration (Saal et al., 2002; Michael and Graham, 2015).



Figure 6.6 CO_2 content against (a) Nb and (b) Ba abundances, and CO_2 trace element ratios against (c) Nb and (d) Ba abundances, for Miðfell melt inclusions compared with undegassed MORB melt inclusions and glasses (Hauri et al., 2002a; Saal et al., 2002; Michael and Graham, 2015; Le Voyer et al., 2017). One sigma error ellipses are at the top of each plot, with relative error dependent on the trace element concentration. Coloured fields are the same as Figure 6.4.

These three suites have variable CO₂/Nb (239–557) and CO₂/Ba (\sim 100); with Borgarhraun CO₂/Nb lying within the MORB range. Miðfell melt inclusions show CO₂/ITE ratios higher than any of the previous measurements from studies on MORB. Figures 6.6c and d show that for any given ITE concentration, Miðfell melt inclusions show higher CO₂/ITE ratios. The highest Miðfell CO₂/Nb is 10–20 times greater than MORB, and the highest CO₂/Ba is 1.3–5.7 times greater than MORB. This indicates that some of the Miðfell melt inclusions, or more generally some of the melts within the Miðfell magmatic system, are more enriched in carbon than equivalent MORB melts.

The Miðfell CO_2/ITE ratios are higher than those recorded within basaltic glass from MOR settings, which suggests that some process or source characteristic has led to carbon

enrichment beneath Miðfell, and that this enrichment is not present within the depleted MORB mantle.

6.4 Carbon enrichment from crustal processes

The following section explores the crustal processes that could have modified Miðfell melts and possibly lead to carbon enrichment. Crustal assimilation was discussed in the context of Miðfell in Chapter 5. There is evidence in Iceland of interaction between ascending melts and crustal material (Nicholson et al., 1991; Eason and Sinton, 2009; Brounce et al., 2012). Miðfell has likely interacted with gabbroic material during magmatic ascent, and in rarer cases (excluded from the melt inclusion dataset) has assimilated just plagioclase. Crustal assimilation becomes particularly important to constrain if assimilated lithologies are carbon-bearing, as this could greatly modify initial CO₂/ITE ratios and produce ratios recorded by Miðfell melt inclusions that are unrepresentative of the source mantle.

6.4.1 Simple plagioclase addition

Chapter 5 identified five melt inclusions from Miðfell that showed very high Sr anomalies $(Sr/Sr^* > 10)$ and coupled ITE dilution, which were interpreted as having a strong plagioclase assimilation signal. Therefore, they were removed from the melt inclusion dataset and not discussed further. Due to the low solubility of carbon within silicate phases, it is expected that plagioclase addition could only reduce CO₂/ITE ratios by diluting CO₂ concentrations (Keppler et al., 2003; Shcheka et al., 2006). Dissolution of plagioclase within silicate melt would change the major element composition of the magma. This process could result in a slight increase in CO₂ solubility due to increased concentrations of Ca and Na (Shishkina et al., 2014).

6.4.2 Interaction with gabbroic material

Gurenko and Sobolev (2006) showed clear evidence for the interaction between primitive Miðfell melts and gabbroic material observed as xenoliths in Miðfell erupted products. This has been supported by modelling of gabbro interaction in Chapter 5, which indicates that trace element patterns change as a result of gabbro assimilation. This interaction can be inferred from trace element anomalies, such as Ba > Nb, as shown by 12 melt inclusion trace element patterns in Figure 5.10c.

Trace element concentrations have been shown to vary due to gabbro addition, however it is important to estimate the CO_2 contribution from gabbroic material. Gurenko and Sobolev



Figure 6.7 Enrichment in CO_2/X compared to Michael and Graham (2015) CO_2/X for each incompatible trace element. (a) Melt inclusions with smooth trace element patterns, four with highest CO_2/Ba plotted; (b) melt inclusions with spiky trace element patterns, i.e. element anomalies present on normalised trace element spider plot, four with highest CO_2/Ba plotted. Other melt inclusion suites plotted for comparison.

(2006) and observations on xenoliths sampled by B. Winpenny (pers. comm.), suggest that there is no carbonate material present in the Miðfell gabbroic xenoliths. If the Miðfell olivines can be assumed to have crystallised at mid-crustal pressures, as suggested by barometry, then perhaps they are less likely to have interacted with carbonated crust, which is more abundant near the surface.

The change in initial CO_2/ITE ratios due to interaction with gabbroic material/melt can be described simply by the following equation:

$$\left(\frac{\text{CO}_2}{\text{ITE}}\right)_{\text{mix}} = \frac{\text{CO}_{2 \text{ initial}} \times (1-F) + \text{CO}_{2 \text{ gabbro}} \times F}{\text{ITE}_{\text{initial}} \times (1-F) + \text{ITE}_{\text{gabbro}} \times F},$$

where *F* is the fraction of gabbro melt in the mixture. If $CO_{2 \text{ gabbro}} = 0$, as suggested by observations on Miðfell xenoliths, then $(CO_2/ITE)_{\text{mix}}$ can only decrease when primitive melts interact with gabbroic material, assuming that $ITE_{\text{gabbro}} > ITE_{\text{initial}}$.

To avoid the complication of gabbroic material interaction, the melt inclusion suite can be filtered to remove melts that likely interacted with gabbro, i.e. those with primitive mantle normalised Ba > Nb and negative Zr anomalies. This filter leaves melt inclusions with a smooth trace element pattern, from which the highest CO₂/ITE ratios preserved are CO₂/Ba_{smooth} = 396 ± 48 and CO₂/Nb_{smooth} = 1832 ± 316 (Table 6.2). Comparing these ratios to the undegassed MORB suites in Figure 6.6 shows that these melt inclusions also have elevated CO₂/ITE ratios, indicating carbon enrichment.

Figure 6.7 shows the relative enrichment in CO_2/ITE ratios of Miðfell melt inclusions with respect to average MORB ratios (Michael and Graham, 2015). Figure 6.7a shows the

highest four melt inclusions with smooth trace element patterns, while Figure 6.7b shows the highest four melt inclusions with spiky trace element patterns. The smooth melt inclusions show considerable enrichment in $CO_2/ITEs$ for the most incompatible elements (Ba, Nb, and La). The spiky melt inclusions show the greatest ratio enrichment with trace elements that have negative anomalies after interaction with gabbro (Nb and Zr). These melt inclusions show higher ratios than the smooth inclusions, however there is a possibility that their high ratios are due to interaction with gabbro, therefore they cannot be used for estimating mantle carbon. Carbon enrichment observed in the melt inclusions with smooth trace element patterns must be have existed before melts entered the crust. Therefore, it is likely that some sort of mantle process or source characteristic has produced the high CO_2/ITE ratios seen in Miðfell.

6.5 Carbon enrichment from mantle processes

The previous section ruled out the involvement of crustal processes in enriching the depleted Miðfell melt inclusions in carbon. Here, processes pertaining to the mantle are explored to explain this carbon enrichment. However, it is well understood that the mantle is heterogeneous, therefore the carbon signal seen in Miðfell melt inclusions may originate from one of many mantle lithologies, represent a mixture of lithologies, or be controlled by the mantle melting process.

Several studies have used trace elements, radiogenic and stable isotopes to conclude that there is heterogeneity in the mantle beneath Iceland, with inferred components of: depleted MORB mantle, primordial mantle, recycled oceanic crust, and a refractory lithology (Fitton et al., 1997; Chauvel and Hémond, 2000; McKenzie et al., 2004; Thirlwall et al., 2004; Peate et al., 2010; Shorttle and Maclennan, 2011; Shorttle et al., 2014). The Icelandic mantle plume, which is deep-rooted in the lower mantle, is thought to be responsible for bringing some of these components up into the melting region beneath Iceland (Rickers et al., 2013; Jenkins et al., 2016; Yuan and Romanowicz, 2017).

Miðfell melt inclusions have CO_2/ITE ratios elevated with respect to MORB values, therefore DMM is unlikely the cause of carbon enrichment within the Icelandic mantle. Instead it is likely that one of the other mantle components is responsible for creating melts with high CO_2/ITE ratios; namely (i) carbonatite melts, (ii) recycled oceanic crust, and (iii) primordial mantle, which will be discussed in the following sections.

6.5.1 Carbonatite melt addition

Carbonatite and carbonated silicate melts erupted on ocean islands are observed to contain high concentrations of CO_2 , and as a result have high CO_2/ITE ratios (Hoernle et al., 2002). However, no carbonatite melts have been observed as erupted products on Iceland. Experiments conducted on carbonated peridotite have suggested that the formation of carbonatite and carbonated silicate melts is key to the initial mobilisation of carbon from the mantle assemblage (Dasgupta et al., 2009; Dasgupta and Hirschmann, 2010; Dasgupta et al., 2013). However the role of carbonatite melt in enriching silicate melts at shallow depths within the melt region beneath Iceland is more uncertain.

The mixing between primitive Miðfell melts and carbonatite melts can be modelled to test if synthetic melts can be made to match the trace element patterns and CO_2 content of Miðfell melt inclusions. Carbonatite melt compositions for use in this model can be acquired from observations of natural samples from ocean islands, or determined from experimental lherzolite-carbonatite melt partition coefficients (Hoernle et al., 2002; Dasgupta et al., 2009). Figure 6.8a shows the primitive mantle normalised trace element compositions of natural carbonatite melts calculated from partition coefficients. The two suites show very similar trace element patterns, but not the same concentrations, with natural samples around 10 times greater.

These carbonatite melts are very rich in trace elements, therefore mixing with primitive Miðfell silicate melts will result in trace element enrichment. Instead a depleted melt composition has been chosen to represent primitive Miðfell melt (Figure 6.8b). Depleted silicate melts have likely formed near the top of the melting region beneath Iceland, therefore carbonatite mixing with a depleted composition is representative of trace element modification by late carbonatite melt addition near the top of the melting region.

Figure 6.8b shows the linear mixing of the initial depleted Miðfell melt composition and the Dasgupta et al. (2009) carbonatite composition, with mixing lines representing fractions increasing in 1% steps, up to 10% carbonatite addition. The fit between the mixed melt composition and a target Miðfell melt inclusion (containing high CO₂/Ba) can be quantified using a reduced chi-squared statistic, χ^2_{ν} :

$$\chi_v^2 = \frac{\chi^2}{v}$$
, where $\chi^2 = \sum_i \frac{(O_i - C_i)^2}{\sigma_i^2}$.

v is the number of degrees of freedom, O_i is the observed concentration of component *i*, C_i is the calculated concentration of component *i*, and σ_i^2 is the variance of repeat measurements of component *i*.



Figure 6.8 (a) Primitive mantle normalised trace element patterns of carbonatite melts (natural carbonatites (Hoernle et al., 2002), beige; average of natural carbonatites, brown; Dasgupta et al. (2009) experimental carbonatite, navy). (b) Mixed melt compositions of depleted Miðfell composition (red) and Dasgupta et al. (2009) carbonatite, up to 10% carbonatite addition in steps of 1%. (c) Estimates of misfit between mixed melt composition and target composition (highest CO_2/Ba melt inclusion composition). Minima give the fraction of carbonatite required to produce the smallest misfit.

Figure 6.8b shows that the mixed melts do not match the trace element profiles of the Miðfell melt inclusion population. The fit between the modelled compositions and a target melt inclusion composition can be calculated using the reduced chi-squared statistic detailed above. Figure 6.8c plots the quality of fit, quantified by the reduced chi-squared statistic, for five different mixed melt compositions against the fraction of carbonatite melt added to the starting composition. The initial starting composition has the best fit for most of the carbonatite compositions mixed in this model.

Geochemically there is very little evidence for the direct interaction of carbonatite melt with depleted Miðfell melt compositions. There is also a melting depth argument that makes it difficult to rationalise late carbonatite melt addition. Carbonatites form at depths greater than 200 km, while depleted silicate melts form at the top of the melting region, < 60 km (Dasgupta et al., 2010, 2013). For a carbonatite melt to interact with a shallow level depleted melt, it must transit 150 km of mantle without reacting with silicate material on the way. Carbonatite melts are thought to be very reactive with silicate residue, therefore

this assumption of no modification during melt transport seems unfounded (Dasgupta and Hirschmann, 2010).

6.5.2 Carbon liberation from the mantle

Carbonatite melts may not have mixed directly with primitive depleted Miðfell melts to produce elevated CO_2/ITE ratios, however they may have been important in the initial mobilisation of carbon from the mantle assemblage and the potential fractionation of carbon, Nb, and Ba (Dasgupta and Hirschmann, 2006; Dasgupta et al., 2009; Rosenthal et al., 2015).

The nature of carbon-rich melt formed at depth in the mantle is dependent on the oxidation state of carbon within the mantle assemblage. Oxidised carbon, i.e. CO_3^{2-} , is very incompatible in silicate minerals, so exists as a carbonate accessory phase, which has a low solidus temperature and will therefore melt to form small fraction carbonatite melts at great depth (> 300 km). Reduced carbon, in the form of graphite or diamond, can undergo redox melting at low pressures, reducing ferric iron within the mantle assemblage to form CO_2 in a carbonated silicate melt (Dasgupta and Hirschmann, 2010).

Carbonatite melt is thought to be highly mobile and easily expelled from a solid matrix, even at low melt volumes, allowing the melt to migrate through the silicate mantle assemblage (Hunter and McKenzie, 1989; Minarik and Watson, 1995; Minarik, 1998; Hammouda and Laporte, 2000). It has been suggested that the process of carbonated silicate melt formation can be thought of as a reaction front, with carbonatite melt percolating through a solid peridotite assemblage, progressively reacting with silicate phases to increase the SiO₂ content of the melt (Dasgupta and Hirschmann, 2010). The evolution of this carbonatite melt into carbonated silicate melt is thought to ultimately lead to the generation of carbonated basalt (Hirose, 1997; Dasgupta et al., 2007a,b; Foley et al., 2009).

The reactive nature of carbonatite melt means that it quickly evolves by dissolving silicate phases, causing the partitioning of trace elements between the solid mantle assemblage and the carbonated silicate melt (Hirose, 1997; Dalton and Presnall, 1998; Moore and Wood, 1998; Gudfinnsson, 2005; Dasgupta et al., 2007a,b; Brey et al., 2008). If a steady state of rising carbonated silicate melt within an upwelling peridotite residue is achieved, then it can be argued that all melts interact with the same solid residue. A solid residue that is progressively more depleted due to reaction with carbonatite/carbonated silicate melt as the residue gets shallower. In this situation carbonated silicate melts would reach the anhydrous peridotite solidus with very similar compositions. However, this assumes that (i) the overlying solid residue is homogeneous at any given depth, (ii) melts ascend at a constant rate, (iii) no melt mixing occurs, and (iv) carbon enrichment is homogeneous, i.e. initial carbonatite melts are formed from the same melt fraction.

However, the mantle is not thought to be homogeneous, therefore ascending melts are likely to interact with streaks of different lithologies, depleting the solid residue of incompatible trace elements and creating isobaric melt heterogeneity, but also variability between depth-integrated melts (Meibom and Anderson, 2003). Melts taking different paths through the carbonated silicate melting region could conceivably interact with different amounts of enriched lithology creating melt heterogeneity by the time they reach the anhydrous peridotite solidus. This process would create melts with trace element compositions weighted by the enrichment of the lithologies they interacted with during ascent; a weighted average of lithologies within the melting region.

If there is significant decoupling between carbonated melt and the solid peridotite residue, then CO_2/ITE ratios measured at the surface are representative of a large volume of heterogeneous mantle. However, if carbonated silicate melts are retained by their residue until depletion of the solid in carbon and ITEs, then CO_2/ITE ratios will represent the small parcel of mantle that the melt was in equilibrium with before extraction.

The strong correlation between CO_2 and ITEs in undegassed MORB glasses suggests that these components remain coupled over the entire mantle melting process, however care should be taken when interpreting these ratios. If the mantle is a heterogeneous mixture of different mantle components, each with a different geochemistry and melting behaviour, then melt heterogeneity reaching the surface is likely to be of reduced variance to that seen in the mantle (Rudge et al., 2013). An individual melt composition likely represents a unique blend of geochemistry, produced from varied proportions of melts from different mantle lithologies and extents of depletion.

6.6 Carbon enrichment in source mantle lithologies

The previous section highlighted the importance of mantle carbon mobilisation when interpreting observed CO_2/ITE ratios and the role of carbonatite melts in this process. This section goes on to discuss the mantle lithologies that could be responsible for the carbon enrichment of Miðfell melt inclusions, such as recycled oceanic crust or primordial mantle (Figure 6.9).

6.6.1 Recycled oceanic crust

There are several pieces of geochemical evidence for the presence of recycled oceanic crust (ROC) within the Icelandic melt source region. Studies have used noble gas, radiogenic, and stable isotopes, as well as trace element data to identify the signature of ROC in Icelandic

melts (Hofmann and White, 1982; Thirlwall et al., 2004). Sr-Hf-Nd isotopes have been measured from Þeistareykir, NNVZ, to infer that there is a recycled OIB component within the Icelandic mantle with an age of 400 Ma (McKenzie et al., 2004). Os-He isotopes have also been used to infer an ancient ROC reservoir within the Icelandic source mantle, which from mixing models is older than 1 Ga, and has mixed with primitive mantle and DMM (Brandon et al., 2007). This observation is in agreement with other studies that suggest the Icelandic mantle plume is triggered by ROC deep within the mantle and potentially linked to ultra-low shear wave velocity zones at the core-mantle boundary (Hofmann and White, 1982; Yuan and Romanowicz, 2017).

Studies comparing the ³He/⁴He ratios, radiogenic (Sr-Nd-Hf-Pb) and stable (δ^{18} O) isotopic compositions, and trace element content of several Icelandic eruptions suggest that trace element enriched melts are not the carriers of high ³He/⁴He (> 15 R/R_a) ratios (Peate et al., 2010). Instead, they imply that enriched melts are the geochemical signal of degassed, recycled components like ROC. Major element and trace element compositions can be used to define depleted and enriched endmembers for the different regional rift zones on Iceland, with trace element ratios then able to predict the contribution of pyroxenite melt to different Icelandic eruptions (Shorttle and Maclennan, 2011; Shorttle et al., 2014). Pyroxenite melt is thought to be the product of ROC melting, which due to high pressure-temperature metamorphism has an eclogite or pyroxenite lithology.

Carbon content of recycled oceanic crust

The presence of ROC within the Icelandic mantle is reasonably well-established, however the carbon content of this lithology is much more uncertain as there is much debate as to the efficiency of carbon recycling through subduction zones and into the deep mantle (Dasgupta and Hirschmann, 2010; Kelemen and Manning, 2015; Clift, 2017; Mason et al., 2017). The efficiency of carbon subduction down into the deep mantle is crucial for the C/ITE ratios of ROC. Carbon stability within subducting slabs is dependent on (i) oxidation state; (ii) the carbon-rich phase present; (iii) carbon mobility in fluids; (iv) carbon reactivity with the subducted assemblage; and (v) the thermal structure of the subduction zone. Experiments on carbonated basalts predict that the near-solidus phase equilibria of carbonated basaltic eclogite are complex, making it difficult to ascertain the stability of carbonate down through a subduction zone (Hammouda, 2003; Dasgupta et al., 2004; Yaxley and Brey, 2004; Dasgupta et al., 2005). Given the variation in modern-day subduction conditions, and the high likelihood of temporal variations to subduction mechanics, it is hard to predict the nature of carbon within deep ancient ROC lithologies.



Figure 6.9 Schematic diagram showing potential carbon-rich reservoirs in the deep mantle: primordial undegassed reservoir; recycled oceanic crust that might be associated with an Ultra Low Velocity Zone (ULVZ), and the ambient lower mantle. Transport of deep volatiles to eruption on Iceland is via a mantle plume, thought to be rooted on an ULVZ. The lithological nature of the ULVZ remains uncertain.

Observations of modern subduction lithologies can be used to predict the behaviour of carbon within currently subducting slabs. Modern oceanic crust predominantly contains carbonate, as Earth's surface environment is oxidising (Dasgupta and Hirschmann, 2010; Kelemen and Manning, 2015). Upon subduction it is thought that carbonate breaks down, mobilising carbon into the overlying mantle, and ultimately back out into the atmosphere through volcanism. Large ion lithophile elements (Ba, Rb, K) are fluid mobile, so will also be lost through subduction zone processing, resulting in the fractionation of Ba, C and Nb. Organic carbon is thought to pass through a subduction zone more readily, due to the fluid immobile state of reduced carbon, potentially providing one of the main ways of fluxing carbon into the deep mantle (Clift, 2017).

Ferric iron estimates for the Miðfell melt inclusions, measured by XANES (see Chapter 2 for details), indicate relatively oxidised conditions with 17% ferric iron in the Miðfell matrix glass. This oxidation state suggests that graphite was unlikely stable in the source lithology and residue. If graphite had been present in the source, then it would have been oxidised to CO_2 or carbonate by the reduction of ferric iron during redox melting upon decompression through the melt region beneath Iceland (Dasgupta et al., 2010; Stagno et al., 2013).

Limitations to our knowledge of subduction zone element processing mean that it is difficult to characterise the geochemistry of Icelandic ROC. Partial melting of ROC is thought to produce trace element enriched melts (Brandon et al., 2007). If ROC also contains high C/ITE ratios, then melts produced will be very high in CO₂, and are more likely to degas deep in magmatic systems, making their initial CO₂/ITE ratios less likely to be preserved for measurement at the surface. If the elevated CO₂/ITE ratios seen in depleted Miðfell inclusions are from a trace element enriched ROC component, then extensive amounts of fractional melting are required to produce melts with such depleted chemistry.

The carbon enrichment in Miðfell is more likely to be from ROC than late stage addition of carbonatite melt, however ROC as the source of carbon is by no means a certainty. Degassing of the enriched Miðfell melts has obscured the amount of trace element enrichment in the carbon-rich source lithology, which could have attributed the carbon enrichment to ROC. The presence of such high CO_2/ITE ratios within depleted melts can be more easily explained by the melting of a mantle lithology more depleted in trace elements than ROC.

6.6.2 Primordial mantle reservoir

Recycled oceanic crust cannot be ruled out as being responsible for carbon enrichment, but here a mantle lithology more likely contributing to carbon enrichment is discussed; primordial undegassed mantle. The presence of an undegassed primordial mantle reservoir, thought to be present in the lower mantle beneath Iceland and tapped by the Icelandic mantle plume, has previously been established by observations of noble gas isotope ratios.

Noble gas isotope systematics

The distribution of high-³He/⁴He isotopic signatures around Iceland has long been known, with Miðfell showing a ³He/⁴He ~ 17 R/R_A (Breddam et al., 2000). High ³He/⁴He ratios indicate enrichment in the primordial ³He isotope, which is associated with primordial undegassed reservoirs that have been isolated from Earth's surface for billions of years. The correlation between high ³He/⁴He and ¹⁸⁷Os/¹⁸⁸Os suggests that a mixture of primordial undegassed mantle and ROC is sourced by the Icelandic mantle plume and expressed by Miðfell lavas (Brandon et al., 2007).

Miðfell also shows a uniquely primitive Ne isotopic signature with moderately nucleogenic ²¹Ne (²¹Ne/²²Ne lower than MORB) and a primordial component indistinguishable from 'solar' ²⁰Ne/²²Ne, again suggesting the presence of an isolated primordial reservoir (Harrison et al., 1999; Trieloff and Kunz, 2005). The heavy noble gases (Ar, Kr, Xe) have an 'atmosphere-like' fingerprint, but not due to recent subduction, suggesting the recycling of heavy noble gases into the deep mantle when the atmospheric noble gas composition was different (Trieloff and Kunz, 2005). This age constraint provides further evidence for the presence of ancient ROC within the Icelandic mantle plume.

Low radiogenic to non-radiogenic Xe (¹²⁹Xe/¹³⁰Xe) from Miðfell glasses is evidence for a relatively undegassed primitive deep-mantle reservoir in the Miðfell source mantle Mukhopadhyay (2012). The low I/Xe in this source (lower than DMM) cannot be explained by the mixing of atmospheric Xe with MORB-type Xe. ¹²⁹I became extinct 100 Ma after the formation of the solar system, suggesting that the low I/Xe source was formed at or before this time due to the differentiation of mantle material in the early Earth.

Primordial undegassed mantle as the source of Miðfell carbon enrichment

There is strong evidence for the presence of both ROC and a primordial undegassed reservoir within the Icelandic mantle plume (Brandon et al., 2007), however there is more uncertainty in attributing high-carbon to just one of those two lithologies in particular. Trace element enriched melts are not carriers of high ³He/⁴He, therefore they are in part the result of undegassed ROC melting (Peate et al., 2010). Miðfell matrix glass has an estimated pyroxenite melt contribution of < 2%, meaning that the more depleted olivine-hosted melt inclusions have an even lower contribution from melting of ROC (Shorttle et al., 2014). These depleted melt inclusions record the high CO₂/ITE ratios, therefore the carbon enrichment is unlikely from ROC.

Miðfell and Borgarhraun melt inclusions show high trace element variability, which in the case of Borgarhraun has been partly attributed to recycled ocean island basalt (Maclennan et al., 2003b; McKenzie et al., 2004). This interpretation suggests that trace element enriched melts in Miðfell are the result of some enriched recycled component. However, the two melt inclusion suites differ when it comes to carbon and ³He/⁴He. Borgarhraun shows a $CO_2/Nb = 391 \pm 16$, similar to values recovered from MORB, and ³He/⁴He ~ 8 (Hauri et al., 2002a; Breddam et al., 2000). Miðfell has elevated CO_2/ITE ratios and ³He/⁴He ~ 17. With the presence of ROC evident in both (although to a lesser extent in Miðfell), then this difference, which couples carbon and He together, must be due to another mantle component, i.e. a primordial undegassed reservoir. All the noble gas isotopes point to this mantle component and it seems logical that if the volatile noble gases can be isolated in the deep mantle, then so too could carbon. Given the assumed primordial nature of this reservoir, then it can be further assumed to have trace element concentrations similar to that of bulk silicate Earth (BSE). However, it is unclear how significant the contribution of BSE carbon is to Miðfell melts and their carbon enrichment.

6.7 Earth's carbon reservoirs

If the carbon enrichment observed in Miðfell melt inclusions is due to an undegassed primordial mantle component residing in the lower mantle with BSE-like chemistry, then Miðfell could contain key information about the carbon content of the bulk mantle. This observation could have implications for Earth's carbon reservoir, which will be discussed in the sections below.

The carbon content of the upper mantle (DMM) has been estimated by several studies of undegassed melt inclusions and MORB glasses from across the MOR system (Saal et al., 2002; Hauri et al., 2002a; Michael and Graham, 2015; Le Voyer et al., 2017). Estimates range from 37–50 ppm carbon, though one sample suite suggested that the enriched MORB source mantle could contain \sim 140 ppm (Cartigny et al., 2008). These estimates indicate heterogeneity within the DMM, which is not surprising as the notion of lithological heterogeneity within the mantle is not new (Stracke, 2012).

Until very recently the only estimates of lower mantle carbon content have come from planetary mass balance calculations. The carbon content of the bulk mantle has been estimated from the global ⁴⁰Ar budget (Marty, 2012). Only a fraction of the ⁴⁰Ar produced by ⁴⁰K decay is in the atmosphere, the rest must be in the solid Earth. Using an estimate of K for the BSE and knowing the decay constant of ⁴⁰K allows the calculation of total ⁴⁰Ar produced by ⁴⁰K decay. The ⁴⁰Ar content of the solid Earth can be calculated as the difference between total decay ⁴⁰Ar and atmospheric ⁴⁰Ar. N/⁴⁰Ar ratios are known from MORB, showing fairly homogeneous values, therefore the N content of the bulk mantle is known. C/N ratios have been measured in MORB, which gives a carbon estimate for the bulk mantle. C/⁴He ratios can be used in a similar way. These calculations predict a carbon concentration of 765 ± 300 ppm for the bulk silicate mantle, which is considerably higher than estimates of DMM.

 CO_2 degassing estimates and probabilistic magma supply rates from Hawaii have been used to estimate the carbon content of the Hawaiian source mantle to be 263 ± 70 ppm (Anderson and Poland, 2017). This value is considerably higher than DMM estimates, but not quite as enriched as bulk mantle estimates (Marty, 2012). The lower mantle is three times larger by mass and twice as large by volume than the upper mantle, therefore the carbon content of this reservoir is likely extremely significant in the carbon budget of the Earth.

6.7.1 Carbon content of Miðfell mantle source lithology

Carbon-rich melt inclusions from Miðfell are likely providing information on the carbon content of the lower mantle and the lithologies it contains. The CO_2/ITE ratios measured in

the Miðfell melt inclusions can be converted into a carbon estimate of the Miðfell mantle source by using mantle ITE concentration estimates. Given the strong evidence that high CO_2/ITE ratios in Miðfell melt inclusions are due to a lower mantle lithology, it is inappropriate to use DMM trace element concentrations for this carbon estimate. The carbon-rich source lithology is inferred to be undegassed primordial mantle, therefore knowledge of the source trace element concentrations must be assumed to convert measured CO_2/ITE ratios to carbon content. Undegassed primordial mantle is assumed to have trace element concentrations similar to BSE, which is justified by the assumed isolated and primordial nature of this reservoir.

The source region beneath Miðfell is very likely a mixture of lithologies, therefore the carbon signal recorded at the surface represents an average of these mantle components. By making the assumption that the CO_2/ITE ratio is solely the result of primordial mantle carbon enrichment, one endmember mantle scenario can be investigated. Due to the trace element enrichment of BSE with respect to DMM, a mixture of these two components would have lower trace element concentrations than the BSE endmember, and therefore contain less carbon than BSE based on the same CO_2/ITE ratio measured in Miðfell melts. Even if the Miðfell carbon is coming from a source with DMM-like trace element concentrations, it will still be enriched in carbon with respect to other segments of the sampled DMM.

The carbon content for the Miðfell source lithology, assuming the carbon contribution was solely from a primordial source, can be estimated as 740 ± 188 ppm (from CO₂/Ba_{smooth}) and 297 \pm 51 ppm (from CO₂/Nb_{smooth}), using BSE Ba and Nb concentrations (Palme and O'Neill, 2014). These values have been calculated using CO₂/ITE ratios from smooth trace element pattern melt inclusions to avoid the influence of crustal assimilation on the carbon estimate. This carbon content estimate is 15 times greater than the carbon concentration estimated for the DMM, however if DMM trace element concentrations are assumed, then is carbon enrichment is only a factor of four greater.

It is interesting to compare the carbon estimate from this BSE endmember scenario to carbon values calculated from planetary mass balance. The Miðfell source mantle value is comparable to estimates of bulk mantle, however the high uncertainty of Miðfell source mantle trace element concentrations suggests that no great significance can be placed on this similarity (Marty, 2012; Hirschmann, 2016). As argued previously, perhaps the Miðfell carbon value is actually more representative of a mixture of lithologies present within the source mantle, i.e. an undegassed primordial reservoir with BSE trace element concentrations mixed with DMM. If so, Miðfell estimates a carbon source value that is an average of lithologies present within the melt region, and therefore more akin to bulk mantle value. However, the method used above to calculate Miðfell source carbon is flawed, as the source trace element

concentration is not known with any certainty. If the source is a mixture of lithologies, then the average source chemistry will not be equivalent to BSE. A different approach is required to establish the ratio of mixed mantle components, so that the carbon content of these endmembers can be established. Such a method could combine measurements from multiple incompatible trace elements to estimate the proportion of BSE material within the source, assuming a simple mixture of BSE and DMM.

6.7.2 Implications of Miðfell carbon enrichment

The endmember scenario for mantle carbon, outlined above, estimates the carbon content of a primordial undegassed reservoir with BSE trace element concentrations. If this value is assumed to be representative of BSE carbon, then a mass balance can be calculated between BSE, DMM and exosphere carbon reservoirs. An estimate for the proportion of mantle considered to be depleted can be calculated by making the assumption that the exosphere carbon reservoir was created by degassing of BSE material down to DMM concentrations. This mass balance uses the Miðfell carbon value as a BSE estimate, the carbon mass of the atmosphere and an estimate of DMM carbon concentration (Le Voyer et al., 2017).

Calculations suggest that $\sim 5\%$ of BSE degassed down to DMM levels to release the mass of the carbon exosphere. This value seems inconsistent with the observation that the DMM (upper mantle) is thought to contribute $\sim 25\%$ of the mantle's total mass, therefore some assumptions made in this mass balance must be incorrect. The DMM proportion of BSE mass could be increased, in line with geophysical and geochemical observations of the DMM, if: (i) BSE carbon has been overestimated, (ii) there is a hidden carbon-rich reservoir unaccounted for in the mass balance, (iii) the exosphere experienced carbon loss after differentiation of the BSE and DMM, (iv) DMM has been underestimated, or (v) the exosphere has been underestimated.

This inconsistency between reservoir size and mass balance calculations can also be seen in planetary mass balance arguments (Marty, 2012). Therefore, it seems unlikely that the Miðfell carbon estimate is representative of the whole lower mantle. However, the observation of high carbon in Miðfell is still an important result because it confirms the presence of carbon-enriched material within the lower mantle. It is conceivable that the CO_2/ITE ratios measured from Miðfell represent a weighted average of several lithologies present in the lower mantle, as geochemical signals pertaining to primordial mantle, ROC, and DMM are expressed in the melt inclusion suite; at least one of which contributes to carbon enrichment.

Large uncertainties on the carbon estimates of Earth's reservoirs mean that a variety of carbon distributions through the Earth are plausible. Planetary mass balance can be used to

narrow this range down, but ultimately better constraints are needed on the carbon content of individual lithologies present in the deep mantle and, more generally, the solid Earth. These constraints could come from better characterisation of processes that replenish the lower mantle, or more direct observations of geochemistry from plume-related volcanism. These constraints can be supported by geophysical and geochemical observations to help pinpoint the potential origin of the signatures seen at the surface (Anderson and Poland, 2017; Mason et al., 2017; Yuan and Romanowicz, 2017).

However, the low solubility of CO_2 within silicate melts means that for the majority of melt suites, degassing obscures the carbon signatures of their source lithologies. This process is particularly significant for carbon-rich lithologies as it causes their carbon signal to remain hidden from direct observation, however their presence, but not as carbon-rich material, could be inferred from trace element chemistry.

6.8 Comparing DMM and lower mantle carbon

The previous section highlighted that the DMM contains less carbon than the lower mantle by up to an order of magnitude. This section discusses the possible causes of such a disparity between mantle carbon reservoirs. DMM is thought to have 37–50 ppm carbon, however the Miðfell estimate of a lower mantle carbon signal is 15 times greater, 740 ± 188 ppm, and calculated from higher CO₂/ITE ratios. If DMM trace element concentrations are used then the carbon enrichment is only a factor of four. Several processes may have acted upon these reservoirs over Earth's history to give the carbon content estimated at the present-day, causing the two distinct carbon reservoirs to have such different concentrations and C/ITE ratios.

6.8.1 Early reservoir differentiation

Xe isotopes measured from Miðfell glass suggest the presence of a primordial undegassed reservoir within the deep mantle, which is inferred to have become isolated during the first 100 Ma after the formation of Earth (Mukhopadhyay, 2012). Other isotopic systems suggest the presence of a primordial component, but do not give any age constraints (Brandon et al., 2007). Early in Earth's history the structure of the Earth was dominated by a magma ocean, which would suggest that the majority of the Earth's volatile budget ought to be in the atmosphere. However, it has been suggested that inefficient melt drainage from the freezing front of Earth's magma ocean caused volatile-rich melt to be trapped within the solidified mantle assemblage (Hier-Majumder and Hirschmann, 2017). Further cooling of these melt

pockets caused the crystallisation of carbon-rich accessory phases. This crystallisation could have been responsible for the initial storage of volatile elements within the mantle, but it is unclear how this process alone could have set up a carbon-rich domain. Instead, it moves the problem of differences between the upper and lower mantle to a process capable of preferentially removing carbon from the upper mantle.

6.8.2 Trace element-carbon recycling

If the upper and lower mantle were not differentiated with respect to carbon during the early Earth, then perhaps later modification processes caused the two to differ. Assuming that the upper and lower mantle were once homogeneous with respect to carbon and ITEs, then to get the two reservoirs estimated at the present-day requires carbon and ITE loss from the upper mantle. However, upper mantle C/ITE ratios suggest that carbon loss was preferential to ITE loss, giving C/ITE ratios lower than the deep mantle.

A potential way of reducing high C/ITE to lower DMM values is the preferential recycling of trace elements over carbon back into the upper mantle. Trace element recycling has previously been suggested for U, with high ²³⁸U/²³⁵U ratios in MORB relative to BSE indicating the presence of subducted U within the DMM (Andersen et al., 2015). Recycled components within the mantle are a familiar concept, however they are often thought to produce trace element enriched melts (Peate et al., 2010), rather than the more depleted compositions associated with MORB (Gale et al., 2013). This problem could potentially be solved if DMM with streaks of recycled oceanic crust underwent a second phase of melting, e.g. early formation of continental crust, reducing the trace element enrichment in the recycled component and further depleting the upper mantle of carbon. However, given that oceanic lithosphere is partly mantle material, then not all recycled lithologies are necessarily enriched in trace elements with respect to the original source lithology.

Modern subduction zones are thought to strongly fractionate large ion lithophile and high field strength elements, due to differences in fluid mobility or the presence of accessory phases (McCulloch and Gamble, 1991; Foley et al., 2000) . Therefore, it would be expected that recycled oceanic crust would preserve this geochemical signature as it mixes into the DMM, causing a greater reduction in C/Nb than C/Ba. Melt inclusions with smooth trace element patterns show the largest differences when compared to CO_2/ITE ratios of MORB for Ba, suggesting that C/Ba ratios have undergone a larger reduction than C/Nb in the DMM (Figure 6.7). Melt inclusions with spiky trace element patterns show the opposite, with the greatest C/ITE reduction from the fluid immobile Nb. However, if recycling is related to metasomatism from subduction-related fluids, then perhaps the reduction in C/Nb.

Preferential recycling of trace elements over carbon into the mantle could modify the C/ITEs of the DMM. However, due to the uncertainties surrounding fluid mobility and carbon flux into the mantle it is unclear if this process acting over billions of years has created the volatile signature of the modern DMM.

6.8.3 Melt depletion processes

The DMM is thought to be the residue from melting of a BSE-like lithology (Workman and Hart, 2005). Observations of undegassed MORB glasses are used to infer that carbon and ITEs have a similar compatibility during the mantle melting process, therefore there is minimal fractionation between these components and the C/ITE of the residue remains invariant. However, experiments on carbonated peridotite suggest that carbon, Nb, and Ba can be fractionated at low extents of melting (Dasgupta and Hirschmann, 2006; Rosenthal et al., 2015). If melt depletion of a primordial BSE source were to create the DMM, then carbon would have to be more incompatible during melting than the ITEs to match observed lower CO_2/ITE ratios. However, experiments suggest that carbon is more incompatible than Nb, but less so than Ba. Production of small fraction melts would reduce the C/Nb ratio of the solid residue, but increase the C/Ba, which is not what is observed when comparing CO_2/ITE ratios between Miðfell and MORB glasses.

6.9 Conclusions

The global dataset of melt inclusions and MORB glasses is dominated by degassing, however undegassed compositions have more likely been preserved in the most depleted melt inclusions and glasses. Three published studies of undegassed melt inclusions and MORB glasses have provided observations of CO_2/ITE ratios, and DMM carbon estimates were calculated from them.

Miðfell melt inclusions preserve the highest CO_2/ITE ratios ever measured in olivinehosted melt inclusions. When compared to the undegassed MORB suites at similar ITE concentrations, Miðfell is seen to be enriched in carbon. Crustal processes can only lead to a decrease in CO_2/ITE ratios, providing assimilated lithologies are not carbonated. Melt inclusions have been identified that have potentially avoided crustal assimilation and they still show elevated CO_2/ITE ratios with respect to MORB glasses. This observation means that the carbon enrichment in Miðfell is not the result of crustal processing, but rather related to enrichment in the mantle. Carbon enrichment is not due to the late interaction of carbonatite melt with depleted silicate melts shallow in the Icelandic melting region. However, the formation of carbonatite and carbonated silicate melts is the likely mode of carbon mobilisation from mantle assemblages.

A small component of recycled oceanic crust is thought to be present within the Miðfell source mantle. It could be the source of carbon enrichment, providing that carbon was able to pass through ancient subduction zones efficiently. However, it is assumed that ROC melting results in trace element enrichment, yet the highest CO_2/ITE ratios are observed in the most depleted Miðfell melt inclusions, suggesting a different origin for the carbon.

Several noble gas isotopic signatures present in Miðfell glass indicate the existence of a primordial undegassed reservoir within the Icelandic source mantle. Due to the coupled high-carbon, high-³He/⁴He, high-²0Ne/²2Ne, and low-¹29Xe/¹30Xe composition of Miðfell, along with the trace element depleted nature of the highest CO₂/ITE melt inclusions, it can be inferred that a primordial undegassed lithology is in part responsible for Miðfell carbon enrichment. Assuming that this primordial reservoir has a trace element content equivalent to BSE, then its carbon content is 740 ± 188 ppm. This value acts as an extreme endmember estimate for the level of carbon enrichment within the Miðfell source region.

Depending on the length scale over which lithologies mixed within the Icelandic mantle plume, this carbon concentration could represent a single carbon-rich reservoir, or it could be a weighted average of several lithologies present within the lower mantle. Regardless, the key observation from the Miðfell melt inclusion suite is that there are carbon-bearing lithologies in the lower mantle that are more carbon-enriched than the DMM.
Chapter 7

Summary

This thesis explores: (i) the reliability of igneous barometers and melt crystallisation models in predicting equilibration pressures and mineral assemblages, (ii) the controls on the extent of trace element variability preserved by olivine-hosted melt inclusions, and (iii) the carbon content of the Icelandic mantle sourced by the Miðfell eruption. These three areas have been investigated through field work in Iceland; analytical work using SIMS, EPMA, and SEM imaging; experimental work conducted in a gas-mixing furnace and piston-cylinder apparatus; and numerical modelling. A diverse set of geochemical measurements have been combined to provide a detailed interpretation of one main eruption, Miðfell, with the Kistufell eruption producing a comparative dataset for assessing lithophile variability within olivine-hosted melt inclusions. Observations from natural samples have been combined with experimental constraints, both conducted in this thesis and previously published, and geophysical observations to help elucidate the fundamental controls on the magmatic system studied. This chapter summarises the principal results of the thesis, placing them in the context of regional and global scales before discussing potential opportunities for further work.

7.1 Experimental petrology and igneous barometry

There is clear evidence from experimental work and thermodynamic modelling that certain igneous phases, such as clinopyroxene, have pressure sensitive compositions (Putirka et al., 1996; Holland and Powell, 1998). This observation is the main basis for a number of igneous barometers, which have used experimental igneous phase compositions for calibration. However, these calibrations are thought to be unreliable due to problems with old experimental procedures and the pressure ranges over which they were conducted (Masotta et al., 2013; Neave and Putirka, 2017). Developments within the last few decades have improved experi-

mental procedures, making their results more reliable and appropriate for the pressure range of interest to igneous petrologists; crustal pressures (Moore et al., 2008). Recent progress has been made with clinopyroxene-liquid barometry, OPAM barometry, and the establishment of multi-reaction average pressure barometry for mafic phases (Neave and Putirka, 2017; Voigt et al., 2017; Ziberna et al., 2017). Improvements to our understanding of post-entrapment melt inclusion behaviour has enabled a more nuanced interpretation of their CO_2 content (Maclennan, 2017). New techniques have allowed more accurate quantification of vapour bubble CO_2 concentrations, to better reconstruct initial CO_2 content and therefore gain more information about the history of these inclusions (Gaetani, Le Roux, Klein; pers. comm.).

Experiments presented in this thesis were originally aimed at producing a full recalibration of the clinopyroxene-liquid barometer, however problems with the results of the newly developed 5 kbar experimental procedure have not allowed for this. Although, they have been useful for testing the capabilities of published barometers and igneous models. The experiments have highlighted the importance of knowing the equilibrium composition of clinopyroxene when using igneous barometers, as disequilibrium compositions can produce wildly inaccurate pressure estimates. Equilibrium clinopyroxene can be identified by comparing observed compositions to expected end-member proportions, and by calculating the Fe-Mg clinopyroxene and liquid partition coefficient, K_D (Putirka, 1999, 2008b; Neave and Putirka, 2017). All tested barometers were able to predict experimental pressure conditions when using equilibrium clinopyroxene compositions. The clinopyroxene-liquid barometer is the easiest to implement as it requires the fewest equilibrium phases, but OPAM barometry also produced good results, particularly when using the χ^2 misfit minimisation method (Yang et al., 1996; Kelley and Barton, 2008; Neave and Putirka, 2017; Voigt et al., 2017). Multi-reaction average pressure barometry requires equilibrium between three crystalline phases, which is potentially difficult to achieve in volcanic rocks (Ziberna et al., 2017).

Results presented in Chapter 4 provide encouragement that igneous barometry is progressing, with the recalibration of old barometers using new experimental data, and the formulation of new barometers. Some of the data from experiments in this thesis has expanded the Icelandic experimental dataset, and could be used in future igneous barometer recalibration, which would allow igneous petrologists to better investigate crustal pressures, and unite petrological and geophysical observations.

7.2 Geochemical heterogeneity within olivine-hosted melt inclusions

Geochemical heterogeneity has been observed in eruptions around Iceland, both on the scale of an individual eruption and across volcanic systems (Maclennan et al., 2003b; Peate et al., 2010; Shorttle and Maclennan, 2011). Different tracers have been used to infer mantle source heterogeneity, with radiogenic isotopes identifying different isotopic reservoirs and lithophile elements highlighting melt enrichment, which could be an indication of source lithological variability. Melt inclusion suites have been used to characterise melt heterogeneity on a single eruption scale, which if the result of source heterogeneity suggests that the mantle tapped by a single eruption has geochemical variability on lengthscales short enough for multiple signals within one eruption, but long enough that heterogeneity is not eliminated immediately by melting. Borgarhraun is one of the most well-characterised of all Icelandic melt inclusion suites, showing high geochemical variability in both lithophile elements and Pb-isotopes (Maclennan et al., 2003b; Maclennan, 2008b).

Several processes are likely responsible for the geochemical variability observed within melt inclusion suites like Borgarhraun. The Miðfell and Kistufell melt inclusion suites show evidence for post-entrapment crystallisation, H_2O diffusive re-equilibration, CO_2 degassing and vapour bubble formation, crustal assimilation of both plagioclase and gabbroic material, and melt mixing (Danyushevsky et al., 2002; Kamenetsky et al., 2002; Gurenko and Sobolev, 2006; Maclennan, 2008a; Bucholz et al., 2013). However, using appropriate geochemical tracers it is possible to reconstruct these processes and assess the variability of melts entering the magmatic systems beneath these eruptions. Melt inclusions provide a snapshot of the melt population just before entrapment, which for Miðfell has high compositional variability, including some undegassed inclusions that have not interacted with crustal material.

Comparisons between multiple melt inclusion suites from across Iceland highlight a potentially interesting relationship between average melt inclusion enrichment and relative variability within the melt inclusion population. Depleted melt inclusion populations show more relative variability than more enriched inclusion suites. A simple model predicting the composition of primary mantle melts suggests that this observation is related to melt mixing within the mantle (Rudge et al., 2013; Jennings et al., 2017). The relationship can be explained by more efficient mixing of low fraction, low volume, enriched melts, which are the first to form by fractional melting of mantle material. At larger melt volumes, with the average melt composition now more depleted due to higher melt fractions, mixing is less efficient, therefore more melt variability is observed. The model presented in Chapter 5 is by

no means perfect, so more rigorous mathematical foundations are required to create a model that adequately links melt mixing to melt fraction and melt variability.

7.3 Deep mantle carbon

Direct observations of the carbon content of Earth's mantle have so far been restricted to estimates of the depleted upper mantle beneath MORs, with measurements coming from suites of undegassed olivine-hosted melt inclusions and MORB glasses (Saal et al., 2002; Michael and Graham, 2015; Le Voyer et al., 2017). These studies used CO_2/Nb ratios to estimate the carbon content of the DMM to be between 37–50 ppm, however no such estimates exist for lower mantle reservoirs. The only predictions of lower mantle carbon concentration have been calculated from planetary mass balance (Marty, 2012).

One melt inclusion suite from Iceland has been identified as containing undegassed melt inclusions, which provide an estimate for mantle carbon similar to predicted DMM values, suggesting that the mantle beneath Borgarhraun is akin to that beneath a normal MOR, i.e. no plume-related carbon signature (Hauri et al., 2002a, 2017). Elsewhere in Iceland, plume-related geochemical signatures are expressed in the form of primordial noble gas isotopes (Harrison et al., 1999; Dixon et al., 2000; Brandon et al., 2007; Füri et al., 2010; Mukhopadhyay, 2012). Miðfell is unique in its well-characterised noble gas isotopic signatures (He, Ne, Xe), which suggest that the Miðfell source mantle contains a primordial undegassed reservoir brought into the melting region by the Icelandic mantle plume. Observations of lithophile and volatile variability in olivine-hosted melt inclusions from Miðfell are presented in this thesis. They suggest that beneath the compositional modifications imposed by crustal processing lies melt heterogeneity inherited from the mantle beneath Miðfell. The primordial noble gases suggest that some of this melt heterogeneity can be attributed to an undegassed primordial reservoir.

The most trace element depleted melt inclusions are shown to preserve undegassed CO_2 concentrations with some of the highest CO_2/ITE ratios recorded in basaltic glass. Discussion of the origin for these high CO_2/ITE ratios and carbon enrichment conclude that an undegassed primordial reservoir from the lower mantle is the most likely source lithology. Assuming that this primordial reservoir has trace element content equivalent to BSE, then its carbon content is 740 \pm 188 ppm. This value matches the bulk mantle estimate calculated from planetary mass balance, and it indicates the presence of a carbon-rich reservoir within the lower mantle (Marty, 2012). However, depending on the lengthscale over which lithologies mixed within the Icelandic mantle plume, this carbon concentration could represent a single carbon-rich reservoir, or it could be a weighted average of several

lithologies present within the lower mantle. Regardless of the origin, the key result is that this observation suggests that there is a carbon-rich lithology of unknown size within the lower mantle, which is more concentrated in carbon than the depleted MORB mantle.

7.4 Further work

7.4.1 Igneous barometry

Igneous barometry could benefit from more experiments conducted at crustal pressures as this would improve the reliability of barometer calibrations at the pressure of interest for igneous petrologists. Further improvements are required for the experimental procedure outlined in Chapter 3 to eliminate Na-gain and Fe-loss in the 5 kbar experimental capsules, and to increase the success rate of holding 5 kbar conditions for the duration of the experimental run. A ridged base plug should help to stabilise the pressure-temperature conditions of the experiment by holding the thermocouple in place, while the use of a double capsule, graphite within noble metal, should prevent Na from diffusing from the pressure medium into the sample.

Expansion of the Icelandic experimental dataset to other bulk compositions is just as important as experimenting upon a range of pressures. The diversity of Icelandic melts, generally distinguished by their enrichment in lithophile elements, requires that igneous barometers are sensitive to a range of compositions, as it is likely that bulk composition has a control over crystallisation paths and mineral assemblage stability (Neave et al., 2017). Ultimately it would be useful to have a set of igneous barometers that are sensitive to crustal pressures and a full range of bulk compositions, allowing for reliable pressure estimation from any igneous assemblage.

The development of multi-reaction average pressure barometry to cover mafic phase assemblages has only been possible due to recent improvements in mineral activity-composition models (Green et al., 2016; Ziberna et al., 2017). The avP program was developed to target gabbroic assemblages, however the barometer could be further improved by building in a melt model. This addition would allow for more endmember reactions to constrain equilibrium pressure, and expand the avP program to volcanic compositions where melt is the key component capable of maintaining equilibrium between the phenocryst phases. A new melt model applicable to mafic compositions, and those that are more SiO₂ rich, would allow THERMOCALC to calculate pressure-temperature pseudosections on mafic igneous compositions and thermodynamically constrain equilibrium assemblages.

7.4.2 Melt inclusion heterogeneity

Careful measurement of more Icelandic melt inclusion suites, and more individual melt inclusion measurements from each suite are required to improve the statistical significance of the observed trend between melt inclusion variability and enrichment. Once this has been achieved then the development of a more thorough mantle melting-melt mixing model is needed to explain the relationship between enrichment, melt fraction, and extent of mixing prior to melt inclusion entrapment.

7.4.3 Mantle volatile reservoirs

Several studies have considered more than just carbon when assessing the volatile budget of the solid Earth. If Miðfell is truly tapping a primordial undegassed reservoir then it would be interesting to see if melt inclusions hold a nitrogen signature related to the lower mantle and the early Earth. Measurements of mantle derived nitrogen have previously been made on fluid inclusions within olivine, and inclusions within diamond. It is possible to measure nitrogen by SIMS, so melt inclusions could be analysed for nitrogen. Miðfell could provide a crucial extra constraint on the nitrogen content of the solid Earth, as currently $\geq 90\%$ of the mantle nitrogen database comes from diamonds (Mikhail and Howell, 2016). It could provide a useful comparison to nitrogen budgets derived from planetary mass balance (Andersen et al., 1995; Regier et al., 2016).

Miðfell is now a very well-characterised eruption, however it would be intriguing to see if the variability shown by the lithophile elements is also reflected in some way by isotopic systems, such as Rb-Sr or U-Pb. Isotopes have been measured in Icelandic melt inclusions before, so it is possible to conduct reliable Pb analyses on them (Maclennan, 2008b). Measurement of isotopes could enable Miðfell's carbon enrichment to be attributed to a particular mantle lithology with more certainty.

Looking on a global scale, it would be useful to identify the carbon content of deep mantle reservoirs from other eruptions. High forsterite content olivines, glassy carrier liquid and melt inclusions, depleted trace element concentrations, and high primordial noble gas isotopes are the key characteristics of Miðfell that enable its carbon enrichment to be attributed to a lower mantle lithology. Observations of CO_2 flux ought to be combined with these characteristics to help identify other eruptions that could unlock the secrets of the deep Earth (Anderson and Poland, 2017). Iceland is an incredible natural laboratory for exploring igneous processes and more general solid Earth systematics, but perhaps there are just as remarkable locations and melt inclusion suites elsewhere around the world that could advance our knowledge of carbon cycling.

References

- Ágústsdóttir, T., Woods, J., Greenfield, T., Green, R. G., White, R. S., Winder, T., Brandsdóttir, B., Steinthórsson, S., and Soosalu, H. (2016). Strike-slip faulting during the 2014 Bárðarbunga-Holuhraun dike intrusion, central Iceland. *Geophysical Research Letters*, 43(4):1495–1503.
- Andersen, M. B., Elliott, T., Freymuth, H., Sims, K. W. W., Niu, Y., and Kelley, K. A. (2015). The terrestrial uranium isotope cycle. *Nature*, 517:356–359.
- Andersen, T., Burke, E. A., and Neumann, E. R. (1995). Nitrogen-rich fluid in the upper mantle: fluid inclusions in spinel dunite from Lanzarote, Canary Islands. *Contributions to Mineralogy and Petrology*, 120(1):20–28.
- Anderson, A. T. and Brown, G. G. (1993). CO₂ contents and formation pressures of some Kilauean melt inclusions. *American Mineralogist*, 78(7-8):794–803.
- Anderson, A. T., Newman, S., Williams, S. N., Druitt, T. H., Skirius, C., and Stolper, E. (1989). H₂O, CO₂, Cl, and gas in Plinian and ash-flow Bishop rhyolite. *Geology*, 17(3):221–225.
- Anderson, K. R. and Poland, M. P. (2017). Abundant carbon in the mantle beneath Hawaii. *Nature Geoscience*, 10(9):704–708.
- Annen, C., Blundy, J. D., and Sparks, R. S. J. (2006). The genesis of intermediate and silicic magmas in deep crustal hot zones. *Journal of Petrology*, 47(3):505–539.
- Asimow, P. D. and Ghiorso, M. S. (1998). Algorithmic modifications extending MELTS to calculate subsolidus phase relations. *American Mineralogist*, 83:1127–1131.
- Aster, E. M., Wallace, P. J., Moore, L. R., Watkins, J., Gazel, E., and Bodnar, R. J. (2016). Reconstructing CO2 concentrations in basaltic melt inclusions using Raman analysis of vapor bubbles. *Journal of Volcanology and Geothermal Research*, 323:148–162.
- Baker, D. R. (2004). Piston-cylinder calibration at 400 to 500 MPa: A comparison of using water solubility in albite melt and NaCl melting. *American Mineralogist*, 89(10):1553– 1556.
- Bédard, J. H. (1993). Oceanic crust as a reactive filter: synkinematic intrusion, hybridization, and assimilation in an ophiolitic magma chamber, western Newfoundland. *Geology*, 21(1):77–80.
- Bédard, J. H. (2005). Partitioning coefficients between olivine and silicate melts. *Lithos*, 83:394–419.
- Berndt, J., Liebske, C., Holtz, F., Freise, M., Nowak, M., Ziegenbein, D., Hurkuck, W., and Koepke, J. (2002). A combined rapid-quench and H₂-membrane setup for internally heated pressure vessels: Description and application for water solubility in basaltic melts. *American Mineralogist*, 87(11-12):1717–1726.

- Bindeman, I. N., Davis, A. M., and Drake, M. J. (1998). Ion Microprobe Study of Plagioclase-Basalt Partition Experiments at Natural Concentration Levels of Trace Elements. *Geochimica et Cosmochimica Acta*, 62(7):1175–1193.
- Blundy, J. D., Falloon, T. J., Wood, B. J., and Dalton, J. A. (1995). Sodium partitioning between clinopyroxene and silicate melts. *Journal of Geophysical Research*, 100(B8):15501– 15515.
- Boettcher, A. L., Windom, K. E., Bohlen, S. R., and Luth, R. W. (1981). Low-friction, anhydrous, low- to high-temperature furnace sample assembly for piston-cylinder apparatus. *Review of Scientific Instruments*, 52(12):1903–1904.
- Bohlen, S. R. (1984). Equilibria for precise pressure calibration and a frictionless furnace assembly for the piston-cylinder apparatus. *Neues Jahrbuch fur Mineralogie, Monatshefte*, 404:404–412.
- Bourgeois, O., Dauteuil, O., and Van Vliet-Lanoë, B. (1998). Pleistocene subglacial volcanism in Iceland: Tectonic implications. *Earth and Planetary Science Letters*, 164(1-2):165– 178.
- Bowen, N. (1928). The Evolution of the Igneous Rocks. Princeton University Press.
- Bowen, N. L. (1915). The later stages of evolution of the igneous rocks. *The Journal of Geology*, 121:1–91.
- Boyd, F. R. and England, J. L. (1960). Apparatus for phase-equilibrium measurements at pressures up to 50 kilobars and temperatures up to 1750 °C. *Journal of Geophysical Research*, 65(2):741.
- Brandon, A. D., Graham, D. W., Waight, T., and Gautason, B. (2007). ¹⁸⁶Os and ¹⁸⁷Os enrichments and high ³He/⁴He sources in the Earth's mantle: Evidence from Icelandic picrites. *Geochimica et Cosmochimica Acta*, 71:4570–4591.
- Brandsdóttir, B., Menke, W., Einarsson, P., White, R. S., and Staples, R. K. (1997). Färoe-Iceland Ridge Experiment 2. Crustal structure of the Krafla central volcano. *Journal of Geophysical Research: Solid Earth*, 102(B4):7867–7886.
- Breddam, K. (2002). Kistufell: Primitive melt from the Iceland mantle plume. *Journal of Petrology*, 43(2):345–373.
- Breddam, K., Kurz, M. D., and Storey, M. (2000). Mapping out the conduit of the Iceland mantle plume with helium isotopes. *Earth and Planetary Science Letters*, 176:45–55.
- Brey, G. P., Bulatov, V. K., Girnis, A. V., and Lahaye, Y. (2008). Experimental Melting of Carbonated Peridotite at 6-10 GPa. *Journal of Petrology*, 49(4):797–821.
- Brounce, M., Feineman, M., LaFemina, P., and Gurenko, A. (2012). Insights into crustal assimilation by Icelandic basalts from boron isotopes in melt inclusions from the 1783–1784 Lakagígar eruption. *Geochimica et Cosmochimica Acta*, 94:164–180.
- Bryndzia, L. T. and Wood, B. J. (1990). Oxygen thermobarometry of abyssal spinel peridotites: the redox state and C-O-H volatile composition of the Earth's sub-oceanic upper mantle. *American Journal of Science*, 290:1903–1116.
- Bucholz, C. E., Gaetani, G. A., Behn, M. D., and Shimizu, N. (2013). Post-entrapment modification of volatiles and oxygen fugacity in olivine-hosted melt inclusions. *Earth and Planetary Science Letters*, 374:145–155.

- Buck, W. R. and Su, W. (1989). Focused mantle upwelling below mid-ocean ridges due to feedback between viscosity and melting. *Geophysical Research Letters*, 16(7):641–644.
- Budd, D. A., Troll, V. R., Dahren, B., and Burchardt, S. (2016). Persistent multi-tiered magma plumbing beneath Katla volcano, Iceland. *Geochemistry, Geophysics, Geosystems*, 17(3):966–980.
- Burnard, P., Reisberg, L., and Colin, A. (2014). An observed link between lithophile compositions and degassing of volatiles (He, Ar, CO₂) in MORBs with implications for Re volatility and the mantle C/Nb ratio. *Earth and Planetary Science Letters*, 395:159–167.
- Cabral, R. A., Jackson, M. G., Koga, K. T., Rose-Koga, E. F., Hauri, E. H., Whitehouse, M. J., Price, A. A., Day, J. M. D., Shimizu, N., and Kelley, K. A. (2014). Volatile cycling of H₂O, CO₂, F, and Cl in the HIMU mantle: A new window provided by melt inclusions from oceanic hot spot lavas at Mangaia, Cook Islands. *Geochemistry, Geophysics, Geosystems*, 15(11):4445–4467.
- Carmichael, I. S. E., Nicholls, J., and Smith, A. L. (1970). Silica activity in igneous rocks. *The American Mineralogist*, 55:246–263.
- Cartigny, P., Pineau, F., Aubaud, C., and Javoy, M. (2008). Towards a consistent mantle carbon flux estimate: Insights from volatile systematics (H₂O/Ce, δ D, CO₂/Nb) in the North Atlantic mantle (14° N and 34° N). *Earth and Planetary Science Letters*, 265(3-4):672–685.
- Cerantola, V., Bykova, E., Kupenko, I., Merlini, M., Ismailova, L., McCammon, C., Bykov, M., Chumakov, A. I., Petitgirard, S., Kantor, I., Svitlyk, V., Jacobs, J., Hanfland, M., Mezouar, M., Prescher, C., Rüffer, R., Prakapenka, V. B., and Dubrovinsky, L. (2017). Stability of iron-bearing carbonates in the deep Earth's interior. *Nature Communications*, 8(May):15960.
- Chauvel, C. and Hémond, C. (2000). Melting of a complete section of recycled oceanic crust: Trace element and Pb isotopic evidence from Iceland. *Geochemistry, Geophysics, Geosystems*, 1(2).
- Chavrit, D., Humler, E., and Grasset, O. (2014). Mapping modern CO₂ fluxes and mantle carbon content all along the mid-ocean ridge system. *Earth and Planetary Science Letters*, 387:229–239.
- Chen, Y., Provost, A., Schiano, P., and Cluzel, N. (2013). Magma ascent rate and initial water concentration inferred from diffusive water loss from olivine-hosted melt inclusions. *Contributions to Mineralogy and Petrology*, 165(3):525–541.
- Clift, P. D. (2017). A revised budget for Cenozoic sedimentary carbon subduction. *Reviews* of Geophysics, 55(1):97–125.
- Colman, A., Sinton, J. M., and Wanless, V. D. (2015). Constraints from melt inclusions on depths of magma residence at intermediate magma supply along the Galápagos Spreading Centre. *Earth and Planetary Science Letters*, 412:122–131.
- Coogan, L. (2014). 4.14 the lower oceanic crust. In Holland, H. D. and Turekian, K. K., editors, *Treatise on Geochemistry (Second Edition)*, pages 497–541. Elsevier, second edition edition.
- Cottrell, E., Walter, M. J., and Walker, D. (2009). Metal silicate partitioning of tungsten at high pressure and temperature : Implications for equilibrium core formation in Earth. *Earth and Planetary Science Letters*, 281(3-4):275–287.

- Dalton, J. A. and Presnall, D. C. (1998). The Continuum of Primary Carbonatitic-Kimberlitic Melt Compositions in Equilibrium with Lherzolite: Data from the System CaO-MgO-Al₂O₃-SiO₂-CO₂ at 6 GPa. *Journal of Petrology*, 39(11-12):1953–1964.
- Danyushevsky, L. V. (2001). The effect of small amounts of H₂O on crystallisation of mid-ocean ridge and backarc basin magmas. *Journal of Volcanology and Geothermal Research*, 110:265–280.
- Danyushevsky, L. V., Della-Pasqua, F. N., and Sokolov, S. (2000). Re-equilibration of melt inclusions trapped by magnesian olivine phenocrysts from subduction-related magmas: petrological implications. *Contributions to Mineralogy and Petrology*, 138(1):68–83.
- Danyushevsky, L. V., Perfit, M. R., Eggins, S. M., and Falloon, T. J. (2003). Crustal origin for coupled 'ultra-depleted' and 'plagioclase' signatures in MORB olivine-hosted melt inclusions: evidence from the Siqueiros Transform Fault, East Pacific Rise. *Contributions* to Mineralogy and Petrology, 144(5):619–637.
- Danyushevsky, L. V. and Plechov, P. (2011). Petrolog3: integrated software for modeling crystallization processes. *Geochemistry Geophysics Geosystems*, 12(7).
- Danyushevsky, L. V., Sobolev, A. V., and Dmitriev, L. V. (1996). Mineralogy Petrology Estimation of the pressure of crystallization and H20 content of MORB and BABB glasses : calibration of an empirical technique. *Mineralogy and Petrology*, 57:185–204.
- Danyushevsky, L. V., Sokolov, S., and Falloon, T. J. (2002). Melt Inclusions in Olivine Phenocrysts : Using Diffusive Re-equilibration to Determine the Cooling History of a Crystal , with Implications for the Origin of Olivine-phyric Volcanic Rocks. *Journal of Petrology*, 43(9):1651–1671.
- Darbyshire, F. A., White, R. S., and Priestley, K. F. (2000). Structure of the crust and uppermost mantle of Iceland from a combined seismic and gravity study. *Earth and Planetary Science Letters*, 181:409–428.
- Dasgupta, R. and Hirschmann, M. M. (2006). Melting in the Earth's deep upper mantle caused by carbon dioxide. *Nature*, 440(7084):659–662.
- Dasgupta, R. and Hirschmann, M. M. (2010). The deep carbon cycle and melting in Earth's interior. *Earth and Planetary Science Letters*, 298(1-2):1–13.
- Dasgupta, R., Hirschmann, M. M., and Dellas, N. (2005). The effect of bulk composition on the solidus of carbonated eclogite from partial melting experiments at 3 GPa. *Contributions* to *Mineralogy and Petrology*, 149:288–305.
- Dasgupta, R., Hirschmann, M. M., McDonough, W. F., Spiegelman, M., and Withers, A. C. (2009). Trace element partitioning between garnet lherzolite and carbonatite at 6.6 and 8.6 GPa with applications to the geochemistry of the mantle and of mantle-derived melts. *Chemical Geology*, 262(1-2):57–77.
- Dasgupta, R., Hirschmann, M. M., and Smith, N. D. (2007a). Partial melting experiments of peridotite + CO₂ at 3 GPa and genesis of alkalic ocean island basalts. *Journal of Petrology*, 48(11):2093–2124.
- Dasgupta, R., Hirschmann, M. M., and Smith, N. D. (2007b). Water follows carbon: CO₂ incites deep silicate melting and dehydration beneath mid-ocean ridges. *Geology*, 35(2):135–138.
- Dasgupta, R., Hirschmann, M. M., and Stalker, K. (2006). Immiscible transition form carbonate-rich to silicate-rich melts in the 3 GPa melting interval of eclogite + CO₂ and genesis of silica-undersaturated ocean island lavas. *Journal of Petrology*, 57(4):647–671.

- Dasgupta, R., Hirschmann, M. M., and Withers, A. C. (2004). Deep global cycling of carbon constrained by the solidus of anhydrous carbonated eclogite under upper mantle conditions. *Earth and Planetary Science Letters*, 227:73–85.
- Dasgupta, R., Jackson, M. G., and Lee, C.-T. A. (2010). Major element chemistry of ocean island basalts Conditions of mantle melt heterogeneity of mantle source. *Earth and Planetary Science Letters*, 289:377–392.
- Dasgupta, R., Mallik, A., Tsuno, K., Withers, A. C., Hirth, G., and Hirschmann, M. M. (2013). Carbon-dioxide-rich silicate melt in the Earth's upper mantle. *Nature*, 493(7431):211–5.
- Debaille, V., Tronnes, R. G., Brandon, A. D., Waight, T. E., Graham, D. W., and Lee, C.-T. A. (2009). Primitive off-rift basalts from Iceland and Jan Mayen: Os-isotopic evidence for a mantle source containing enriched subcontinental lithosphere. *Geochimica et Cosmochimica Acta*, 73:3423–3449.
- DeMets, C., Gordon, R. G., Argus, D. F., and Stein, S. (1994). Effect of recent revisions to the geomagnetic reversal time scale on estimates of current plate motions. *Geophysical Research Letters*, 21(20):2191–2194.
- Des Marais, D. J. (2001). Isotopic Evolution of the Biogeochemical Carbon Cycle During the Precambrian. *Reviews in Mineralogy and Geochemistry*, 43(1):555–578.
- Di Muro, A., Metrich, N., Vergani, D., Rosi, M., Armienti, P., Fougeroux, T., Deloule, E., Arienzo, I., and Civetta, L. (2014). The Shallow Plumbing System of Piton de la Fournaise Volcano (La Reunion Island, Indian Ocean) Revealed by the Major 2007 Caldera-Forming Eruption. *Journal of Petrology*, 55(7):1287–1315.
- Dixon, E. T., Honda, M., McDougall, I., Campbell, I. H., and Sigurdsson, I. (2000). Preservation of near-solar neon isotopic ratios in Icelandic basalts. *Earth and Planetary Science Letters*, 180:309–324.
- Dixon, J. E., Clague, D. A., and Stolper, E. M. (1991). Degassing history of water, sulfur, and carbon in submarine lavas from Kilauea volcano, Hawaii. *Journal of Geology*, 99(3):371–394.
- Dixon, J. E. and Stolper, E. M. (1995). An experimental study of water and carbon dioxide solubilities in mid-ocean ridge basaltic liquids.Part II: Applications to degassing. *Journal of Petrology*, 36(6):1633–1646.
- Eason, D. E. and Sinton, J. M. (2009). Lava shields and fissure eruptions of the Western Volcanic Zone, Iceland: Evidence for magma chambers and crustal interaction. *Journal of Volcanology and Geothermal Research*, 186:331–348.
- Edmonds, M. (2008). New geochemical insights into volcanic degassing. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 366(1885):4559–4579.
- Fitton, G. J., Saunders, A. D., Norry, M. J., Hardarson, B. S., and Taylor, R. N. (1997). Thermal and chemical structure of the Iceland plume. *Earth and Planetary Science Letters*, 153:197–208.
- Foley, S., Yaxley, G., Rosenthal, A., Buhre, S., Kiseeva, E., Rapp, R., and Jacob, D. (2009). The composition of near-solidus melts of peridotite in the presence of CO₂ and H₂O between 40 and 60 kbar. *Lithos*, 112:274–283.
- Foley, S. F., Barth, M. G., and Jenner, G. A. (2000). Rutile/melt partition coefficients for trace elements and an assessment of the influence of rutile on the trace element characteristics of subduction zone magmas. *Geochimica et Cosmochimica Acta*, 64(5):933–938.

- Ford, C. E., Russell, D. G., Groven, J. A., and Fisk, M. A. (1983). Distribution coefficients of Mg²⁺, Fe²⁺, Ca²⁺ and Mn²⁺ between olivine and melt. *Journal of Petrology*, 24(3):256–265.
- Frost, D. J., Liebske, C., Langenhorst, F., McCammon, C. A., Tronnes, R. G., and Rubie, D. C. (2004). Experimental evidence for the existence of iron-rich metal in the Earth's lower mantle. *Nature*, 428:409–412.
- Frost, D. J. and McCammon, C. A. (2008). The redox state of the Earth. *Annual Reviews of Earth and Planetary Sciences*, 36:389–420.
- Füri, E., Hilton, D. R., Halldórsson, S. A., Barry, P. H., Hahm, D., Fischer, T. P., and Grönvold, K. (2010). Apparent decoupling of the He and Ne isotope systematics of the Icelandic mantle: The role of He depletion, melt mixing, degassing fractionation and air interaction. *Geochimica et Cosmochimica Acta*, 74:3307–3332.
- Gaetani, G. a. and Grove, T. L. (1998). The influence of water on melting of mantle peridotite. *Contributions to Mineralogy and Petrology*, 131(4):323–346.
- Gaetani, G. a., O'Leary, J. a., Shimizu, N., Bucholz, C. E., and Newville, M. (2012). Rapid reequilibration of H₂O and oxygen fugacity in olivine-hosted melt inclusions. *Geology*, 40(10):915–918.
- Gale, A., Dalton, C. A., Langmuir, C. H., Su, Y., and Schilling, J.-G. (2013). The mean composition of ocean ridge basalts. *Geochemistry, Geophysics, Geosystems*, 14(3).
- Geiger, H., Mattsson, T., Deegan, F. M., Troll, V. R., Burchardt, S., Gudmundsson, O., Tryggvason, A., Krumbholz, M., and Harris, C. (2016). Magma plumbing for the 2014– 2015 Holuhraun eruption, Iceland. *Geochemistry, Geophysics, Geosystems*, 17(8):2953– 2968.
- Ghiorso, M. S. and Sack, R. O. (1995). Chemical mass transfer in magmatic processes IV. A revised and internally consistent thermodynamic model for the interpolation and extrapolation of liquid-solid equilibria in magmatic systems at elevated temperatures and pressures. *Contributions to Mineralogy and Petrology*, 119:197–212.
- Gill, R. (2014). Modern Analytical Geochemistry: An Introduction to Quantitative Chemical Analysis Techniques for Earth, Environmental and Materials Scientists. Longman Geochemistry Series. Taylor & Francis.
- Green, D. H. (1970). A review of experimental evidence on the origin of basaltic and nephelinitic magmas. *Physics of the Earth and Planetary Interiors*, 3(C):221–235.
- Green, E. C., White, R. W., Diener, J. F., Powell, R., Holland, T. J., and Palin, R. M. (2016). Activity–composition relations for the calculation of partial melting equilibria in metabasic rocks. *Journal of Metamorphic Geology*, 34(9):845–869.
- Greenfield, T. and White, R. S. (2015). Building icelandic igneous crust by repeated melt injections. *Journal of Geophysical Research: Solid Earth*, 120(11):7771–7788.
- Greenfield, T., White, R. S., and Roecker, S. (2016). The magmatic plumbing system of the Askja central volcano, Iceland, as imaged by seismic tomography. *Journal of Geophysical Research: Solid Earth*, 121(10):7211–7229.
- Grove, T. L., Kinzler, R. J., and Bryan, W. B. (1992). Fractionation of Mid-Ocean Ridge Basalt (MORB). *Geophysical Monograph*, 71:281–310.
- Gudfinnsson, G. H. (2005). Continuous Gradations among Primary Carbonatitic, Kimberlitic, Melilititic, Basaltic, Picritic, and Komatiitic Melts in Equilibrium with Garnet Lherzolite at 3-8 GPa. *Journal of Petrology*, 46(8):1645–1659.

- Gudmundsson, A. (1987). Lateral magma flow, caldera collapse, and a mechanism of large eruptions in Iceland. *Journal of Volcanology and Geothermal Research*, 34(1-2):65–78.
- Gudmundsson, A. (2000). Dynamics of volcanic systems in iceland: Example of tectonism and volcanism at juxtaposed hot spot and mid-ocean ridge systems. *Annual Review of Earth and Planetary Sciences*, 28(1):107–140.
- Gurenko, A. A., Bindeman, I. N., and Sigurdsson, I. A. (2015). To the origin of Icelandic rhyolites: insights from partially melted leucocratic xenoliths. *Contributions to Mineralogy and Petrology*, 169(5):1–21.
- Gurenko, A. A. and Chaussidon, M. (1995). Enriched and primitive melts included in olivine from Icelandic tholeiites: origin by continuous melting of a single mantle column. *Geochimica et Cosmochimica Acta*, 59:2905–2917.
- Gurenko, A. A. and Sobolev, A. V. (2006). Crust-primitive magma interaction beneath neovolcanic rift zone of Iceland recorded in gabbro xenoliths from Midfell, SW Iceland. *Contributions to Mineralogy and Petrology*, 151(5):495–520.
- Halldórsson, S. A., Barnes, J. D., Stefánsson, A., Hilton, D. R., Hauri, E. H., and Marshall, E. W. (2016). Subducted lithosphere controls halogen enrichments in the Iceland mantle plume source. *Geology*, 44(8):679–682.
- Hammouda, T. (2003). High-pressure melting of carbonated eclogite and experimental constraints on carbon recycling and storage in the mantle. *Earth and Planetary Science Letters*, 214:357–368.
- Hammouda, T. and Laporte, D. (2000). Ultrafast mantle impregnation by carbonatite melts. *Geology*, 28(3):283.
- Hansteen, T. H. (1991). Multi-stage evolution of the picritic Mælifell rocks, SW Iceland: constraints from mineralogy and inclusions of glass and fluid in olivine. *Contributions to Mineralogy and Petrology*, 109(2):225–239.
- Harrison, D., Burnard, P., and Turner, G. (1999). Noble gas behaviour and composition in the mantle: constraints from the Iceland Plume. *Earth and Planetary Science Letters*, 171(2):199–207.
- Hartley, M. E., Maclennan, J., Edmonds, M., and Thordarson, T. (2014). Reconstructing the deep CO₂ degassing behaviour of large basaltic fissure eruptions. *Earth and Planetary Science Letters*, 393:120–131.
- Hartley, M. E., Neave, D. A., Maclennan, J., Edmonds, M., and Thordarson, T. (2015). Diffusive over-hydration of olivine-hosted melt inclusions. *Earth and Planetary Science Letters*, 425:168–178.
- Hartley, M. E., Shorttle, O., Maclennan, J., Moussallam, Y., and Edmonds, M. (2017). Olivine-hosted melt inclusions as an archive of redox heterogeneity in magmatic systems. *Earth and Planetary Science Letters*, 479:192–205.
- Hartley, M. E. and Thordarson, T. (2013). The 1874-1876 volcano-tectonic episode at Askja, North Iceland: Lateral flow revisited. *Geochemistry, Geophysics, Geosystems*, 14(7):2286–2309.
- Hauri, E., Gronvöld, K., Oskarsson, N., and McKenzie, D. (2002a). Abundance of carbon in the Icelandic mantle: constraints from melt inclusions. In *AGU Spring Meeting Abstracts*, volume 1, page 03.
- Hauri, E., Wang, J., Dixon, J. E., King, P. L., Mandeville, C., and Newman, S. (2002b). SIMS analysis of volatiles in silicate glasses. *Chemical Geology*, 183(1-4):99–114.

- Hauri, E. H., Maclennan, J., McKenzie, D., Gronvold, K., Oskarsson, N., and Shimizu, N. (2017). CO₂ content beneath northern Iceland and the variability of mantle carbon. *Geology*, (1):1–4.
- Hayes, J. M. and Waldbauer, J. R. (2006). The carbon cycle and associated redox processes through time. *Philosophical Transactions of the Royal Society B*, 361:931–950.
- Hays, J. F. (1966). Lime-alumina-silica. *Carnegie Institution of Washington Yearbook*, 65:234–239.
- Head, E. M., Shaw, A. M., Wallace, P. J., Sims, K. W. W., and Carn, S. A. (2011). Insight into volatile behavior at Nyamuragira volcano (D.R. Congo, Africa) through olivine-hosted melt inclusions. *Geochemistry, Geophysics, Geosystems*, 12(10):1–22.
- Helo, C., Longpré, M.-A., Shimizu, N., Clague, D. A., and Stix, J. (2011). Explosive eruptions at mid-ocean ridges driven by CO2-rich magmas. *Nature Geoscience*, 4(4):260–263.
- Hémond, C., Arndt, N., Litchenstein, U., Hofmann, A., Oskarsson, N., and Steinthorsson, S. (1993). The heterogeneous Iceland plume: Nd-Sr-O isotopes and trace element constraints. *Journal of Geophysical Research*, 98:15833–15850.
- Henstock, T. J., Woods, A. W., and White, R. S. (1993). The accretion of oceanic crust by episodic sill intrusion. *Journal of Geophysical Research*, 98:4143.
- Hier-Majumder, S. and Hirschmann, M. M. (2017). The origin of volatiles in the Earth's mantle. *Geochemistry, Geophysics, Geosystems*, 18(8):3078–3092.
- Hirose, K. (1997). Partial melt compositions of carbonated peridotite at 3 GPa and role of CO₂ in alkali-basalt magma generation. *Geophysical Research Letters*, 24(22):2837–2840.
- Hirschmann, M. M. (2016). Constraints on the early delivery and fractionation of Earth's major volatiles from C/H, C/N, and C/S ratios. *American Mineralogist*, 101(3):540–553.
- Hirschmann, M. M. and Dasgupta, R. (2009). The H/C ratios of Earth's near-surface and deep reservoirs, and consequences for deep Earth volatile cycles. *Chemical Geology*, 262:4–15.
- Hirschmann, M. M., Ghiorso, M. S., and Stolper, M. (1999). Calculation of Peridotite Partial Melting from Thermodynamic Models of Minerals and Melts. II. Isobaric Variations in Melts near the Solidus and owing to Variable Source Composition. *Journal of Petrology*, 40(2):297–313.
- Hoernle, K., Tilton, G., Le Bas, M. J., Duggen, S., and Garbe-Schönberg, D. (2002). Geochemistry of oceanic carbonatites compared with continental carbonatites: mantle recycling of oceanic crustal carbonate. *Contributions to Mineralogy and Petrology*, 142(5):520–542.
- Hofmann, A. W. and White, W. M. (1982). Mantle plumes from ancient oceanic crust. *Earth* and *Planetary Science Letters*, 57:421–436.
- Holland, T. and Blundy, J. (1994). Non-ideal interactions in calcic amphiboles and their bearing on amphibole-plagioclase thermometry. *Contributions to Mineralogy and Petrology*, 116(4):433–447.
- Holland, T. J. B., Hudson, N. F. C., Powell, R., and Harte, B. (2013). New thermodynamic models and calculated phase equilibria in NCFMAS for basic and ultrabasic compositions through the transition zone into the uppermost lower mantle. *Journal of Petrology*, 54(9):1901–1920.

- Holland, T. J. B. and Powell, R. (1998). An internally consistent thermodynamic dataset for phases of petrological interest. *Journal of Metamorphic Geology*, 16:309–343.
- Holland, T. J. B. and Powell, R. (2011). An improved and extended internally consistent thermodynamic dataset for phases of petrological interest, involving a new equation of state for solids. *Journal of Metamorphic Geology*, 29:333–383.
- Holness, M. B. (2006). Melt–solid dihedral angles of common minerals in natural rocks. *Journal of Petrology*, 47(4):791–800.
- Holtz, F. (2001). The water-undersaturated and dry Qz-Ab-Or system revisited. Experimental results at very low water activities and geological implications. *Contributions to Mineralogy and Petrology*, 141(3):347–357.
- Hooft, E. E. E., Brandsdóttir, B., Mjelde, R., Shimamura, H., and Murai, Y. (2006). Asymmetric plume-ridge interaction around Iceland: The Kolbeinsey ridge Iceland seismic experiment. *Geochemistry, Geophysics, Geosystems*, 7(5).
- Hudgins, T. R., Mukasa, S. B., Simon, A. C., Moore, G., and Barifaijo, E. (2015). Melt inclusion evidence for CO₂-rich melts beneath the western branch of the East African Rift: implications for long-term storage of volatiles in the deep lithospheric mantle. *Contributions to Mineralogy and Petrology*, 169(5):46.
- Hudon, P., Baker, D. R., and Toft, P. B. (1994). A high-temperature assembly for 1.91-cm (3/4 -in.) piston-cylinder apparatus. *American Mineralogist*, 79(1-2):145–147.
- Hudson, T. S., White, R. S., Greenfield, T., Ágústsdóttir, T., Brisbourne, A., and Green, R. G. (2017). Deep crustal melt plumbing of Bárðarbunga volcano, Iceland. *Geophysical Research Letters*, 44(17):8785–8794.
- Hunter, R. H. and McKenzie, D. (1989). The equilibrium geometry of carbonate melts in rocks of mantle composition. *Earth and Planetary Science Letters*, 92(3-4):347–356.
- Iacovino, K., Moore, G., Roggensack, K., Oppenheimer, C., and Kyle, P. (2013). H₂O-CO₂ solubility in mafic alkaline magma: Applications to volatile sources and degassing behavior at Erebus volcano, Antarctica. *Contributions to Mineralogy and Petrology*, 166(3):845–860.
- Jakobsson, S. (1979). *Petrology of recent basalts of the Eastern Volcanic Zone, Iceland.* Contributions to geology. Icelandic Museum of Natural History.
- Jakobsson, S., Blundy, J., and Moore, G. (2014). Oxygen fugacity control in piston-cylinder experiments: a re-evaluation. *Contributions to Mineralogy and Petrology*, 167(6):1007.
- Jakobsson, S. P. and Gudmundsson, M. T. (2008). Subglacial and intraglacial volcanic formations in Iceland. *Jökull*, (58):182–186.
- Jarosewich, E., Nelen, J., and Norbers, J. A. (1980). Reference Samples for Electron Microprobe Analysis. *Geostandards Newsletter*, 4(1):43–47.
- Javoy, M. and Pineau, F. (1991). The volatiles record of a 'popping' rock from the Mid-Atlantic Ridge at 14° N: chemical and isotopic composition of gas trapped in the vesicles. *Earth and Planetary Science Letters*, 107(3-4):598–611.
- Jenkins, J., Cottaar, S., White, R. S., and Deuss, A. (2016). Depressed mantle discontinuities beneath Iceland: Evidence of a garnet controlled 660 km discontinuity? *Earth and Planetary Science Letters*, 433:159–168.

- Jenner, F. E. and O'Neill, H. S. C. (2012). Analysis of 60 elements in 616 ocean floor basaltic glasses. *Geochemistry, Geophysics, Geosystems*, 13(1).
- Jennings, E. S., Gibson, S. A., Maclennan, J., and Heinonen, J. S. (2017). Deep mixing of mantle melts beneath continental flood basalt provinces: constraints from olivine-hosted melt inclusions in primitive magmas. *Geochimica et Cosmochimica Acta*, 196:36–57.
- Jennings, E. S. and Holland, T. J. B. (2015). A simple thermodynamic model for melting of peridotite in the system NCFMASOCr. *Journal of Petrology*, 56(5):869–892.
- Jochum, K. P., Stoll, B., Herwig, K., Willbold, M., Hofmann, A. W., Amini, M., Aarburg, S., Abouchami, W., Hellebrand, E., Mocek, B., Raczek, I., Stracke, A., Alard, O., Bouman, C., Becker, S., Dücking, M., Brätz, H., Klemd, R., de Bruin, D., Canil, D., Cornell, D., de Hoog, C.-J., Dalpé, C., Danyushevsky, L., Eisenhauer, A., Gao, Y., Snow, J. E., Groschopf, N., Günther, D., Latkoczy, C., Guillong, M., Hauri, E. H., Höfer, H. E., Lahaye, Y., Horz, K., Jacob, D. E., Kasemann, S. A., Kent, A. J. R., Ludwig, T., Zack, T., Mason, P. R. D., Meixner, A., Rosner, M., Misawa, K., Nash, B. P., Pfänder, J., Premo, W. R., Sun, W. D., Tiepolo, M., Vannucci, R., Vennemann, T., Wayne, D., and Woodhead, J. D. (2006). MPI-DING reference glasses for in situ microanalysis: New reference values for element concentrations and isotope ratios. *Geochemistry, Geophysics, Geosystems*, 7(2):1–44.
- Jochum, K. P., Weis, U., Stoll, B., Kuzmin, D., Yang, Q., Raczek, I., Jacob, D. E., Stracke, A., Birbaum, K., Frick, D. A., Günther, D., and Enzweiler, J. (2011). Determination of Reference Values for NIST SRM 610-617 Glasses Following ISO Guidelines. *Geostandards* and Geoanalytical Research, 35(4):397–429.
- Jochum, K. P., Willbold, M., Raczek, I., Stoll, B., and Herwig, K. (2005). Chemical Characterisation of the USGS Reference Glasses GSA-1G, GSC-1G, GSD-1G, GSE-1G, BCR-2G, BHVO-2G and BIR-1G Using EPMA, ID-TIMS, ID-ICP-MS and LA-ICP-MS. *Geostandards and Geoanalytical Research*, 29(3):285–302.
- Johannes, W., Bell, P. M., Mao, H. K., Boettcher, A. L., Chipman, D. W., Hays, J. F., Newton, R. C., and Seifert, F. (1971). An interlaboratory comparison of piston-cylinder pressure calibration using the albite-breakdown reaction. *Contributions to Mineralogy and Petrology*, 32(1):24–38.
- Jóhannesson, H. and Sæmundsson, K. (2009). *Geological map of Iceland.* 1:600 000. *Tectonics (1st edition).* Icelandic Institute of Natural History, Reykjavik.
- Jónasson, K. (1994). Rhyolite volcanism in the Krafla central volcano, north-east Iceland. *Bulletin of Volcanology*, 56(6-7):516–528.
- Jones, S. M., White, N., Clarke, B., Rowley, E., and Gallagher, K. (2002). Present and past influence of the Iceland plume on sedimentation. *Geological Society, London, Special Publications*, 196:12–25.
- Kamenetsky, V., Davidson, P., Mernagh, T., Crawford, A., Gemmell, J., Portnyagin, M., and Shinjo, R. (2002). Fluid bubbles in melt inclusions and pillow-rim glasses: hightemperature precursors to hydrothermal fluids? *Chemical Geology*, 183(1):349–364.
- Katz, R. F. and Weatherley, S. M. (2012). Consequences of mantle heterogeneity for melt extraction at mid-ocean ridges. *Earth and Planetary Science Letters*, 335:226–237.
- Kawakami, Y., Yamamoto, J., and Kagi, H. (2003). Micro-Raman Densimeter for CO₂ Inclusions in Mantle-Derived Minerals. *Applied Spectroscopy*, 57(11):1333–1339.
- Keiding, J. K. and Sigmarsson, O. (2012). Geothermobarometry of the 2010 Eyjafjallajökull eruption: New constraints on Icelandic magma plumbing systems. *Journal of Geophysical Research*, 117(November 2011):B00C09.

- Kelemen, P. B., Hirth, G., Shimizu, N., Spiegelman, M., and Dick, H. J. B. (1997a). A review of melt migration processes in the adiabatically upwelling mantle beneath oceanic spreading ridges. *Philosophical Transactions: Mathematical, Physical and Engineering Sciences*, 355:283–318.
- Kelemen, P. B., Koga, K., and Shimizu, N. (1997b). Geochemistry of gabbro sills in the crust-mantle transition zone of the Oman ophiolite: implications for the origin of the oceanic lower crust. *Earth and Planetary Science Letters*, 146:475–488.
- Kelemen, P. B. and Manning, C. E. (2015). Reevaluating carbon fluxes in subduction zones, what goes down, mostly comes up. *Proceedings of the National Academy of Sciences*, 112(30):E3997–E4006.
- Kelemen, P. B., Shimizu, N., and Salters, V. J. (1995). Extraction of mid-ocean-ridge basalt from the upwelling mantle by focused flow of melt in dunite channels. *Nature*, 375(6534):747–753.
- Kelley, D. F. and Barton, M. (2008). Pressures of Crystallization of Icelandic Magmas. *Journal of Petrology*, 49(3):465–492.
- Keppler, H., Wiedenbeck, M., and Shcheka, S. S. (2003). Carbon solubility in olivine and the mode of carbon storage in the Earth's mantle. *Nature*, 424(6947):414–416.
- Kessel, R., Schmidt, M. W., Ulmer, P., and Pettke, T. (2005). Trace element signature of subduction–zone fluids, melts and supercritical liquids at 120–180 km depth. *Nature*, 437:724–727.
- Kim, D., Brown, L. D., Árnason, K., Águstsson, K., and Blanck, H. (2017). Magma reflection imaging in Krafla, Iceland, using microearthquake sources. *Journal of Geophysical Research: Solid Earth*, 122(7):5228–5242.
- Klein, E. M. and Langmuir, C. H. (1987). Global correlations of ocean ridge basalt chemistry with axial depth and crustal thickness. *Journal of Geophysical Research*, 92(B8):8089–8115.
- Kokfelt, T. F., Hoernle, K., and Hauff, F. (2003). Upwelling and melting of the Iceland plume from radial variation of ²³⁸U-²³⁰Th disequilibria in postglacial volcanic rocks. *Earth and Planetary Science Letters*, 214:167–186.
- Koleszar, A., Saal, A., Hauri, E., Nagle, A., Liang, Y., and Kurz, M. (2009). The volatile contents of the Galapagos plume; evidence for H₂O and F open system behavior in melt inclusions. *Earth and Planetary Science Letters*, 287(3-4):442–452.
- Kress, V. C. and Carmichael, I. S. E. (1991). The compressibility of silicate liquids containing Fe₂O₃ and the effect of composition, temperature, oxygen fugacity and pressure on their redox states. *Contributions to Mineralogy and Petrology*, 108:82–92.
- Kress, V. C. and Ghiorso, M. S. (2004). Thermodynamic modelling of post-entrapment crystallization in igneous phases. *Journal of Volcanology and Geothermal Research*, 137(4):247–260.
- Lambart, S., Laporte, D., and Schiano, P. (2009a). An experimental study of focused magma transport and basalt-peridotite interactions beneath mid-ocean ridges: Implications for the generation of primitive MORB compositions. *Contributions to Mineralogy and Petrology*, 157(4):429–451.
- Lambart, S., Laporte, D., and Schiano, P. (2009b). An experimental study of pyroxenite partial melts at 1 and 1.5 GPa: implications for the major-element composition of midocean ridge basalts. *Earth and Planetary Science Letters*, 282:335–347.

- Langmuir, C. H. (1989). Geochemical consequences of *in situ* crystallisation. *Nature*, 340:199–205.
- Langmuir, C. H., Hanson, G. N., and O'Hara, M. J. (1980). An evaluation of major element heterogeneity in the mantle sources of basalts. *Philosophical Transactions of the Royal Society of London. Series A, Mathematical and Physical Sciences*, 297(1431):383–407.
- Langmuir, C. H., Klein, E. M., and Plank, T. (1992). Petrological systematics of mid-ocean ridge basalts: Constraints on melt generation beneath ocean ridges. *American Geophysical Union Monograph*, 71:183–280.
- Laporte, D., Toplis, M. J., Seyler, M., and Devidal, J.-L. (2004). A new experimental technique for extracting liquids from peridotite at very low degrees of melting: application to partial melting of depleted peridotite. *Contributions to Mineralogy and Petrology*, 146:463–484.
- Laubier, M., Schiano, P., Doucelance, R., Ottolini, L., and Laporte, D. (2007). Olivine-hosted melt inclusions and melting processes beneath the FAMOUS zone (Mid-Atlantic-Ridge). *Chemical Geology*, 240:129–150.
- Le Voyer, M., Kelley, K., Cottrell, E., and Hauri, E. (2017). Heterogeneity in mantle carbon content from CO₂-undersaturated basalts. *Nature Communications*, 8:14062.
- Lees, J. M. (2007). Seismic tomography of magmatic systems. *Journal of Volcanology and Geothermal Research*, 167(1-4):37–56.
- Lord, O. T., Walter, M. J., Dasgupta, R., Walker, D., and Clark, S. M. (2009). Melting in the Fe-C system to 70 GPa. *Earth and Planetary Science Letters*, 284(1-2):157–167.
- Lowenstern, J. B. (1995). Applications of silicate-melt inclusions to the study of magmatic volatiles. *Magmas, Fluids and Ore Deposits. Mineralogical Association of Canada Short Course*, 23(January 1995):71–99.
- Lowenstern, J. B. (2015). Bursting the bubble of melt inclusions. *American Mineralogist*, 100(4):672–673.
- Maclennan, J. (2008a). Concurrent mixing and cooling of melts under Iceland. *Journal of Petrology*, 49(11):1931–1953.
- Maclennan, J. (2008b). Lead isotope variability in olivine-hosted melt inclusions from Iceland. *Geochimica et Cosmochimica Acta*, 72:4159–4176.
- Maclennan, J. (2017). Bubble formation and decrepitation control the CO₂ content of olivine-hosted melt inclusions. *Geochemistry, Geophysics, Geosystems*, 18(2):597–616.
- Maclennan, J., Gaetani, G. A., Hartley, M. E., Neave, D. A., and Winpenny, B. (2012). Petrological constraints on the crustal structure under rift zones. In *AGU Fall Meeting Abstracts*, volume T41G-07.
- Maclennan, J., Jull, M., McKenzie, D., Slater, L., and Grönvold, K. (2002). The link between volcanism and deglaciation in Iceland. *Geochemistry, Geophysics, Geosystems*, 3.
- Maclennan, J., McKenzie, D., and Grönvold, K. (2001a). Plume-driven upwelling under central Iceland. *Earth and Planetary Science Letters*, 194:67–82.
- Maclennan, J., McKenzie, D., Grönvold, K., Shimizu, N., Eiler, J. M., and Kitchen, N. (2003a). Melt mixing and crystallisation under Theistareykir, northeast Iceland. *Geochemistry, Geophysics, Geosystems*, 4(11).

- Maclennan, J., McKenzie, D., Grönvold, K., and Slater, L. (2001b). Crustal accretion under northern Iceland. *Earth and Planetary Science Letters*, 191:295–310.
- Maclennan, J., McKenzie, D., Hilton, F., Gronvöld, K., and Shimizu, N. (2003b). Geochemical variability in a single flow from northern Iceland. *Journal of Geophysical Research*, 108(B1).
- Manning, C. E. (2004). The chemistry of subduction-zone fluids. *Earth and Planetary Science Letters*, 223(1-2):1–16.
- Manning, C. E. and Boettcher, S. L. (1994). Rapid-quench hydrothermal experiments at mantle pressures and temperatures. *American Mineralogist*, 79(11-12):1153–1158.
- Manning, D. A. C. (1981). The effect of fluorine on liquidus phase relationships in the system Q-Ab-Or with excess water at 1 kbar. *Contrib. Mineral. Petrol.*, 76:205–215.
- Marty, B. (2012). The origins and concentrations of water, carbon, nitrogen and noble gases on Earth. *Earth and Planetary Science Letters*, 313-314(1):56–66.
- Marty, B. and Jambon., A. (1987). C/³He in volatile flux from the solid Earth: Implications for carbon dynamics. *Earth and Planetary Science Letters*, 83:16–26.
- Marty, B. and Tolstikhin, I. N. (1998). CO2 fluxes from mid-ocean ridges, arcs and plumes. *Chemical Geology*, 145(3-4):233–248.
- Mason, E., Edmonds, M., and Turchyn, A. V. (2017). Remobilization of crustal carbon may dominate volcanic arc emissions. *Science*, 357(6348):290–294.
- Masotta, M., Freda, C., Paul, T. A., Moore, G. M., Gaeta, M., Scarlato, P., and Troll, V. R. (2012). Low pressure experiments in piston cylinder apparatus: Calibration of newly designed 25 mm furnace assemblies to P = 150 MPa. *Chemical Geology*, 312-313(December):74–79.
- Masotta, M., Mollo, S., Freda, C., Gaeta, M., and Moore, G. (2013). Clinopyroxene-liquid thermometers and barometers specific to alkaline differentiated magmas. *Contributions to Mineralogy and Petrology*, 166(6):1545–1561.
- Matzen, A. K., Baker, M. B., Beckett, J. R., and Stolper, E. M. (2011). Fe-Mg partitioning between olivine and high-magnesian melts and the nature of Hawaiian parental liquids. *Journal of Petrology*, 52(7-8):1243–1263.
- McCulloch, M. T. and Gamble, J. A. (1991). Geochemical and geodynamical constraints on subduction zone magmatism. *Earth and Planetary Science Letters*, 102(3-4):358–374.
- McDonough, W. F. and Sun, S. (1995). The composition of the Earth. *Chemical Geology*, 120:223–253.
- McKenzie, D. (2000). Constraints on melt generation and transport rates from U-series activity ratios. *Chemical Geology*, 162:81–94.
- McKenzie, D. and O'Nions, R. K. (1991). Partial melt distributions from inversion of rare earth element concentrations. *Journal of Petrology*, 32(5):1021–1091.
- McKenzie, D., Stracke, A., Blichert-Toft, J., Albaréde, F., Grönvold, K., and O'Nions, R. K. (2004). Source enrichment processes responsible for isotopic anomalies in oceanic island basalts. *Geochimica et Cosmochimica Acta*, 68(12):2699–2724.
- McNutt, S. R. (1996). Seismic monitoring and eruption forecasting of volcanoes: a review of the state-of-the-art and case histories. In *Monitoring and mitigation of volcano hazards*, pages 99–146. Springer.

- Meibom, A. and Anderson, D. L. (2003). The statistical upper mantle assemblage. *Earth and Planetary Science Letters*, 217:123–139.
- Metrich, N., Zanon, V., Creon, L., Hildenbrand, A., Moreira, M., and Marques, F. O. (2014). Is the 'Azores Hotspot' a Wetspot? Insights from the Geochemistry of Fluid and Melt Inclusions in Olivine of Pico Basalts. *Journal of Petrology*, 55(2):377–393.
- Michael, P. (1995). Regionally distinctive sources of depleted MORB: evidence from trace elements and H₂O. *Earth and Planetary Science Letters*, 131:301–320.
- Michael, P. J. and Graham, D. W. (2015). The behavior and concentration of CO₂ in the suboceanic mantle: Inferences from undegassed ocean ridge and ocean island basalts. *Lithos*, 236-237:338–351.
- Mikhail, S. and Howell, D. (2016). A petrological assessment of diamond as a recorder of the mantle nitrogen cycle. *American Mineralogist*, 101(4):780–787.
- Minarik, W. G. (1998). Complications to Carbonate Melt Mobility due to the Presence of an Immiscible Silicate Melt. *Journal of Petrology*, 39(11-12):1965–1973.
- Minarik, W. G. and Watson, E. (1995). Interconnectivity of carbonate melt at low melt fraction. *Earth and Planetary Science Letters*, 133(3-4):423–437.
- Mirwald, P. W., Getting, I. C., Kennedy, G. C., and Angeles, L. (1975). Low-Friction Cell for Piston-Cylinder High-Pressure Apparatus Hard-fired Pyrophyllite Soft-fired Stainless Steel Boron Nitride Disk Lead Foil + MoS 2 Grease. *Journal of Geophysical Research*, 80(11):1519–1525.
- Mitchell, M. A., White, R. S., Roecker, S., and Greenfield, T. (2013). Tomographic image of melt storage beneath Askja Volcano, Iceland using local microseismicity. *Geophysical Research Letters*, 40(19):5040–5046.
- Mollo, S., Del Gaudio, P., Ventura, G., Iezzi, G., and Scarlato, P. (2010). Dependence of clinopyroxene composition on cooling rate in basaltic magmas: Implications for thermo-barometry. *Lithos*, 118(3-4):302–312.
- Mollo, S., Putirka, K., Misiti, V., Soligo, M., and Scarlato, P. (2013). A new test for equilibrium based on clinopyroxene-melt pairs: Clues on the solidification temperatures of Etnean alkaline melts at post-eruptive conditions. *Chemical Geology*, 352:92–100.
- Moore, G. (2008). Interpreting H₂O and CO₂ Contents in Melt Inclusions: Constraints from Solubility Experiments and Modelling. *Reviews in Mineralogy and Geochemistry*, 69(1):333–362.
- Moore, G., Roggensack, K., and Klonowski, S. (2008). A low-pressure-high-temperature technique for the piston-cylinder. *American Mineralogist*, 93(1):48–52.
- Moore, K. R. and Wood, B. J. (1998). The Transition from Carbonate to Silicate Melts in the CaO–MgO–SiO2–CO2 System. *Journal of Petrology*, 39(11-12):1943–1951.
- Moore, L. R. (2014). Bubbles matter: An assessment of the contribution of vapor bubbles to melt inclusion volatile budgets. 100(Mi):806–823.
- Moune, S., Sigmarsson, O., Schiano, P., Thordarson, T., and Keiding, J. K. (2012). Melt inclusion constraints on the magma source of Eyjafjallajökull 2010 flank eruption. *Journal of Geophysical Research: Solid Earth*, 117(B9).
- Mukhopadhyay, S. (2012). Early differentiation and volatile accretion recorded in deepmantle neon and xenon. *Nature*, 486:101–104.

- Mysen, B. O., Cody, G. D., and Smith, A. (2004). Solubility mechanisms of fluorine in peralkaline and meta-aluminous silicate glasses and in melts to magmatic temperatures. *Geochimica et Cosmochimica Acta*, 68(12):2745–2769.
- Namur, O., Charlier, B., Toplis, M. J., and Vander Auwera, J. (2012). Prediction of plagioclase-melt equilibria in anhydrous silicate melts at 1-atm. *Contributions to Mineral*ogy and Petrology, 163(1):133–150.
- Neave, D. A., Maclennan, J., Edmonds, M., and Thordarson, T. (2014). Melt mixing causes negative correlation of trace element enrichment and CO₂ content prior to an Icelandic eruption. *Earth and Planetary Science Letters*, 400:272–283.
- Neave, D. A., Maclennan, J., Thordarson, T., and Hartley, M. E. (2015). The evolution and storage of primitive melts in the Eastern Volcanic Zone of Iceland: the 10 ka Grímsvötn tephra series (i.e. the Saksunarvatn ash). *Contributions to Mineralogy and Petrology*, 170(2):21.
- Neave, D. A., Namur, O., and Holtz, F. (2017). The effect of mantle-derived variability on the mineralogy of primitive basalts: Experimental constraints from icelandic systems. In *IAVCEI Meeting Abstracts*, volume 284, page 756.
- Neave, D. A., Passmore, E., Maclennan, J., Fitton, G., and Thordarson, T. (2013). Crystal– melt relationships and the record of deep mixing and crystallisation in the AD 1783 Laki eruption, Iceland. *Journal of Petrology*, 54:1661–1690.
- Neave, D. A. and Putirka, K. D. (2017). A new clinopyroxene-liquid barometer, and implications for magma storage pressures under Icelandic rift zones. *American Mineralogist*, 102(4):777–794.
- Nelson, S. T. and Montana, A. (1992). Sieve-textured plagioclase in volcanic rocks produced by rapid decompression. *American Mineralogist*, 77(11-12):1242–1249.
- Newman, S. and Lowenstern, J. B. (2002). Volatile Calc : a silicate melt H₂O CO₂ solution model written in Visual Basic for excel. *Computers & Geosciences*, 28:597–604.
- Nicholson, H., Condomines, M., Fitton, J. G., Fallick, A. E., Gronvöld, K., and Rogers, G. (1991). Geochemical and isotopic evidence for crustal assimilation beneath Krafla, Iceland. *Journal of Petrology*, 32(5):1005–1020.
- Nimis, P. (1995). A clinopyroxene geobarometer for basaltic systems based on crystalstructure modeling. *Contributions to Mineralogy and Petrology*, 121(2):115–125.
- Oganov, A. R., Ono, S., Ma, Y., Glass, C. W., and Garcia, A. (2008). Novel high-pressure structures of MgCO₃, CaCO₃ and CO₂ and their role in Earth's lower mantle. *Earth and Planetary Science Letters*, 273(1-2):38–47.
- Padrón-Navarta, J. A., Hermann, J., and O'Neill, H. S. C. (2014). Site-specific hydrogen diffusion rates in forsterite. *Earth and Planetary Science Letters*, 392:100–112.
- Palme, H. and O'Neill, H. (2014). Cosmochemical Estimates of Mantle Composition. In *Treatise on Geochemistry*, volume 3, pages 1–39. Elsevier, 2 edition.
- Peate, D. W., Breddam, K., Baker, J. A., Kurz, M. D., Barker, A. K., Prestvik, T., Grassineau, N., and Skovgaard, A. C. (2010). Compositional characteristics and spatial distribution of enriched Icelandic mantle components. *Journal of Petrology*, 51(7):1447–1475.
- Peate, D. W. and Hawkesworth, C. J. (2005). U series disequilibria: Insights into mantle melting and the timescales of magma differentiation. *Reviews of Geophysics*, 43(1). RG1003.

- Peate, D. W., Hawkesworth, C. J., van Calsteren, P. W., Taylor, R. N., and Murton, B. J. (2001). ²³⁸U–²³⁰Th constraints on mantle upwelling and plume–ridge interaction along the Reykjanes Ridge. *Earth and Planetary Science Letters*, 187:259–272.
- Peslier, A. H., Bizimis, M., and Matney, M. (2015). Water disequilibrium in olivines from Hawaiian peridotites: Recent metasomatism, H diffusion and magma ascent rates. *Geochimica et Cosmochimica Acta*, 154:98–117.
- Pichavant, M., Di Carlo, I., Le Gac, Y., Rotolo, S. G., and Scaillet, B. (2009). Experimental Constraints on the Deep Magma Feeding System at Stromboli Volcano, Italy. *Journal of Petrology*, 50(4):601–624.
- Pilidou, S., Priestley, K., Debayle, E., and Gudmundsson, O. (2005). Rayleigh wave tomography in the North Atlantic: High resolution images of the Iceland, Azores and Eifel mantle plumes. *Lithos*, 79(3-4 SPEC. ISS.):453–474.
- Plank, T. and Langmuir, C. H. (1992). Effects of the melting regime on the composition of the oceanic crust. *Journal of Geophysical Research*, 97(B13):19749–19770.
- Powell, R. and Holland, T. J. B. (2008). On thermobarometry. *Journal of Metamorphic Geology*, 26(2):155–179.
- Putirka, K. (1999). Clinopyroxene + liquid equilibria to 100 kbar and 2450 K. *Contributions to Mineralogy and Petrology*, 135(2-3):151–163.
- Putirka, K. (2008a). Excess temperatures at ocean islands: implications for mantle layering and convection. *Geology*, 36:283–286.
- Putirka, K., Johnson, M., Kinzler, R., Longhi, J., and Walker, D. (1996). Thermobarometry of mafic igneous rocks based on clinopyroxene-liquid equilibria, 0-30 kbar. *Contributions to Mineralogy and Petrology*, 123(1):92–108.
- Putirka, K. D. (2008b). Thermometers and Barometers for Volcanic Systems. *Reviews in Mineralogy and Geochemistry*, 69(1):61–120.
- Putirka, K. D., Mikaelian, H., Ryerson, F., and Shaw, H. (2003). New clinopyroxeneliquid thermobarometers for mafic, evolved, and volatile-bearing lava compositions, with applications to lavas from Tibet and the Snake River Plain, Idaho. *American Mineralogist*, 88(10):1542–1554.
- Reed, S. J. B. (2005). *Electron Microprobe Analysis and Scanning Electron Microscopy in Geology*. Cambridge University Press.
- Regier, M. E., Hervig, R. L., Myers, M. L., Roggensack, K., and Wilson, C. J. (2016). Analysing nitrogen in natural and synthetic silicate glasses by secondary ion mass spectroscopy. *Chemical Geology*, 447:27–39.
- Reverso, T., Vandemeulebrouck, J., Jouanne, F., Pinel, V., Villemin, T., Sturkell, E., and Bascou, P. (2014). A two-magma chamber model as a source of deformation at Grímsvötn Volcano, Iceland. *Journal of Geophysical Research: Solid Earth*, 119(6):4666–4683.
- Rickers, F., Fitchner, A., and Trampert, J. (2013). The Iceland-Jan Mayen plume system and its impact on mantle dynamics in the North Atlantic region: Evidence from full-waveform inversion. *Earth and Planetary Science Letters*, 367:39–51.
- Risku-Norja, H. (1985). Gabbro nodules from picritic pillow basalt, SW Iceland. Nordic Volcanological Institute Prof Paper, 8501:32–49.

Roedder, E. (1984). Fluid inclusions. *Reviews in Mineralogy and Geochemistry*, 12.

- Roedder, E. and Weiblen, P. W. (1970). Silicate liquid immiscibility in lunar magmas, evidenced by melt inclusions in lunar rocks. *Science*, 167:641–644.
- Rosenthal, A., Hauri, E., and Hirschmann, M. (2015). Experimental determination of C, F, and H partitioning between mantle minerals and carbonated basalt, CO₂/Ba and CO₂/Nb systematics of partial melting, and the CO₂ contents of basaltic source regions. *Earth and Planetary Science Letters*, 412:77–87.
- Rudge, J. F., Maclennan, J., and Stracke, A. (2013). The geochemical consequences of mixing melts from a heterogeneous mantle. *Geochimica et Cosmochimica Acta*, 114:112–143.
- Saal, A. E., Hauri, E., Langmuir, C. H., and Perfit, M. (2002). Vapour undersaturation in primitive mid-ocean-ridge basalt and the volatile content of Earth's upper mantle. *Nature*, 419:451–455.
- Schilling, J. G. (1973). Iceland mantle plume: Geochemical study of Reykjanes ridge. *Nature*, 242:565–571.
- Schipper, C. I., Le Voyer, M., Moussallam, Y., White, J. D. L., Thordarson, T., Kimura, J.-I., and Chang, Q. (2016). Degassing and magma mixing during the eruption of Surtsey Volcano (Iceland, 1963–1967): the signatures of a dynamic and discrete rift propagation event. *Bulletin of Volcanology*, 78(4):33.
- Shcheka, S. S., Wiedenbeck, M., Frost, D. J., and Keppler, H. (2006). Carbon solubility in mantle minerals. *Earth and Planetary Science Letters*, 245(3-4):730–742.
- Shejwalkar, A. and Coogan, L. A. (2013). Experimental calibration of the roles of temperature and composition in the Ca-in-olivine geothermometer at 0.1MPa. *Lithos*, 177:54–60.
- Shen, Y. and Forsyth, D. (1995). Geochemical constraints on initial and final depths of melting beneath mid-ocean ridges. *Journal of Geophysical Research*, 100(B2):2211–2237.
- Shishkina, T., Botcharnikov, R., Holtz, F., Almeev, R., and Portnyagin, M. (2010). Solubility of H₂O- and CO₂-bearing fluids in tholeiitic basalts at pressures up to 500 MPa. *Chemical Geology*, 277(1-2):115–125.
- Shishkina, T. A., Botcharnikov, R. E., Holtz, F., Almeev, R. R., Jazwa, A. M., and Jakubiak, A. A. (2014). Compositional and pressure effects on the solubility of H₂O and CO₂ in mafic melts. *Chemical Geology*, 388:112–129.
- Shorttle, O. (2015). Geochemical variability in MORB controlled by concurrent mixing and crystallisation. *Earth and Planetary Science Letters*, 424:1–14.
- Shorttle, O. and Maclennan, J. (2011). Compositional trends of Icelandic basalts: Implications for short-lengthscale lithological heterogeneity in mantle plumes. *Geochemistry*, *Geophysics*, *Geosystems*, 12(11).
- Shorttle, O., Maclennan, J., and Lambart, S. (2014). Quantifying lithological variability in the mantle. *Earth and Planetary Science Letters*, 395:24–40.
- Shorttle, O., Rudge, J. F., Maclennan, J., and Rubin, K. H. (2016). A Statistical Description of Concurrent Mixing and Crystallization during MORB Differentiation: Implications for Trace Element Enrichment. *Journal of Petrology*, 57(11-12):2127–2162.
- Sides, I., Edmonds, M., Maclennan, J., Houghton, B., Swanson, D., and Steele-MacInnis, M. (2014). Magma mixing and high fountaining during the 1959 Kilauea Iki eruption, Hawaii. *Earth and Planetary Science Letters*, 400:102–112.

- Siewert, R., Buttner, H., and Rosenhauer, M. (1998). Experimental investigation of thermodynamic melting properties in the system NaCl-KCl at pressures of up to 7000 bar. *Neues Jahrbuch für Mineralogie - Abhandlungen*, 172:259–278.
- Sigmundsson, F., Hreinsdóttir, S., Hooper, A., Árnadóttir, T., Pedersen, R., Roberts, M. J., Óskarsson, N., Auriac, A., Decriem, J., Einarsson, P., Geirsson, H., Hensch, M., Ófeigsson, B. G., Sturkell, E., Sveinbjörnsson, H., and Feigl, K. L. (2010). Intrusion triggering of the 2010 Eyjafjallajökull explosive eruption. *Nature*, 468(7322):426–430.
- Slater, L., McKenzie, D., Gronvöld, K., and Shimizu, N. (2001). Melt generation and movement beneath Theistareykir, NE Iceland. *Journal of Petrology*, 42(2):321–354.
- Sobolev, A. V., Hofmann, A. W., Kuzmin, D. V., Yaxley, G., Arndt, N. T., Chung, S.-L., Danyushevsky, L. V., Elliott, T., Frey, F. A., Garcia, M. O., Gurenko, A. A., Kamenetsky, V. S., Kerr, A. C., Krivolutskaya, N. A., Matvienkov, V. V., Nikogosian, I. K., Rocholl, A., Sigurdsson, I. A., Sushchevskaya, N. M., and Teklay, M. (2007). The amount of recycled crust in sources of mantle derived melts. *Science*, 316:590–597.
- Sobolev, A. V., Hofmann, A. W., and Nikogosian, I. K. (2000). Recycled oceanic crust observed in 'ghost plagioclase' within the source of Mauna Loa lavas. *Nature*, 404(6781):986–990.
- Sobolev, A. V. and Shimizu, N. (1993). Ultra-depleted primary melt included in an olivine from the Mid-Atlantic Ridge. *Nature*, 363:151–154.
- Stagno, V., Ojwang, D. O., McCammon, C. A., and Frost, D. J. (2013). The oxidation state of the mantle and the extraction of carbon from the Earth's interior. *Nature*, 493:84–88.
- Steele-Macinnis, M., Esposito, R., and Bodnar, R. J. (2011). Thermodynamic Model for the Effect of Post-entrapment Crystallization on the H₂O-CO₂ Systematics of Vapor-saturated, Silicate Melt Inclusions. *Journal of Petrology*, 52(12):2461–2482.
- Stracke, A. (2012). Earth's heterogeneous mantle: a product of convection driven interaction between crust and mantle. *Chemical Geology*, 330–331:274–299.
- Stracke, A. and Bourdon, B. (2009). The importance of melt extraction for tracing mantle heterogeneity. *Geochimica et Cosmochimica Acta*, 73(1):218–238.
- Stracke, A., Hofmann, A. W., and Hart, S. R. (2005). FOZO, HIMU, and the rest of the mantle zoo. *Geochemistry, Geophysics, Geosystems*, 6(5).
- Stracke, A., Zindler, A., Salters, V. J. M., McKenzie, D., Blichert-Toft, J., Albarède, F., and Grönvold, K. (2003). Theistareykir revisited. *Geochemistry, Geophysics, Geosystems*, 4(2).
- Sugawara, T. (2000). Empirical relationships between temperature, pressure, and MgO content in olivine and pyroxene saturated liquid. *Journal of Geophysical Research*, 105(B4):8457.
- Tarasewicz, J., White, R. S., Brandsdóttir, B., and Schoonman, C. M. (2014). Seismogenic magma intrusion before the 2010 eruption of Eyjafjallajökull volcano, Iceland. *Geophysical Journal International*, 198(2):906–921.
- Tepley, F. J., Davidson, J. P., and Clynne, M. a. (1999). Magmatic interactions as recorded in plagioclase phenocrysts of Chaos Crags, Lassen Volcanic Center, California. *Journal of Petrology*, 40(5):787–806.
- Thirlwall, M. F. (1995). Generation of Pb isotopic characteristics of the Iceland plume. *Journal of the Geological Society, London*, 152:991–996.

Thirlwall, M. F., Gee, M. A. M., Taylor, R. N., and Murton, B. J. (2004). Mantle components in Iceland and adjacent ridges investigated using double-spike Pb isotope ratios. *Geochimica et Cosmochimica Acta*, 68(2):361–386.

Thordarson, T. and Hoskuldsson, A. (2002). Iceland. Classic geology in Europe. Terra.

- Thordarson, T. and Höskuldsson, Á. (2008). Postglacial volcanism in iceland. *Jökull*, 58(198):e228.
- Thordarson, T. and Larsen, G. (2007). Volcanism in Iceland in historical time: Volcano types, eruption styles and eruptive history. *Journal of Geodynamics*, 43(1):118–152.
- Trieloff, M. and Kunz, J. (2005). Isotope systematics of noble gases in the Earth's mantle: possible sources of primordial isotopes and implications for mantle structure. *Physics of the Earth and Planetary Interiors*, 148(1):13–38.
- Trønnes, R. G. (1990). Basaltic Melt Evolution of the Hengill Volcanic System, SW Iceland, and Evidence for Clinopyroxene Assimilation in Primitive Tholeiitic Magmas. *Journal of Geophysical Research*, 95(B10):15893–15910.
- Vernon, R. (2004). A Practical Guide to Rock Microstructure. Cambridge University Press.
- Vink, G. E. (1984). A hotspot model for Iceland and the Vøring Plateau. *Journal of Geophysical Research*, 89(B12):9949.
- Voigt, M., Coogan, L. A., and von der Handt, A. (2017). Experimental investigation of the stability of clinopyroxene in mid-ocean ridge basalts: The role of Cr and Ca/Al. *Lithos*, 274-275:240–253.
- Walter, M. J. (1998). Melting of garnet peridotite and the origin of komatiite and depleted lithosphere. *Journal of Petrology*, 39(1):29–60.
- Wanless, V. D., Behn, M. D., Shaw, A. M., and Plank, T. (2014). Variations in melting dynamics and mantle compositions along the Eastern Volcanic Zone of the Gakkel Ridge: insights from olivine-hosted melt inclusions. *Contributions to Mineralogy and Petrology*, 167(5):1005.
- Wanless, V. D., Perfit, M. R., Klein, E. M., White, S., and Ridley, W. I. (2012). Reconciling geochemical and geophysical observations of magma supply and melt distribution at the 9°N overlapping spreading center, East Pacific Rise. *Geochemistry, Geophysics, Geosystems*, 13(11):1–22.
- Wanless, V. D. and Shaw, A. M. (2012). Lower crustal crystallization and melt evolution at mid-ocean ridges. *Nature Geosci*, 5(9):651–655.
- Wanless, V. D., Shaw, A. M., Behn, M. D., Soule, S. A., Escartín, J., and Hamelin, C. (2015). Magmatic plumbing at Lucky Strike volcano based on olivine-hosted melt inclusion compositions. *Geochemistry, Geophysics, Geosystems*, 16(1):126–147.
- Weatherley, S. M. and Katz, R. F. (2012). Melting and channelized magmatic flow in chemically heterogeneous, upwelling mantle. *Geochemistry, Geophysics, Geosystems*, 13(1):226–237.
- Welsch, B., Hammer, J., Baronnet, A., Jacob, S., Hellebrand, E., and Sinton, J. (2016). Clinopyroxene in postshield Haleakala ankaramite: 2. Texture, compositional zoning and supersaturation in the magma. *Contributions to Mineralogy and Petrology*, 171(1):1–19.
- White, R. and McKenzie, D. (1989). Magmatism at rift zones: The generation of volcanic continental margins and flood basalts. *Journal of Geophysical Research*, 94(B6):7685.

- White, R. S. and McKenzie, D. (1995). Mantle plumes and flood basalts. *Journal of Geophysical Research*, 100(B9):17543–17585.
- White, R. S., McKenzie, D., and O'Nions, K. (1992). Oceanic crustal thickness from seismic measurements and rare earth element inversions. *Journal of Geophysical Research*, 97(B13):19683–19715.
- Winpenny, B. and Maclennan, J. (2011). A partial record of mixing of mantle melts preserved in Icelandic phenocrysts. *Journal of Petrology*, 52(9):1791–1812.
- Winpenny, B. and Maclennan, J. (2014). Short Length Scale Oxygen Isotope Heterogeneity in the Icelandic Mantle: Evidence from Plagioclase Compositional Zones. *Journal of Petrology*, 55(12):2537–2566.
- Witham, F., Blundy, J., Kohn, S. C., Lesne, P., Dixon, J., Churakov, S. V., and Botcharnikov, R. (2012). SolEx: A model for mixed COHSCl-volatile solubilities and exsolved gas compositions in basalt. *Computers & Geosciences*, 45:87–97.
- Workman, R. K. and Hart, S. R. (2005). Major and trace element composition of the depleted MORB mantle (DMM). *Earth and Planetary Science Letters*, 231:53–72.
- Workman, R. K., Hauri, E., Hart, S. R., Wang, J., and Blusztajn, J. (2006). Volatile and trace elements in basaltic glasses from Samoa: Implications for water distribution in the mantle. *Earth and Planetary Science Letters*, 241(3-4):932–951.
- Yang, H.-J., Kinzler, R. J., and Grove, T. L. (1996). Experiments and models of anhydrous, basaltic olivine-plagioclase-augite saturated melts from 0.001 to 10 kbar. *Contributions to Mineralogy and Petrology*, 124:1–18.
- Yaxley, G. M. and Brey, G. P. (2004). Phase relations of carbonate bearing eclogite assemblages from 2.5 to 5.5 GPa: implications for petrogenesis of carbonatites. *Contributions to Mineralogy and Petrology*, 146:606–619.
- Yuan, K. and Romanowicz, B. (2017). Seismic evidence for partial melting at the root of major hot spot plumes. *Science*, 357(6349):393–397.
- Zhu, W., Gaetani, G. A., Fusseis, F., Montési, L. G. J., and De Carlo, F. (2011). Microtomography of partially molten rocks: Three-dimensional melt distribution in mantle peridotite. *Science*, 332(6025):88–91.
- Ziberna, L., Green, E. C. R., and Blundy, J. D. (2017). Multiple-reaction geobarometry for olivine-bearing igneous rocks. *American Mineralogist*, in press.
- Zinner, E. and Crozaz, G. (1986). A method for the quantitative measurement of rare earth elements in the ion microprobe. *International Journal of Mass Spectrometry and Ion Processes*, 69(1):17–38.

Appendix A

Data corrections and standards

Table A.1 Secondary standard corrections for SIMS trace element analyses from Miðfell sessions one (January) and two (June), and Kistufell session one (April). Silica corrections for experimental glass measurements from three EPMA sessions.

	Miðfel	l MIs	Kistufell MIs	Expe	rimental	glass
Element	January	June	April	March	July	August
SiO ₂	-	-	-	1.012	0.995	1.007
Li	1.266	1.298	-	-	-	-
Κ	1.242	1.250	1.313	-	-	-
Ba	1.298	1.312	1.298	-	-	-
Nb	1.065	1.069	1.086	-	-	-
La	1.202	1.158	1.138	-	-	-
Ce	1.265	1.219	1.206	-	-	-
Pr	1.249	1.222	1.206	-	-	-
Sr	1.215	1.214	1.211	-	-	-
Nd	1.191	1.137	1.129	-	-	-
Zr	1.108	1.097	1.076	-	-	-
Ti	1.170	1.212	1.202	-	-	-
Sm	1.252	1.231	1.207	-	-	-
Eu	1.184	1.189	1.176	-	-	-
Gd	1.180	1.185	1.283	-	-	-
Gd	1.113	1.191	1.263	-	-	-
Tb	1.208	1.164	1.179	-	-	-
Dy	1.234	1.181	1.171	-	-	-
Но	1.177	1.150	1.136	-	-	-
Y	1.146	1.148	1.144	-	-	-
Er	1.241	1.201	1.173	-	-	-
Tm	1.166	1.133	1.136	-	-	-
Yb	1.172	1.111	1.104	-	-	-
Lu	1.112	1.064	1.098	-	-	-

	fit	(1)		fit(2)	
Element	a	b	d	е	f
Li	0.01056	0.03533	na	na	na
F	0.15371	-2.01883	na	na	na
Cl	0.62250	-1.58219	na	na	na
Κ	0.05323	-0.81151	na	na	na
Ti	0.02704	-0.05281	na	na	na
Sr	0.02490	0.27036	na	na	na
Y	0.02052	0.04136	na	na	na
Zr	0.02257	0.25336	na	na	na
Nb	0.03401	0.07696	0.06332	-1.33824	0.03973
Ba	0.03184	0.20409	1.20644	-2.77474	0.03543
La	0.02308	0.12317	0.03166	-6.15235	0.03735
Ce	0.03436	0.16188	0.16173	-0.99876	0.03434
Pr	0.02226	0.05148	0.06001	-0.81868	0.01807
Nd	0.04059	0.15467	na	na	na
Sm	0.02252	0.16566	0.17951	-1.59913	0.03443
Eu	0.06708	0.04961	na	na	na
Gd	0.07956	0.84666	22.98523	-10.67858	0.23437
Tb	0.02937	0.05115	0.03898	-1.26870	0.03363
Dy	0.04611	0.18173	na	na	na
Но	0.02092	0.05276	0.02594	-2.27307	0.02832
Er	0.02727	0.11521	0.13559	-1.47075	0.03215
Tm	0.02681	0.01897	0.00845	-1.54260	0.03020
Yb	0.04714	0.21898	na	na	na
Lu	0.02857	0.02534	na	na	na
H_2O	-0.36172	0.00074	na	na	na
CO_2	0.10994	-2.24825	na	na	na

Table A.2 Curve fitting parameters for Miðfell melt inclusion relative error estimates.

Coefficients in bold font indicate that the equation fits the standards data reasonably well and can be used to calculate the relative error of measured unknowns.

Appendix B

Dataset references

Global melt inclusion dataset

Table B.1 Sources of global melt inclusion and MORB glass data from different magmatic settings:Mid Ocean Ridge (MOR); Ocean Island (OI); Continental Rift Zone (CRZ).

Source	Location	Setting	No. samples
Melt inclusions			
Colman et al. (2015)	Galápagos Ridge	MOR	81
Helo et al. (2011)	Axial Seamount, Juan de Fuca Ridge	MOR	49
Le Voyer et al. (2017)	Mid-Atlantic Ridge	MOR	23
Saal et al. (2002)	Siqueiros, East Pacific Rise	MOR	107
Wanless et al. (2012)	East Pacific Rise	MOR	118
Wanless and Shaw (2012)	Juan de Fuca Ridge	MOR	45
Wanless et al. (2014)	Gakkel Ridge, Arctic	MOR	66
Wanless et al. (2015)	Lucky strike, Mid-Atlantic Ridge	MOR	63
Anderson and Brown (1993)	Kilauea, Hawaii	OI	50
Cabral et al. (2014)	Mangaia, Pacific	OI	51
Di Muro et al. (2014)	La Réunion, Indian	OI	67
Hartley et al. (2014)	Laki, Iceland	OI	126
Koleszar et al. (2009)	Galápagos	OI	120
Metrich et al. (2014)	Azores, Atlantic	OI	53
Moune et al. (2012)	Fimmvörðuháls, Iceland	OI	26
Neave et al. (2014)	Skuggafjöll, Iceland	OI	112
Neave et al. (2015)	Saksunarvatn, Iceland	OI	133
Schipper et al. (2016)	Surtsey, Iceland	OI	22
Sides et al. (2014)	Kilauea, Hawaii	OI	133
Wallace et al. (2015)	Hawaii	OI	29
Head et al. (2011)	Nyamuragira, East African Rift	CRZ	34
Hudgins et al. (2015)	Western East African Rift	CRZ	101
MORB glasses			
Michael and Graham (2015)	Various MORs	MOR	51

Appendix C

Extended data tables

Table C.1 Volatile and trace element data from Kistufell melt inclusions measured by SIMS. Glasses (Gl) ol003-hg1 to ol092-hg2 and melt inclusions (MI) ol003-hm1 to ol101-hm1 were measured in April 2014, glasses ol103-hg1 to ol149-hg1 and melt inclusions ol102-hm1 to ol170-hm1 were measured in August 2014. Ol - olivine number, In - inclusion number. H₂O in wt%, all other elements in ppm.

Sample	Eruption	ō	Ц	Phase	H ₂ O	co ₂	н	۵	м	ц	Sr	Y	Zr]	ЧN	Ba Lá	a Č	e Pr	ΡN	Sm	Eu ¹	56Gd ¹	57 Gd	Tb	Dy	юН	Er]	E E	- P	3
ol003-hg1	Kistufell	б	1	Ū	0.121	19.0	99.2	29.0	551 5	1325 11	13.7 17	7.36 35	3.09 2.	.69 13.	.53 2.50	5 6.32	2 0.93	5.10	2.04	0.72	1.38	2.35	0.43	2.59	0.54 1	1.53 0.	21 1.	83 0.	25
ol004-hg1	Kistufell	4	1	G	0.110	4.7	87.6	25.9	546 5	5357 11	14.0 15	7.50 35	8.13 2.	.61 12.	.96 2.5	5 6.5(0.94	5.34	1.81	0.80	0.79	2.48	0.48	2.90	0.67	1.58 0.	23 2.	18 0.	56
ol057-hg1	Kistufell	57	,	G	0.210	0.0 1	18.2	35.1	675 5	5750 12	23.3 17	7.39 41	.26 3.	.29 15.	.97 2.7	1 6.55	9 0.99	5.82	1.86	0.78	1.76	2.15	0.38	3.21	0.68	.84 0.	31 1.	70 0.	31
ol059-hg1	Kistufell	59	'	G	0.117	0.0 15	05.6	26.2	; 699	5627 12	23.1 17	7.30 41	.73 3.	.34 15.	.95 2.6-	4 6.66	5 1.02	6.37	1.99	0.77	1.09	2.72	0.44	3.04	0.70 2	2.10 0.	32 1.	95 0.	23
ol060-hg1	Kistufell	60	1	G	0.131	na 1	01.4	24.6	546 5	5359 11	15.2 17	7.62 35	0.23 2.	.74 13.	.00 2.4	8 6.24	4 0.92	5.92	1.84	0.74	1.62	2.67	0.44	3.06	0.66	.83 0.	24 2.	10 0.	21
ol062-hg1	Kistufell	62	'	Ū	0.141	na	94.4	29.6	553 5	5425 11	16.5 17	7.31 35	0.83 2.	39 13.	.79 2.8.	3 6.37	7 0.95	5.58	1.63	0.73	1.79	2.60	0.34	2.82	0.61 2	0.00	27 1.	84 0.	12
ol069-hg1	Kistufell	69	'	G	0.131	na	90.8	20.5	568 5	5325 11	15.6 17	7.24 35	3.72 2.	.93 14.	.17 2.6	8 6.31	1 0.89	5.97	1.71	0.84	1.98	2.72	0.40	2.69	0.63 1	.84 0.	24 1.	80 0.	30
ol071-hg1	Kistufell	71	'	G	0.137	na	9.99	27.7	545 5	11 2023	13.9 16	5.97 35	0.03 2.	51 12.	.94 2.6	8 6.15	3 0.92	6.09	1.89	0.60	2.03	2.53	0.43	2.99	0.63 1	0.77	28 2.0	03 0.	26
ol089-hg1	Kistufell	89	'	Ū	0.174	19.1 1	11.3	34.0	619 5	500 11	17. 17	7.13 35	0.87 2.	.88 15.	.36 2.6	8 6.5t	5 1.12	5.58	1.91	0.82	1.52	2.31	0.49	3.01	0.67	.87 0.	26 2.	24 0.	23
ol090-hg1	Kistufell	90	'	Ū	0.135	na	95.3	24.9	536 5	11 1397	16.3 17	7.23 4(.51 2.	57 13.	.11 2.7(J 6.8(0 1.04	5.56	1.90	0.73	1.20	2.09	0.40	2.89	0.55 2	2.05 0.	25 1.	86 0.	72
ol092-hg2	Kistufell	92	'	G	0.134	na	91.2	29.5	556 5	335 11	16.2 17	7.08 35	3.64 2.	.67 13.	.53 2.4.	2 6.5t	5 0.89	5.48	2.04	0.76	2.05	2.16	0.40	2.77	0.64	.67 0.	28 1.	84 0.	25
ol103-hg1	Kistufell	103	'	Ū	0.167	2.0	90.3	7.76	511 5	5004 10	34.1 16	5.04 37	7.11 2.	35 11.	.92 2.3.	2 5.79	9 1.08	5.92	1.54	0.67	2.44	1.82	0.42	3.19	0.58 1	1.78 0.	21 1.	53 0.	51
ol149-hg1	Kistufell	149	ī	Ū	0.154	28.1	94.4	95.7	527 5	5126 10	J6.4 I£	5.25 37	7.78 2.	.79 13.	.24 2.4	8 6.16	5 1.03	5.19	1.85	0.69	2.11	2.34	0.44	2.90	0.61	0 66.1	28 1.	47 0.	22
ol003-hm1	Kistufell	3	-	IW	0.118	545.6	97.9	28.6	728 €	403 14	43.1 19	.20 41	.05 2.	92 17.	.61 3.1:	5 7.80) 1.28	6.67	2.21	0.88	2.02	2.69	0.59	3.21	0.79 2	2.27 0.	35 1.	93 0.	1 %
ol004-hm1	Kistufell	4	-	IM	0.117	796.8	88.9	26.0	655 6	306 13	36.6 19	35 35	.97 2.	68 14.	.70 2.72	2 7.05	101 €	5.68	2.03	0.87	1.51	2.70	0.50	3.69 (0.77 2	2.36 0.	21 2.	67 0.	35
ol007-hm1	Kistufell	٢	-	IM	0.116 2	876.3	87.9	16.8	648 ¢	\$269 13	39.0 19	0.24 35	0.40 2.	59 16.	.14 2.4	7 6.90	0.95	6.28	2.09	0.81	1.48	2.85	0.41	2.99	0.60 2	2.20 0.	24 2.	24 0.	51
ol013-hm1	Kistufell	13	-	IM	0.113	745.2	84.5	19.8	686 E	479 14	10.7 19	0.61 35	.44 2.	60 16.	.06 2.4;	5 6.27	7 1.26	6.49	2.51	0.86	2.00	2.28	0.51	3.99 (0.71 1	.93 0.	24 2.	16 0.	56
ol015-hm1	Kistufell	15	1	IM	0.042	945.6	96.0	34.2	9 699	205 13	38.6 19	35 35	1.22 2.	60 16.	.64 2.6	1 7.15	5 1.32	5.25	1.96	0.74	1.80	2.51	0.50	2.97	0.63 2	2.14 0	33 2.	11 0.	32
ol016-hm1	Kistufell	16	-	IM	0.124	617.1	88.3	25.5	704 6	360 14	12.5 19	7.7 41	.26 2.	79 16.	55 3.18	8 7.15	5 1.15	6.25	2.04	0.00	2.49	2.55	0.49	3.21 (0.73 1	.95 0.	29 2.	10 0.	30
ol016-hm2	Kistufell	16	2	IM	0.115 8	822.3	98.3	25.5	672 €	571 14	46.8 2C	0.32 41	.00	87 17.	.53 3.00) 7.2€	5 1.25	5.88	2.20	0.87	2.74	3.36	0.47	2.85	0.73 1	0 96.1	32 2.	34 0.	26
ol016-hm3	Kistufell	16	3	IW	0.116 8	853.6	93.2	26.4	688 6	549 14	47.8 2C	0.00 43	34 3.	05 16.	.57 2.78	8 8.50) 1.14	5.77	2.20	0.72	2.12	2.95	0.55	3.37 (0.76 2	2.26 0.	28 2.	34 0.	53
ol016-hm4	Kistufell	16	4	IW	0.118	769.9	88.2	24.3	9 679	567 14	41.9 19	0.77 41	.51 2.	81 16.	.70 2.7	4 7.90) 1.13	6.20	2.22	0.80	1.93	2.90	0.60	3.10	0.66 2	2.14 0	32 2.	40 0.	25
ol019-hm1	Kistufell	19	-	IM	0.123	784.4	88.3	19.0	625 €	385 13	39.7 19	0.11 35	0.46 2.	46 16.	.06 2.39	9 6.23	3 1.04	6.33	2.24	0.88	2.30	2.85	0.51	3.67 (0.76 2	0.06	33 2.	12 0.	61
ol020-hm1	Kistufell	20	-	IM	0.125	762.0	90.4	18.8	646 (5453 14	41.2 19	.53 41	.29 3.	23 16.	.56 2.7:	5 7.45	3 1.18	6.16	1.87	0.89	2.32	3.19	0.67	3.71	0.62 2	2.21 0.	31 2.0	05 0.	00
ol020-hm2	Kistufell	20	5	IM	0.129	112.1	92.7	24.8	693 (5481 14	40.4 20	1.13 41	.23 3.	.03 16.	.27 3.1	3 8.35	5 1.17	6.25	2.15	0.86	1.43	2.99	0.50	2.80	0.82 2	2.05 0.	32 2.	30 0.	26
ol024-hm1	Kistufell	24	1	IW	0.106	349.9 1:	56.8	38.0 1	888 10	1493 It	58.2 27	7.31 84	1.08 7.	.18 42.	.19 6.2:	5 17.02	2 2.35	11.73	3.57	1.27	1.94	4.31	0.56	5.56	6.97 E	3.13 0.	40 2.	90 0.	45
ol024-hm2	Kistufell	24	2	IM	0.100 1	169.5 1	79.9	35.8 2	122 10	925 17	76.7 28	3.37 89	0.40 8.	.67 44.	.90 6.8-	4 18.25	3 2.51	11.66	3.62	1.47	2.06	5.13	0.69	5.18	1.25 3	3.42 0.	42 3.	05 0.	39
ol025-hm1	Kistufell	25	-	IM	0.104	792.9 1	09.4	27.4	805 7	7319 16	51.4 21	.79 46	5.50 3.	52 18.	.28 3.1	4 8.04	4 1.28	7.55	2.68	1.06	3.41	3.35	0.42	4.30	0.76 2	0.77 0.	37 2.0	05 0.	34
ol026-hm1	Kistufell	26	-	IM	0.104	711.1 1	14.3	24.7	837 (861 15	53.7 20).63 44	1.81 3.	.35 20.	.00 3.1:	5 8.21	1 1.20	7.08	2.31	0.92	1.74	3.24	0.59	3.56	0.72 2	2.08 0.	36 2.	56 0.	32
ol026-hm2	Kistufell	26	2	IM	0.110	45.1 1	07.5	25.8	806 (5836 15	50.7 20	.68 45	5.55 3.	.28 19.	53 3.0	7 8.00	3 1.21	6.37	2.33	0.91	2.46	3.19	0.64	2.78	0.81 2	2.30 0.	30 2.	39 0.	32
ol030-hm1	Kistufell	30	-	IW	0.130	803.8 14	02.9	31.5	770 7	7559 16	58.3 22	2.46 44	1.10 3.	59 18.	.88 3.1(0 8.20	0 1.40	7.89	2.82	1.06	1.83	3.62	0.57	3.67	0.90	2.24 0.	41 2.	67 0.	38
ol030-hm2	Kistufell	30	5	IM	0.117	724.7 1:	20.8	30.8	868 7	700 16	56.0 23	3.19 45	0.17 4.	.03 22.	.07 3.2	9 9.25	9 1.48	7.83	2.90	1.11	2.74	3.41	0.67	4.36	1.00	2.51 0.	38 2.	62 0.	36
ol031-hm1	Kistufell	31	-	IW	0.124	795.3 14	08.2	29.3	790 J	7459 16	54.1 21	.91 47	7.21 3.	55 20.	:25 3.1-	4 8.45	8 1.18	6.16	2.29	0.91	2.53	4.31	0.68	3.99	0.79 2	2.70 0.	40 2	38 0.	36
ol036-hm1	Kistufell	36	-	IW	0.106	711.2 1	10.6	24.0	676 7	7006 15	58.1 20	.87 41	.03 3.	.16 18.	.08 3.0	2 8.06	8 1.25	6.27	2.37	0.88	2.00	3.90	0.63	3.93	0.90	2.25 0.	31 2.	30 0.	33
ol041-hm1	Kistufell	41	-	IM	0.126 8	871.0 1	17.3	29.7	822 7	7937 17	72.3 22	2.77 48	3.80 3.	.73 19.	.71 3.20	5 8.64	4 1.31	7.63	2.96	1.08	1.62	3.78	0.67	4.25	0.75 2	2.54 0.	34 3.	01 0.	32
ol041-hm2	Kistufell	41	5	IM	0.128 8	820.3 1	15.2	24.8	833 7	7822 17	72.0 22	2.90 47	7.03 3.	39 20.	.60 3.39	9 8.35	5 1.44	7.79	3.01	1.12	2.41	2.61	0.55	4.15	0.97 2	2.64 0.	45 2.	37 0.	0
ol043-hm1	Kistufell	43	-	IM	0.126 8	863.4 10	9.60	23.3	794 7	751 16	59.6 22	2.55 45	5.51 3.	41 19.	.68 3.3	1 8.10	0 1.31	7.53	2.86	0.99	3.12	3.82	0.58	3.82	0.80 2	2.28 0.	41 2.	34 0.	36
ol045-hm1	Kistufell	45	1	IM	0.112	577.3	96.9	9.2	588 5	5058 20	32.8 14	1.98 25	3.97 1.	.69 15.	.65 1.8:	5 4.65	3 0.67	3.48	1.36	0.77	0.80	2.19	0.45	2.30	0.64	.93 0.	26 2.1	03 0.	35
ol046-hm1	Kistufell	46	-	IM	0.123	896.5 1	10.8	33.8	810 7	7447 16	57.8 22	2.32 46	5.98 3.	.38 19.	35 3.2.	2 8.85	9 1.35	8.19	2.39	1.10	2.35	3.15	0.57	3.82	0.97 2	0.29 0.	35 2.	20 0.	31
ol054-hm1	Kistufell	54	-	IM	0.139	779.8 14	04.7	28.5	765 7	7498 16	56.6 22	2.87 44	1.00 3.	.08 19.	.16 2.8-	4 7.84	4 1.31	7.68	2.32	1.09	2.46	3.26	0.52	4.33	0.85 2	2.56 0.	33 2.	39 0.	Ħ
ol056-hm1	Kistufell	56	-	IM	0.121	776.4 1	19.9	38.1	929 7	524 15	55.4 20	.89 45	0.72 4.	.10 21.	.61 3.49	9 9.25	3 1.54	7.48	3.02	1.00	2.26	3.58	0.58	3.51	0.71 2	2.30 0.	39 2.	49 0.	36
ol057-hm1	Kistufell	57	-	IM	0.120	693.4 1	12.5	26.5	859 (853 15	52.1 20	.76 45	5.73 3.	.35 19.	.28 2.9	5 8.27	7 1.33	6.72	2.50	0.89	1.68	3.57	0.46	3.28	0.75	.87 0.	30 2.	36 0.	30
ol059-hm1	Kistufell	59	-	MI	0.119	164.3	0.66	26.3	876 €	828 15	51.6 19	84 44	1.71 3.	27 19.	36 2.9	1 7.80	1.24	6.50	2.22	0.89	3.01	3.88	0.50	3.85	0.75 2	08 0	37 2	28 0.	28

(continued)
C.1
Table

sample	Eruption	01 I	P.	hase F	H ₂ 0 C	02	ч	5	2	5 H	šr J	Zr	qN	Ba	La	Ce	Pr	S PN	Ē	u ¹⁵⁶ Gd	¹⁵⁷ Gd	Tb	Dy	Но	Er	Tm	Yb	Γn
ol060-hm1	Kistufell	09	1 N	11 0.	120 72	6.7 101.	.7 22.	.8 75	67.	37 150.	9 19.35	40.95	3.15	17.68	2.62	7.10	.24	.55 1.	95 0.8	9 1.64	3.00	0.49	3.63	0.82	2.18	0.36	2.18 0	.25
ol060-hm2	Kistufell	09	N N	11 0	124 71	9.0 110	.4 25.	.6 78	32 67.	30 148.	6 18.8	2 40.57	3.02	17.45	2.56	7.38	60.	.93 2.	14 0.8	7 1.47	3.12	0.52	3.33	0.71	2.06	0.32	5.08	-26
ol062-hm1	Kistufell	62	N N	11 O.	138 79	9.6 99.	.8 32.	2 77.	2 65	38 149.	7 19.92	2 42.88	3.27	19.12	3.20	8.07	3 61.	.98 2.	53 0.8	9 2.25	2.90	0.63	3.37	0.71	2.32	0.36	2.47 (.25
ol064-hm1	Kistufell	49	N N	11 O.	134 73	1.3 87.	.8 19.	.8 79	12 66	12 149.	2 19.18	\$ 44.10	2.82	18.38	2.73	7.43 1	16 6	.97 2.	10 0.9	2 1.02	3.79	0.49	3.52	0.74	2.12	0.38	.83	.33
ol066-hm1	Kistufell	99	N N	11 O.	127 52	5.6 107.	.1 29.	.4 83	12 66	11 147.	2 19.51	43.89	3.28	19.09	3.17	8.02	.22	.03 2.	49 0.9	9 2.19	3.33	0.55	3.48	0.79	1.99	0.32	<u>2.16</u>	34
ol066-hm2	Kistufell	99	2 V	11 O.	122	0.5 105.	.4 25.	.8 84	4 675	97 151.	0 20.7() 43.34	3.35	18.79	2.75	8.46 1	.32 6	.51 2.	81 0.8:	5 1.31	2.25	0.52	3.19	0.74	2.54	0.31	2.31 (.33
ol068-hm1	Kistufell	68	N N	11 0.	125 75	8.5 117.	.3 26.	4 75	3 682	29 150.	9 20.15	3 40.83	2.90	17.31	2.73	7.46 1	.16	56 1.	84 0.8	0 1.66	3.00	0.51	3.08	0.74	2.11	0.35	2.40	.33
ol069-hm1	Kistufell	69	N	11 0.	134 80	9.2 117.	.4 32.	.2 78) 69 61	53 155.	5 20.05	3 45.05	3.17	19.74	3.18	8.22	.24	.11 2.	34 1.0	1 2.38	2.56	0.51	3.58	0.79	2.21	0.31	2.24 (30
ol069-hm2	Kistufell	69	2 V	11 O.	131 12	0.2 123.	.1 29.	.4 79	1 689	33 151.	7 20.75	5 44.02	3.47	18.71	3.15	8.53 1	.25 5	.82 2.	61 1.0	7 1.42	3.51	0.46	4.26	0.80	2.47	0.32	2.27 (.31
ol071-hm1	Kistufell	71	N	11 0.	136 23	0.9 111.	.9 26.	.6 74)69 I;	153.	6 21.65	5 44.75	3.24	20.11	3.27	8.32 1	24 8	.34 2.	57 0.8:	3 1.48	3.24	0.64	3.81	0.86	2.22	0.41	2.33 (.29
ol072-hm1	Kistufell	72	N	11 O.	119 80	12.5 125.	.2 32.	4 93	6 68.	75 156.	0 19.76	5 44.90	3.37	20.64	3.11	8.17	E.	.08 2.	63 0.7	5 2.10	3.16	0.57	3.65	0.91	2.65	0.33	2.14	.33
ol074-hm1	Kistufell	74	N	11 0.	137 72	1.1 109.	.8 19.	.8 75	0 655	36 147.	3 19.52	: 41.91	3.16	16.62	2.78	7.19 1	.14	22 2.	06 0.8	9 3.14	2.86	0.55	3.73	0.78	2.28	0.32	0.09	.31
ol076-hm1	Kistufell	76	N	11 0.	152 75	6.3 112.	.6 29.	2 67	2 635	37 147.	4 19.15	38.69	2.68	16.56	2.64	6.58 (3 86.0	.97 1.	97 0.8′	7 2.31	3.39	0.46	4.20	0.69	2.12	0.28	18.1	.27
ol078-hm1	Kistufell	78	N I	11 0.	146 94	6.0 108.	.3 27.	62 0.	6 672	29 151.	6 20.35	3 43.62	3.08	19.49	3.23	7.94 1	.26	24 2.	03 0.8	5 1.98	3.06	0.50	4.30	0.88	2.35	0.34	2.55 (32
ol078-hm2	Kistufell	78	2	11 0.	144 96	9.5 122.	.2 33.	.5 81	4 685	38 155.	3 20.71	45.41	3.48	21.05	2.92	8.04	.18	08 2.	40 1.0	0 1.81	3.39	0.56	4.55	0.86	2.47	0.33	2.35 (.33
ol078-hm4	Kistufell	78 -	4 X	11 0.	136 92	7.9 129.	5 25.	.8 85	189 68 I	17 153.	9 19.80	1 43.78	3.29	20.57	3.39	7.68 1	.33 5	.96 2.	32 0.8	8 2.27	3.01	0.49	4.10	0.78	2.18	0.39	2.28 (24
ol085-hm1	Kistufell	85	N	11 O.	137 81	5.1 109.	.3 19.	.3 72	5 655	55 146.	3 19.02	39.43	2.89	16.64	2.59	7.05	.17 6	.14 2.	03 0.8	0 2.15	3.54	0.50	4.09	0.70	2.12	0.28	96.1	.33
ol086-hm1	Kistufell	86	N	II 0.	135 88	8.3 117.	4 29.	78 78	671	15 148.	1 19.71	42.53	3.26	18.66	2.69	7.38 1	.23 6	.05 2.	52 0.8	5 1.56	2.98	0.57	3.70	0.73	2.33	0.30	0 (1.79	30
ol086-hm2	Kistufell	86	2 W	U 0.	138 16	1.5 109.	.0 23.	97 6.	09 IU	145.	7 20.01	43.80	3.53	19.03	3.08	8.02	.23 6	.50 2.	14 0.8	9 1.61	3.16	0.61	3.83	0.77	2.53	0.32	2.23 (.26
ol087-hm1	Kistufell	87	N	11 O.	136 82	1.9 118.	2 25.	.8 71	0 664	11 149.	8 20.04	(39.69	2.89	16.43	2.55	6.68	90.	.58 2.	17 0.8	7 2.22	3.41	0.49	4.06	0.67	2.27	0.34	2.45 (31
ol087-hm2	Kistufell	87	N N	11 O.	136 74	4.4 98.	.1 17.	5 75.	3 655)6 146.	0 19.76	43.43	2.92	18.15	2.72	6.64]	.03	85 2.	76 0.8	7 2.88	3.06	0.39	3.95	0.75	2.42	0.35	2.23 (32
ol087-hm3	Kistufell	87	2 N	П 0.	135 79	5.7 116.	4 20.	.5 71	1 655	56 147.	8 19.94	1 38.96	2.50	17.36	2.55	6.92	40.	.63 2.	10 0.9	6 0.62	3.05	0.46	3.48	0.80	2.16	0.25	2.84	36
ol088-hm1	Kistufell	88	2	п 0.	131 92	7.2 117.	0 23.	9 79	8 704	157.	6 20.70	45.47	3.56	20.12	3.38	8.52	33	36 2	54 0.9	7 2.38	3.14	0.56	3.57	0.75	2.18	033	51 (24
ol089-hm 1	Kichtall	08	. 2		147 00	101	27 27	<i>LL</i> 0	- ESS	147	2 10 83	47 56	3 30	10.01	2 08	6 23	10	1	05 08	71.0 1	2 08	0.57	3.48	0.81	100	70.0	12	00
	Vietufall	600	2 2 		174 70	171 6.0	0 r 7 7	- r - r	199 2	.6 145	10.01 2	30.04 0	00.0	12.61	2.12	1 20.0	17.00		0.0 CF	1.7 4	06.7 7 0 0	150	2.60	10.0	12.2	0.25	717	25
11111-06010	Instanton	R S	≤ ;		171 17	111 111		; ;		11 140	1.02 1	01.01	00.0	71.01	0.1.0		07.	10	11 0.0	10.2	1.01	10.0	00.0	1.0	00.7		071	1
ol092-hm1	Kistufell	77 77	2;	11 0 0	130 79	0.2 104	.9 22 22	6 i 0 i	100	43 143. · · · · ·	5 20.14	1 40.49	2.51	17.79	2.62	60.7	4	19	43 0.7	4 2.39	3.23	0.61	3.61	0.66	1.83	0.31	9.19	34
ol095-hm1	Kistufell	95	2	0	.115 80	33.6 103	.7 21	1/ 1/	.6 66.	11 145.	.1 18.82	2 40.35	2.83	17.51	2.61	7.03	4	.03 2.	34 0.8	3 1.58	3.02	0.51	3.46	0.72	1.88	0.31	2.45	-29
ol096-hm1	Kistufell	96	N.	11 0.	118 83	3.2 100.	.4 31.	.7 70	4 64	90 141.	1 19.58	3 40.30	2.95	17.40	2.68	6.59 (.96	.39 2.	11 0.9	1 3.11	3.05	0.46	3.43	0.84	2.16	0.31	5.38	.33
ol097-hm1	Kistufell	67	N N	11 0.	117 72	0.6 114.	.9 38.	.5 79.	0 665	90 147.	5 19.48	\$ 43.04	3.35	18.60	2.93	7.64	.23 5	.90 2.	59 0.97	7 1.76	2.96	0.54	3.63	0.78	2.00	0.32	2.46 (.28
ol101-hm1	Kistufell	101	N	11 O.	130 75	4.6 111.	.4 22.	2 69	15 672	22 146.	3 20.35	\$ 42.87	3.20	17.59	2.96	8.03	.19 6	.74 2.	05 0.9	5 1.42	3.03	0.49	3.62	0.67	2.06	0.28	2.29 (.29
ol102-hm1	Kistufell	102	N N	11 O.	184 11	3.2 95.	.1 72.	4 49	16 500	06 105.	3 16.44	1 36.45	2.66	12.75	2.30	5.88 1	.04 2	.35 2.	02 0.7	7 2.22	2.20	0.47	2.50	0.62	1.59	0.30	69.1	22
ol103-hm1	Kistufell	103	N N	11 0.	169 13	12.2 81.	.8 63.	.0 53	t6 50'.	75 109.	0 16.62	3 38.61	2.77	13.41	2.59	6.73 (5 66.0	.85 1.	78 0.6	9 2.23	2.18	0.46	2.79	0.72	1.84	0.26	.68	20
ol104-hm1	Kistufell	104	N	11 O.	175 74	.8.9 97.	.8 79.	1 47.	3 49;	36 106.	0 15.61	34.64	2.11	12.09	2.37	5.97 (.92 5	.97 1.	42 0.6	9 1.96	2.68	0.45	2.82	0.69	1.67	0.28	99.1	.27
ol105-hm1	Kistufell	105	N N	11 O.	171 58	86.6 100.	.6 508.	.5 62	8 51(7 108.	5 16.44	1 36.86	2.80	13.04	2.49	6.65 (.93 5	.45 1.	63 0.6	7 1.34	2.00	0.42	2.69	0.65	1.73	0.28	1.73	.26
ol106-hm1	Kistufell	106	N N	11 O.	171 72	5.9 104.	.2 103.	.7 50	17 50	34 108.	4 15.96	37.03	3.02	12.92	2.23	6.56 (.92 4	85 2.	05 0.6	7 1.80	2.22	0.41	2.69	0.56	1.99	0.23	.62	.23
ol108-hm1	Kistufell	108	N N	11 O.	175 76	1.0 91.	.3 97.	.6 50	6 518	31 108.	8 17.00	38.40	2.67	13.36	2.33	6.07	01.	.99 1.	85 0.6	3 1.81	1.80	0.39	2.75	0.56	1.98	0.27	1.54 (.18
ol109-hm1	Kistufell	109	N N	11 O.	147	na I	1a I	na n	1 1	n n	a ní	1 na	na	na	na	na	na	na	na n	a na	na	na	na	na	na	na	na	na
ol112-hm1	Kistufell	112	N N	11 O.	172	7.1 118.	.0 87.	.6 51	0 532	28 113.	5 16.71	37.37	2.91	13.30	2.86	6.86 (3 66.0	.61 1.	81 0.6	5 1.91	2.28	0.49	2.99	0.59	1.76	0.27	2.08	21
ol112-hm2	Kistufell	112	2 M	11 0.	180 22	8.8 97.	.3 85.	.7 45	3 49-	11 104.	6 16.2t	35.28	2.31	12.91	2.42	6.53 (.87 4	.69 1.	86 0.6	5 1.33	2.35	0.51	3.03	0.73	1.74	0.23	1.85 (.19
ol113-hm1	Kistufell	113	N N	11 0.	171 81	6.1 105.	.9 68.	2 51	4 51(00 109.	6 16.60	36.87	2.81	11.81	2.39	6.18 (5 76.0	.42 2.	07 0.7	1 1.56	2.04	0.37	3.08	0.71	1.68	0.23	1.75 (22
ol113-hm2	Kistufell	113	2 V	11 0.	172 70	1.1 102.	.2 71.	.3 50	12 515	33 109.	1 16.60	35.73	3.08	12.46	2.27	6.29 1	.01	.10 1.	70 0.5	9 1.12	1.89	0.50	2.95	0.63	1.70	0.26	96.1	.22
ol115-hm1	Kistufell	115	N I	11 0.	171 5	1.1 71.	.2 62.	.1 42	8 422	26 88.	8 13.27	31.62	2.26	10.30	2.05	5.08 (.80 4	.07 1.	55 0.6	7 1.37	1.71	0.35	2.12	0.47	1.47	0.16	1.45 (.23
ol117-hm1	Kistufell	117	N	11 0.	178 67	0.7 87.	.3 66.	.5 50	17 502	25 103.	6 16.22	37.53	2.58	12.27	2.25	6.02 (5 76.0	.39 1.	80 0.6	2 1.81	1.86	0.36	2.99	0.54	1.90	0.21	1.62 (.27
ol118-hm1	Kistufell	118	1 N	11 0.	151	na I	1a I	na n	1 1	n n	ia ni	n na	na	na	na	na	na	na	na n	a na	na	na	na	na	na	na	na	na
ol119-hm1	Kistufell	119	L N	11 O.	182 79	16.1 95.	.6 119.	.8 47	0 502	21 107.	4 16.13	34.53	2.35	12.06	2.32	6.35 (.94	.09 1.	45 0.5	9 1.83	2.08	0.47	2.15	0.56	1.60	0.21	.80	.18
ol119-hm2	Kistufell	119	2 V	11 0.	177	na I	1a I	na n	1 1	п п	a ní	n na	na	na	na	na	na	na	na n	a na	na	na	na	na	na	na	na	na
ol120-hm1	Kistufell	120	N N	11 0.	186 54	.7.6 94.	.1 86.	3 51	0 50	33 107.	0 16.25	36.86	2.59	12.20	2.15	6.42 (4 67.0	.88 1.	83 0.7	3 1.74	2.40	0.42	2.70	0.66	1.67	0.22	06.1	.21

(continued)
C.1
Table

Sample	Eruption	0 I	In Phi	ase H	1 ₂ 0 C(02	СС	I K	ц	Sr	Υ	Zr	Νb	Ba	La	Ce	Pr	S PN	im Eı	1 ¹⁵⁶ Gd	¹⁵⁷ Gd	Tb	Dy	Но	Er	Tm	Υb	Γſ
ol120-hm2	Kistufell	120	2 MI	0.	188 505	8.4 85.	1 73.5	5 429	4613	9.99	14.40	33.80	2.34	10.98	2.37	5.14 0	.74 5	.15 1.	62 0.6	1.22	2.16	0.28	2.53	0.47	1.45	0.18	.47 0	.20
ol120-hm3	Kistufell	120	3 MI	0.	183 12	.2 89.	7 88.7	7 488	5058	106.6	17.00	36.74	2.84	12.89	2.27	5.85 1	00.	.43 1.	7.0 0.7	3 1.71	2.27	0.52	2.92	0.50	1.93	0.26	09.	.27
ol122-hm1	Kistufell	122	1 MI	0.0	137	na n	a na	a na	na	na	na	na	na	na	na	na	na	na	na ni	a na	na	na	na	na	na	na	na	na
ol124-hm1	Kistufell	124	IM	0.	178 97	.0 94.	4 88.8	3 515	5109	110.9	15.98	36.01	2.72	12.78	2.32	5.30 0	96.	17 1.	62 0.7	4 1.82	2.22	0.39	3.05	0.66	1.85	0.25	.94	5
ol126-hm1	Kistufell	126	1 M	.0.	165 (.0 79.	8 74.() 438	4453	93.6	13.95	31.52	2.54	11.09	5.04	5.67 0	86 4	.76 1.	42 0.60	5 1.61	1.87	0.44	2.19	0.48	1.38	0.22	.39 0	.16
ol126-hm2	Kistufell	126	5 WI	0	172 542	2.5 81. 50 20	2 94.3	3 516	4656	98.6	14.41	35.14	2.19	12.71	2.30	6.15 0	87	.51 5. 1.	36 0.60 21 0.60	3 1.97	2.00	0.36	2.41	0.53	1.54	0.22	38	5.6
01128-hm1	Kistufell	871	IW I		SI 771	.66 6.0	5.08	1 490	5140	107.9	76.01	37.61	00.2	12.3/	CS.2	0.40	c 76.	.1 20.	0.03	x 1.80	1.97	0.39	7.80	00.0	1.8/	670	8	87.
ol129-hm2	Kistufell	129	5 MI	0	128	na n	a na	a na	na	na	na	na	na	na	na	na	na	na 22	na na	a na	na	na	na	na	na	na	na	na
ol129-hm1	Kistufell	129	IM		10/ 10/	111.	5 103.5		91c	112.1	16.24	37.26	2.78	13.14	2.04	5.60 1	90. 0	.8.2	89 0.7	IC.I 2	2.11	0.44	3.12	0.67	06.1	15.0	0 69.	77
ol130-hm1	Kistufell	130	IM	0.	156 42	.1 104.	5 85.0	540	5124	109.3	16.22	37.10	2.80	13.24	2.37	5.58 0	.97 5	.06	80 0.7:	5 1.90	1.95	0.41	2.67	0.61	1.97	0.27	.68	.29
ol130-hm2	Kistufell	130	2 MI	(O.	159 115	.9 100.	3 68.3	3 549	5213	112.0	16.84	39.27	2.72	13.32	2.33	5.69 1	.05 6	.09	62 0.7	3 1.03	2.02	0.44	2.67	0.66	1.98	0.29	0 69.	.21
ol131-hm1	Kistufell	131	1 MI	0.	167 668	8.7 95.	8 74.6	5 524	5284	114.5	16.38	38.08	2.69	14.08	2.67	5.82 0	.93 5	.22 1.	86 0.87	7 2.63	2.76	0.46	2.69	0.63	2.04	0.27	.91 0	.25
ol131-hm2	Kistufell	131	2 MI	0.	160 665	.9 101.	1 97.8	3 525	5209	112.7	16.37	38.13	3.10	14.07	2.52	5.72 1	.05 5	.33 1.	76 0.88	8 1.91	2.18	0.50	2.98	0.72	1.84	0.21	0 69	20
ol132-hm1	Kistufell	132	1 MI	0	155 637	.8 98.	4 90.8	8 564	5216	108.1	16.22	38.58	3.03	13.60	2.61	6.32 0	.95 5	.71 1.	92 0.68	3 1.69	2.09	0.44	2.42	0.52	1.86	0.22	.36 0	21
ol132-hm2	Kistufell	132	2 MI	0	151 639	.5 86.	3 85.1	1 525	5005	105.7	16.06	37.23	2.69	12.58	2.67	6.13 0	98 4	.96 2.	10 0.69	9 1.50	2.15	0.38	2.88	0.58	2.13	0.25	.55 0	.26
ol133-hm1	Kistufell	133	1 MI	0	161 347	.5 104.	3 121.1	1 535	5031	107.2	15.56	36.92	2.78	13.56	2.38	6.42 1	.00	.28 1.	41 0.6	3 1.09	1.86	0.39	2.24	0.70	1.52	0.26	.67 0	24
ol134-hm1	Kistufell	134	1 MI	0.	164 671	.8 103.	5 87.2	2 513	4975	105.5	15.74	36.55	2.51	12.39	2.40	5.52 0	89 5	.25 1.	62 0.7	1 1.25	2.11	0.40	2.08	0.57	1.73	0.23	.53 0	22
ol135-hm1	Kistufell	135	1 MI	0	167 207	.4 100.	4 93.3	3 519	5031	106.0	15.69	37.19	2.70	12.10	2.34	6.56 0	5 86.	.05 2.	9.0 0.8	1 1.40	2.23	0.38	2.67	0.55	1.75	0.27	.67 0	.25
ol137-hm1	Kistufell	137	1 MI	0	173 811	.7 93.	5 78.5	5 482	5004	106.2	15.60	36.85	2.33	12.02	2.29	5.17 1	.06	.49 1.	81 0.73	2 0.57	2.16	0.44	2.90	0.56	1.75	0.23	.68	27
ol139-hm1	Kistufell	139	1 MI	0	169 181	3 107	2 116.3	3 1490	10886	123.1	21.39	55 37	414	26.49	4.17	1 12 1	52 7	35 2	76 1.07	7 1.55	2.85	0.58	4.03	0.78	2 42	032	43	28
ol141-hm1	Kistufell	141	M	C	167 146	111	8 76.8	\$ 526	5282	110.5	16.51	37.89	3 00	13.21	74	5.25 1	18	29	48 0.6	3 2.65	2.19	0.50	2.57	0.56	1.74	0.24	48	26
ol142-bm1	Kietufall	142	IM I		140 141	0 85	75/ 0	1 406	4816	101 6	15.40	24.84	2.68	12.04	17 0	5 03 0	- L8	55 1	10 0.60	G I V	100	0.36	2 50	0.53	181	0.00	27 2	2
01142-11111	Instantal	1 1 1			141 240	110.00		+ + + 20	0104	0.101	01 CT	10.10	01.7	10.71	t i	0 03 3	, o		10.0	7.5	17:7	00.0	0000	220	10.1			1.2
01145-nm1	Nsurell	143	I		1/1 348	.4 118.	7 104.	710 0	0000	0.611	1/.10	C7.8C	60.2	0/.71	4.7	0 60.0	6 6	.1.	/9 0.81	17.7	2.1.2	0.54	06.7	00.0	cn.2	0.24	80.	9
ol144-hm1	Kistufell	144	I MI	0.	167 715	.3 105.	20.6	508	5054	107.5	15.95	35.88	2.82	12.30	2.27	5.19 0	.94 5	.45 1.	83 0.6	3 1.01	2.15	0.44	2.73	0.58	1.84	0.23	00	50
ol144-hm2	Kistufell	44	2 MI	0.	162 145	.6 109.	0 92.7	7 513	5274	112.1	16.39	38.10	2.71	12.97	2.25	6.69 1	.13	.88 2.	90 0.7	5 1.58	2.11	0.47	3.03	0.64	2.01	0.22	.81	52
ol145-hm1	Kistufell	145	1 MI	0.0	109 575	2. 72.	1 73.2	2 430	3916	83.7	11.96	29.40	2.20	11.02	1.81	4.64 0	.75 4	.13 1.	28 0.5	5 1.08	1.32	0.31	2.04	0.45	1.31	0.18	.03	.16
ol146-hm1	Kistufell	146	1 MI	10.	170 778	3.3 95.	1 89.3	3 451	5073	107.3	15.57	34.53	2.26	11.26	2.22	5.34 0	.97 4	.58 1.	64 0.66	5 2.10	1.96	0.45	3.01	0.58	1.50	0.24	0 69	21
ol146-hm2	Kistufell	146	2 MI	0.	167 161	1 97.	5 93.4	4 529	5110	109.8	16.28	36.53	2.64	12.05	2.88	6.38 0	.90 5	.83 2.	07 0.79	0.71	2.11	0.43	2.98	0.58	1.70	0.22	02 0	27
ol147-hm1	Kistufell	147	1 MI	0.	156 206	6.1 104.	9 82.5	5 490	5348	112.0	17.33	39.19	2.70	13.12	2.61	6.54 0	.93 6	.48 1.	71 0.80	0 1.88	2.37	0.42	3.20	0.61	1.79	0.26	.94 0	.23
ol148-hm1	Kistufell	148	1 MI	0.	161 854	.99 .99.	5 92.2	2 486	5313	113.3	17.09	37.83	2.69	12.36	2.45	5.59 1	.06 5	.14 1.	91 0.67	7 2.50	2.15	0.47	3.36	0.72	1.87	0.22	0 69.	.25
ol148-hm2	Kistufell	148	2 MI	0.	161 125	6.9 113.	1 90.5	5 513	5399	115.1	17.27	38.60	2.63	13.68	2.69	5.81 1	.11 6	.83 2.	11 0.8	4 1.85	1.81	0.47	2.72	0.73	1.85	0.20	.91 0	.22
ol149-hm1	Kistufell	149	I MI	0.	14	na n	a na	a na	na	na	na	na	na	na	na	na	na	na	na n:	a na	na	na	na	na	na	na	na	na
ol150-hm1	Kistufell	150	1 MI	1 0.	154 961	.1 90.	4 88.5	5 639	6443	116.2	17.17	43.34	2.93	17.84	2.87	7.30 1	.15 6	.11 2.	23 0.87	7 2.19	2.67	0.55	3.30	0.61	1.99	0.20	.61	.23
ol150-hm2	Kistufell	150	2 MI	0.	152 556	6.1 108.	5 89.2	2 528	5113	109.2	16.23	37.08	2.81	12.89	2.47	5.30 0	.86 5	.81 2.	02 0.58	8 1.87	2.34	0.46	3.06	0.62	1.90	0.23	06.	32
ol151-hm1	Kistufell	151	1 MI	0.0	164 203	0. 125.	8 140.8	3 1535	11251	128.1	22.58	57.01	5.10	29.36	4.61 1	0.64 1	.74 8	.98 3.	11 0.99	3.90	3.49	0.60	4.43	0.75	2.44	0.39	233 0	37
ol152-hm1	Kistufell	152	1 MI	10.	173 828	8.6 111.	1 78.3	7 438	5142	109.2	17.18	36.03	2.37	11.54	2.35	6.23 0	.91 5	.80 1.	64 0.63	3 1.56	2.10	0.38	2.41	0.50	1.87	0.26	2.33 0	.25
ol154-hm1	Kistufell	154	1 MI	0.0	167 86	6.4 85.	2.67 7	9 519	5075	106.7	16.16	37.55	2.74	13.18	2.63	6.08 0	.89 5	.02 1.	44 0.6	7 2.32	2.13	0.39	2.37	0.59	1.68	0.22	.96	24
ol154-hm2	Kistufell	154	2 MI	0.	158 745	6.8 114.	5 66.2	2 494	5146	110.1	16.79	37.76	2.34	12.55	2.06	5.89 1	.02 5	.07 1.	59 0.7(0 2.00	1.84	0.43	3.07	0.65	1.88	0.25	.86	.26
ol155-hm1	Kistufell	155	1 MI	0.	163 140	.2 102.	7 87.0	502	5067	108.0	16.15	37.97	2.60	12.88	2.38	5.25 0	92 4	.62 1.	76 0.7	4 2.10	2.54	0.49	2.95	0.64	1.75	0.26	.63	24
ol156-hm1	Kistufell	156	1 MI	0.0	170 821	.1 95.	2.77 7	965 (5159	110.5	16.28	37.32	2.56	12.03	2.19	5.90 0	.84	.96 1.	69 0.7(0.2.09	2.43	0.33	2.26	0.65	1.86	0.25		.18
ol156-hm2	Kistufell	156	2 MI	0.	173 832	.7 102.	2 63.8	8 493	5021	107.6	15.77	36.20	2.91	12.25	2.40	6.17 0	89 4	.00	80 0.66	5 1.48	2.20	0.42	2.85	0.56	1.95	0.25	.68	20
ol157-hm1	Kistufell	157	1 MI	0.	161 758	3.9 101.	5 88.3	3 503	5060	107.6	15.62	37.40	2.90	13.14	2.24	5.26 0	.86 4	.95 1.	89 0.66	5 1.72	2.01	0.37	2.41	0.59	1.79	0.25	.71 0	24
ol158-hm1	Kistufell	158	1 MI	0.0	160 205	0.1 105.	0 89.1	1 497	5528	114.5	18.19	39.81	2.76	13.63	2.65	6.90 1	.02 6	.34 1.	73 0.79	9 1.03	2.49	0.51	2.98	0.66	2.20	0.29	2.23 0	30
ol159-hm1	Kistufell	159	1 MI	.0.	152 706	6.2 103.	8 74.4	4 360	4294	105.9	15.56	29.06	1.81	11.71	2.24	5.85 0	86 4	.88 1.	87 0.7:	5 1.72	1.89	0.37	2.86	0.60	1.59	0.24	.89	.19
ol160-hm1	Kistufell	160	1 MI	10.	160 815	6. 110.	3 101.5	5 493	5121	109.4	16.35	36.37	2.75	12.93	2.27	5.21 1	.03 6	.60 1.	72 0.8	4 2.31	2.27	0.41	2.87	0.59	1.64	0.26		.23
ol161-hm1	Kistufell	161	1 MI	0.	156 592	.9 107.	3 83.7	7 528	5148	107.5	15.71	37.84	2.62	13.26	2.51	6.17 1	.01 5	.54 2.	10 0.6	7 1.59	2.38	0.54	2.89	0.58	1.66	0.28	0.07	.18
ol163-hm1	Kistufell	163	1 MI	0.0	154 831	.2 105.	3.69 (3 512	5095	108.1	15.77	37.08	2.49	12.43	2.68	5.49 0	.92 5	.65 1.	79 0.66	5 2.47	2.36	0.46	3.03	0.62	1.64	0.25	.61	.23

Table C.1 (continued)

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Sample	Eruption	Ю	In P.	hase H	20 C	0_2	F	C	К	ц	Sr	Y	Zr	Nb	Ba	La	Ce	Pr	pN	Sm]	Eu 156	Gd ¹⁵⁷	Gd	Tb	Dy J	но	Er T	ш	(b I	n, I
ol164-hm1	Kistufell	164	1 V	11 0.1	49 16	3.6 95	3.6 9.	5.2 5	538 5	5090 1(36.7 15	5.64 3	7.15 2	2.69 1	3.23 2	2.58 (.58 1	00.	.77 1	.0 69	54	60.	.63 0	45 2.	84 0.	.64 1.	65 0.3	23 1.	52 0.2	
ol165-hm1	Kistufell	165	1 N	1I 0.1	51 100	0.4 97	7.1 8	1.0 5	539 5	5146 1(38.8 16	5.36 30	6.88	2.86 1	3.41	2.52 6	34 0	.93	57 2	03 0.	74 (96.0	.28 0	34 2	72 0.	.64 1.	73 0.	27 1.	32 0.2	2
ol167-hm1	Kistufell	167	1 N	1I 0.1	48 70-	4.9 9(5.1 10	6.7 5	526 5	5131 10	38.2 10	5.48 3	7.03	2.74 1	3.26	2.32	.15 0	98 (.26 1	.04 0.	88	.04	.45 0	51 2	54 0.	.62 1.	68 0.	25 1.	59 O.I	6
ol169-hm1	Kistufell	169	1 V	11 0.1	53 79.	2.6 9-	1.6 6	2.8 4	154 5	5231 1	11.6 1(5.71 3:	5.51 2	2.45 1	2.30	2.42	30 0	.93	.92 1	62 0.	80	.30	.49 0	.41 3.	24 0.	.66 1.	84 0.	26 1.	72 0.2	4
ol170-hm1	Kistufell	170	1 N	4I 0.1	51	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	1 I	ла

Table C.2 Major and volatile element data from Kistufell melt inclusions measured by EPMA. Ol -
olivine number, In - inclusion number, PEC - post-entrapment crystallisation correction (Ol%). All in
wt%.

Sample	Eruption	Ol	In	Phase	SiO_2	TiO_2	$\mathrm{Al}_2\mathrm{O}_3$	Cr_2O_3	FeOt	MnO	MgO	CaO	Na ₂ O	K_2O	$\mathrm{P}_2\mathrm{O}_5$	SO_2	F	Cl	Total	PEC
ol003-hg1	Kistufell	3	-	Gl	48 07	0.98	16.21	0.04	9 17	0.15	9.62	12.83	1.67	0.06	0.08	0.16	0.0212	0.0000	99.08	na
ol003 hg1	Kistufell	4	-	Gl	48.26	0.94	16.32	0.02	9.15	0.21	9.70	12.85	1.71	0.08	0.07	0.16	0.0143	0.0000	99.48	na
ol057-hg1	Kistufell	57	-	Gl	48.87	0.94	16.75	0.04	9.28	0.14	9.82	12.85	1.78	0.06	0.05	0.15	0.0065	0.0000	100.74	na
ol059-hg1	Kistufell	59	-	Gl	48.51	0.97	16.41	0.07	9.14	0.16	9.68	12.78	1.72	0.07	0.07	0.16	0.0232	0.0000	99.76	na
ol060-hg1	Kistufell	60	-	Gl	48.33	0.98	16.55	0.00	9.27	0.22	9.58	12.83	1.72	0.08	0.06	0.16	0.0094	0.0000	99.79	na
ol062-hg1	Kistufell	62	-	Gl	48.74	0.95	16.56	0.07	9.14	0.11	9.70	12.59	1.78	0.07	0.07	0.16	0.0072	0.0000	99.93	na
ol069-hg1	Kistufell	69	-	Gl	48.13	0.93	16.42	0.09	9.04	0.17	9.64	13.12	1.76	0.09	0.04	0.16	0.0241	0.0000	99.62	na
ol071-hg1	Kistufell	71	-	Gl	48.18	0.96	16.46	0.08	9.22	0.16	9.59	12.84	1.73	0.07	0.07	0.16	0.0166	0.0000	99.54	na
01089-ng1	Kistufell	89 00	-	GI	48.52	0.95	16.37	0.06	9.22	0.17	9.75	12.81	1.80	0.06	0.05	0.10	0.0096	0.0000	99.94	na
ol092-hg2	Kistufell	92	-	Gl	48.28	0.96	16.38	0.06	9.02	0.12	9.73	12.81	1.76	0.06	0.07	0.15	0.0141	0.0000	99.43	na
ol103-hg1	Kistufell	103	-	Gl	47.79	0.94	16.52	0.03	9.23	0.13	9.84	12.99	1.77	0.09	0.06	0.16	0.0017	0.0061	99.56	na
ol149-hg1	Kistufell	149	-	Gl	48.29	0.98	16.57	0.05	9.15	0.21	9.94	13.04	1.81	0.10	0.06	0.15	0.0195	0.0041	100.36	na
ol003-hm1	Kistufell	3	1	мі	47.92	0.87	15.90	0.06	9 59	0.14	11.01	12.71	1.68	0.05	0.07	0.21	0.0182	na	100.23	2.26
ol004-hm1	Kistufell	4	1	MI	48.19	0.90	15.96	0.06	9.16	0.10	11.10	12.82	1.59	0.06	0.06	0.20	0.0207	na	100.22	2.08
ol007-hm1	Kistufell	7	1	MI	48.21	0.85	16.13	0.02	9.30	0.17	10.74	12.69	1.75	0.07	0.08	0.20	0.0196	na	100.22	1.62
ol013-hm1	Kistufell	13	1	MI	48.38	0.90	15.98	0.08	9.14	0.16	10.71	12.85	1.67	0.06	0.05	0.21	0.0087	na	100.22	1.78
ol015-hm1	Kistufell	15	1	MI	48.47	0.92	16.97	0.07	9.07	0.18	8.72	13.68	1.77	0.08	0.07	0.21	0.0176	na	100.23	na
ol016-hm1	Kistufell	16	1	MI	48.21	0.92	15.92	0.11	9.44	0.16	10.86	12.62	1.60	0.09	0.06	0.21	0.0000	na	100.21	1.78
ol016-hm2	Kistufell	16	2	MI	48.20	0.94	16.59	0.07	9.05	0.15	10.44	12.60	1.81	0.08	0.06	0.21	0.0211	na	100.23	2.71
ol016 hm4	Kistufell	16	3	MI MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
ol010-hm1	Kistufell	10	1	MI	48 38	0.91	15.81	0.08	8 97	0.15	11 30	12 54	1.68	0.04	0.06	0.22	0.0032	na	100.22	1.96
ol020-hm1	Kistufell	20	1	MI	47.90	0.95	15.85	0.02	9.28	0.19	11.38	12.67	1.61	0.04	0.06	0.20	0.0032	na	100.22	2.49
ol020-hm2	Kistufell	20	2	MI	47.74	0.91	15.26	0.06	9.88	0.23	12.11	12.13	1.58	0.05	0.05	0.19	0.0087	na	100.20	3.56
ol024-hm1	Kistufell	24	1	MI	49.33	1.49	15.38	0.09	8.94	0.14	10.36	11.91	2.00	0.16	0.19	0.22	0.0097	na	100.23	1.02
ol024-hm2	Kistufell	24	2	MI	49.23	1.44	15.45	0.01	9.14	0.16	10.59	11.72	1.92	0.18	0.16	0.22	0.0157	na	100.24	1.50
ol025-hm1	Kistufell	25	1	MI	48.62	0.89	16.58	0.04	8.62	0.22	10.49	12.51	1.89	0.06	0.07	0.21	0.0099	na	100.22	3.36
ol026-hm1	Kistufell	26	1	MI	48.30	0.90	15.70	0.01	9.35	0.15	11.32	12.59	1.52	0.08	0.07	0.21	0.0061	na	100.22	2.66
ol026-hm2	Kistufell	26	2	MI	47.75	0.89	14.94	0.07	9.96	0.16	12.07	12.44	1.59	0.08	0.06	0.18	0.0092	na	100.19	4.11
ol030-hm2	Kistufell	30	2	MI	48.23	0.91	15.91	0.08	8.92 0.25	0.15	11.02	12.07	1.02	0.06	0.06	0.23	0.0187	na	100.25	5.25
ol031_hm1	Kistufell	31	1	MI	48.38	0.94	16.14	0.05	8 78	0.15	10.40	13.30	1.50	0.04	0.00	0.23	0.0217	na	100.25	4 80
ol036-hm1	Kistufell	36	1	MI	47.71	0.89	15.19	0.05	9.77	0.17	11.39	13.19	1.51	0.07	0.05	0.23	0.0176	na	100.23	4.22
ol041-hm1	Kistufell	41	1	MI	49.94	1.14	19.54	0.05	7.77	0.12	3.72	15.60	1.96	0.07	0.09	0.24	0.0191	na	100.26	na
ol041-hm2	Kistufell	41	2	MI	49.86	1.09	19.48	0.01	8.07	0.13	3.72	15.50	1.99	0.08	0.06	0.23	0.0142	na	100.24	na
ol043-hm1	Kistufell	43	1	MI	48.66	0.91	16.87	0.03	8.39	0.16	10.02	13.12	1.72	0.05	0.06	0.23	0.0156	na	100.24	5.06
ol045-hm1	Kistufell	45	1	MI	48.06	0.66	16.45	0.01	9.37	0.14	10.97	12.46	1.78	0.04	0.06	0.13	0.0254	na	100.16	3.38
ol046-hm1	Kistufell	46	1	MI	48.14	0.93	16.21	0.07	8.96	0.14	10.48	13.38	1.58	0.06	0.05	0.23	0.0150	na	100.25	4.71
ol054-hm1	Kistufell	54	1	MI	48.22	0.89	15.94	0.07	8.96	0.13	10.83	13.26	1.59	0.06	0.06	0.22	0.0109	na	100.23	5.08
ol057 hm1	Kistufell	50 57	1	MI	48.44	0.93	16.17	0.05	9.05	0.12	10.59	12.80	1.0/	0.10	0.07	0.22	0.0062	na	100.23	2.01
ol059-hm1	Kistufell	59	1	MI	48.30	0.90	15.97	0.08	9.50	0.15	11.88	12.44	1.74	0.05	0.05	0.20	0.0132	na	100.21	1.85
ol060-hm1	Kistufell	60	1	MI	48.10	0.90	15.99	0.06	9.20	0.17	11.08	12.75	1.68	0.03	0.05	0.20	0.0255	na	100.23	1.99
ol060-hm2	Kistufell	60	2	MI	48.13	0.92	16.14	0.04	9.15	0.15	11.03	12.66	1.65	0.06	0.07	0.21	0.0033	na	100.21	1.88
ol062-hm1	Kistufell	62	1	MI	47.91	0.83	15.64	0.09	9.52	0.12	11.55	12.62	1.60	0.06	0.06	0.21	0.0233	na	100.23	2.98
ol064-hm1	Kistufell	64	1	MI	47.99	0.89	15.97	0.02	9.27	0.14	11.13	12.78	1.67	0.06	0.07	0.21	0.0113	na	100.22	1.67
ol066-hm1	Kistufell	66	1	MI	47.69	0.92	16.39	0.06	9.36	0.18	10.93	12.65	1.69	0.05	0.07	0.20	0.0096	na	100.21	0.73
ol066-hm2	Kistufell	66	2	MI	48.04	0.94	16.12	0.09	9.25	0.15	10.79	12.84	1.65	0.06	0.07	0.21	0.0166	na	100.23	2.70
ol068-hm1	Kistutell	68	1	MI	46.32	1.00	14.84	0.07	10.24	0.17	11.88	13.41	1.97	0.05	0.05	0.19	0.0226	na	100.21	3.41
01069-nm1	Kistufell	69 60	2	MI	48.21	0.91	15.80	0.08	9.11	0.15	11.24	12.00	1.73	0.08	0.04	0.22	0.0122	na	100.23	3.91
ol071-hm1	Kistufell	71	1	MI	48 53	1.00	16.81	0.00	9.53	0.19	8.42	13.58	1.56	0.07	0.00	0.20	0.0231	na	100.21	ч.04 na
ol072-hm1	Kistufell	72	1	MI	48.15	0.90	15.82	0.09	9.53	0.14	10.90	12.68	1.67	0.07	0.07	0.22	0.0098	na	100.23	2.99
ol074-hm1	Kistufell	74	1	MI	48.30	0.88	16.11	0.03	9.04	0.08	11.26	12.52	1.68	0.05	0.05	0.20	0.0116	na	100.22	2.04
ol076-hm1	Kistufell	76	1	MI	48.34	0.89	16.25	0.06	8.89	0.20	10.64	13.00	1.63	0.07	0.04	0.20	0.0154	na	100.22	1.83
ol078-hm1	Kistufell	78	1	MI	47.87	0.89	15.94	0.06	9.66	0.19	11.09	12.53	1.63	0.07	0.07	0.22	0.0200	na	100.24	3.62
ol078-hm2	Kistufell	78	2	MI	48.23	0.91	16.22	0.02	9.29	0.19	10.68	12.49	1.84	0.07	0.05	0.21	0.0260	na	100.24	3.52
ol078-hm4	Kistufell	78	4	MI	48.23	0.88	15.83	0.07	9.52	0.23	10.96	12.55	1.59	0.06	0.08	0.22	0.0138	na	100.23	3.21
ol085-hm1	Kistufell	85	1	MI	48.72	0.96	16.94	0.07	8.92	0.13	9.14	13.29	1.71	0.05	0.07	0.20	0.0220	na	100.23	na
01086 hm2	Kistufell	80 86	1	MI	48.20	0.90	15.04	0.05	9.20	0.15	10.80	12.68	1.80	0.06	0.06	0.21	0.0148	na	100.23	1.4/
ol087-hm1	Kistufell	87	2 1	MI	48.02	0.94	15.54	0.04	9.29	0.15	11.55	12.19	1.00	0.00	0.07	0.19	0.0109	n9	100.21	2.05
ol087-hm?	Kistufell	87	2	MJ	47.86	0.93	15.54	0.03	9.34	0.22	11.62	12.74	1.59	0.05	0.07	0.21	0.0276	na	100.22	3.07
ol087-hm3	Kistufell	87	3	MI	47.88	0.87	15.33	0.05	9.53	0.10	11.85	12.74	1.55	0.04	0.07	0.21	0.0097	na	100.22	3.31
ol088-hm1	Kistufell	88	1	MI	48.14	0.91	15.65	0.06	9.43	0.13	11.06	12.80	1.69	0.06	0.06	0.23	0.0198	na	100.25	3.73
ol089-hm1	Kistufell	89	1	MI	47.86	0.94	15.46	0.03	9.74	0.18	11.37	12.66	1.63	0.08	0.06	0.19	0.0035	na	100.20	3.24
Sample	Eruption	Ol	In	Phase	SiO ₂	TiO ₂	$\mathrm{Al}_2\mathrm{O}_3$	Cr_2O_3	FeOt	MnO	MgO	CaO	Na ₂ O	K ₂ O	P_2O_5	SO_2	F	Cl	Total	PEC
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ol090-hm1	Kistufell	90	1	MI	48.27	0.91	15.81	0.09	9.35	0.19	10.84	12.79	1.63	0.07	0.06	0.21	0.0155	na	100.23	2.70
ol092-hm1	Kistufell	92	1	MI	47.73	0.88	15.27	0.05	9.59	0.18	12.08	12.50	1.60	0.04	0.08	0.20	0.0052	na	100.21	2.88
ol095-hm1	Kistufell	95	1	MI	48.40	0.96	16.43	0.08	8.87	0.15	10.31	12.93	1.74	0.07	0.06	0.20	0.0171	na	100.22	0.94
ol096-hm1	Kistufell	96	1	MI	48.03	0.95	16.29	0.10	9.12	0.18	10.68	12.84	1.70	0.05	0.05	0.20	0.0236	na	100.22	1.27
ol101-hm1	Kistufell	101	1	MI	48.40	0.92	16.11	0.07	9.05	0.15	10.52	12.80	1.77	0.00	0.00	0.21	0.0033	na	100.21	2 79
ol102-hm1	Kistufell	102	1	MI	47.90	0.90	15.94	0.04	9.38	0.19	11.17	12.68	1.67	0.06	0.07	0.19	0.0120	0.0048	100.21	2.32
ol103-hm1	Kistufell	103	1	MI	48.18	0.89	16.10	0.07	9.18	0.19	10.99	12.52	1.72	0.08	0.07	0.19	0.0047	0.0019	100.20	1.53
ol104-hm1	Kistufell	104	1	MI	47.85	0.89	15.90	0.07	9.30	0.16	11.32	12.75	1.62	0.08	0.07	0.20	0.0090	0.0005	100.21	2.29
ol105-hm1	Kistufell	105	1	MI	47.71	0.91	15.75	0.05	9.50	0.18	11.68	12.50	1.62	0.06	0.04	0.19	0.0183	0.0048	100.22	2.93
ol106-hm1	Kistufell	106	1	MI	47.81	0.86	15.90	0.05	9.45	0.21	11.42	12.63	1.57	0.06	0.05	0.21	0.0112	0.0042	100.22	2.65
ol108-hm1	Kistufell	108	1	MI	48.17	0.89	16.01	0.03	9.18	0.19	10.93	12.92	1.54	0.05	0.09	0.22	0.0091	0.0022	100.23	2.42
ol112-hm1	Kistufell	112	1	MI	48.01	0.91	15.86	0.01	9.28	0.13	11.17	12.79	1.70	0.07	0.07	0.14	0.0091	0.0041	100.15	3.61
ol112-hm2	Kistufell	112	2	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
ol113-hm1	Kistufell	113	1	MI	47.98	0.90	16.04	0.06	9.13	0.18	11.21	12.75	1.64	0.06	0.06	0.22	0.0106	0.0026	100.23	1.97
ol113-hm2	Kistufell	113	2	MI	47.99	0.93	16.35	0.08	9.05	0.16	10.77	12.90	1.66	0.06	0.06	0.22	0.0108	0.0037	100.23	1.95
ol115-hm1	Kistufell	115	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
oll17-hml	Kistufell	117	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
ol119-hm1	Kistufell	110	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
ol119-hm2	Kistufell	119	2	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	па	na
ol120-hm1	Kistufell	120	1	MI	47.86	0.92	15.75	0.05	9.32	0.19	11.29	12.87	1.63	0.07	0.05	0.21	0.0167	0.0044	100.23	1.98
ol120-hm2	Kistufell	120	2	MI	45.15	0.89	16.45	0.08	10.06	0.18	12.40	12.95	1.71	0.05	0.06	0.20	0.0147	0.0049	100.22	3.73
ol120-hm3	Kistufell	120	3	MI	47.60	0.90	15.77	0.03	9.39	0.20	11.57	12.83	1.56	0.07	0.08	0.17	0.0065	0.0018	100.18	3.23
ol122-hm1	Kistufell	122	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
ol124-hm1	Kistufell	124	1	MI	47.87	0.96	15.86	0.06	9.35	0.21	10.96	12.96	1.64	0.06	0.07	0.20	0.0128	0.0037	100.21	1.93
ol126-hm2	Kistufell	120	2	MI	43.55	0.98	17 37	3.86	9.88	0.22	12.17	11.28	1.81	0.06	0.05	0.15	0.0000	0.0028	100.13	2.16
ol128-hm1	Kistufell	128	1	MI	47.88	0.90	15.92	0.12	9.53	0.13	10.93	12.90	1.57	0.07	0.05	0.16	0.0062	0.0046	100.17	1.65
ol129-hm2	Kistufell	129	2	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
ol129-hm1	Kistufell	129	1	MI	47.55	0.89	15.81	0.06	9.55	0.17	11.33	12.95	1.56	0.08	0.06	0.20	0.0079	0.0042	100.21	2.66
ol130-hm1	Kistufell	130	1	MI	47.69	0.91	15.72	0.05	9.50	0.17	11.56	12.72	1.55	0.06	0.06	0.17	0.0136	0.0060	100.19	2.56
ol130-hm2	Kistufell	130	2	MI	47.82	0.92	15.85	0.05	9.33	0.21	11.37	12.67	1.66	0.07	0.05	0.19	0.0139	0.0038	100.21	1.90
ol131-hm1	Kistufell	131	1	MI	47.83	0.89	16.00	0.09	9.15	0.19	11.24	12.85	1.60	0.07	0.07	0.22	0.0156	0.0045	100.24	2.74
ol132-hm1	Kistufell	132	1	MI	48.03	0.97	16 30	0.06	8.93	0.21	10.75	12.81	1 80	0.06	0.07	0.22	0.0112	0.0030	100 23	0.85
ol132-hm2	Kistufell	132	2	MI	47.92	0.92	16.19	0.05	9.15	0.17	11.01	12.69	1.73	0.10	0.07	0.21	0.0117	0.0052	100.22	1.28
ol133-hm1	Kistufell	133	1	MI	47.84	0.89	16.15	0.03	9.14	0.15	11.24	12.76	1.64	0.06	0.08	0.13	0.0089	0.0029	100.14	1.68
ol134-hm1	Kistufell	134	1	MI	47.76	0.88	16.23	0.06	9.13	0.18	11.05	12.94	1.66	0.06	0.05	0.21	0.0196	0.0043	100.24	1.44
ol135-hm1	Kistufell	135	1	MI	47.83	0.93	16.29	0.05	9.10	0.15	11.03	12.78	1.71	0.07	0.06	0.19	0.0070	0.0066	100.21	1.34
ol137-hm1	Kistufell	137	1	MI	47.90	0.88	16.28	0.06	9.07	0.17	10.96	12.91	1.65	0.07	0.05	0.20	0.0106	0.0044	100.22	1.46
ol139-hm1	Kistufell	139	1	MI	50.33	0.90	14.60	0.03	8.92 0.44	0.11	10.13	11.28	2.32	0.22	0.10	0.08	0.0154	0.0064	100.10	0.89
ol142-hm1	Kistufell	142	1	MI	+7.70 na	na	na	0.07 na	na	na	na	12.00 na	na	na	na	na	0.0059 na	na	100.15 na	5.27 na
ol143-hm1	Kistufell	143	1	MI	47.77	0.94	15.97	0.12	9.34	0.17	11.05	12.85	1.67	0.07	0.06	0.22	0.0144	0.0011	100.23	3.41
ol144-hm1	Kistufell	144	1	MI	48.03	0.85	15.97	0.02	9.23	0.19	11.08	12.87	1.61	0.07	0.07	0.22	0.0100	0.0029	100.24	2.72
ol144-hm2	Kistufell	144	2	MI	47.93	0.87	15.84	0.09	9.33	0.17	11.21	12.81	1.62	0.07	0.06	0.19	0.0149	0.0018	100.21	2.34
ol145-hm1	Kistufell	145	1	MI	na 17 co	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
ol146-hm2	Kistufell	140	2	MI	47.68	0.92	16.15	0.06	9.05	0.17	11.10	12.14	1.57	0.05	0.05	0.20	0.0154	0.0022	100.22	2.41
ol147-hm1	Kistufell	140	1	MI	47.41	0.90	16.37	0.05	9.28	0.15	11.22	12.82	1.69	0.07	0.07	0.11	0.0114	0.0033	100.13	2.12
ol148-hm1	Kistufell	148	1	MI	47.97	0.88	15.68	0.06	9.30	0.15	11.39	12.95	1.53	0.05	0.05	0.22	0.0089	0.0044	100.23	3.50
ol148-hm2	Kistufell	148	2	MI	49.14	0.86	14.82	0.06	9.24	0.16	11.34	12.81	1.43	0.07	0.07	0.19	0.0041	0.0037	100.20	2.61
ol149-hm1	Kistufell	149	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
ol150-hm1	Kistufell	150	1	MI	48.98	1.17	15.81	0.08	8.60	0.17	10.65	12.44	1.93	0.09	0.08	0.20	0.0164	0.0041	100.22	1.41
ol150-hm2	Kistufell	150	2	MI	47.61	0.91	16.11	0.06	9.19	0.18	11.37	12.78	1.65	0.08	0.07	0.20	0.0000	0.0042	100.21	2.30
ol151-hm1	Kistufell	151	1	MI	49.76	2.00	14.27	0.02	9.60	0.24	10.68	10.84	2.27	0.21	0.10	0.08	0.0114	0.0033	100.09	1.88
ol154-hm1	Kistufell	154	1	MI	47.77	0.92	16.01	0.04	9.59	0.16	10.94	12.85	1.57	0.04	0.07	0.17	0.0111	0.0034	100.18	2.24
ol154-hm2	Kistufell	154	2	MI	48.26	0.93	15.80	0.04	9.34	0.17	10.64	12.99	1.73	0.05	0.07	0.21	0.0170	0.0026	100.23	0.97
ol155-hm1	Kistufell	155	1	MI	47.39	0.93	15.75	0.05	9.59	0.15	11.82	12.59	1.59	0.08	0.06	0.19	0.0063	0.0031	100.20	3.25
ol156-hm1	Kistufell	156	1	MI	47.77	0.90	16.11	0.06	9.12	0.17	11.21	12.91	1.62	0.03	0.10	0.21	0.0151	0.0039	100.22	1.85
ol156-hm2	Kistufell	156	2	MI	48.27	0.90	16.16	0.05	8.99	0.09	11.07	12.82	1.54	0.06	0.05	0.21	0.0154	0.0036	100.23	1.75
ol157-hm1	Kistufell	157	1	MI	47.86	0.91	16.14	0.06	9.14	0.14	11.14	12.89	1.62	0.04	0.06	0.21	0.0173	0.0031	100.23	1.90
01158-hml	Kistufell	158	1	MI	41.18	0.89	15.65	0.08	9.61	0.17	11.38	12.79	1.51	0.08	0.05	0.21	0.0075	0.0030	100.22	4.40 2.96
ol160-hm1	Kistufell	160	1	MI	40.50	0.72	15.50	0.08	9.02	0.17	11.24	13.00	1.61	0.04	0.03	0.21	0.0081	0.0034	100.22	2.00
ol161-hm1	Kistufell	161	1	MI	47.90	0.92	16.32	0.06	9.06	0.22	11.05	12.74	1.63	0.05	0.06	0.21	0.0176	0.0034	100.23	na
ol163-hm1	Kistufell	163	1	MI	47.88	0.90	16.28	0.08	9.04	0.15	10.99	12.93	1.62	0.07	0.06	0.21	0.0213	0.0046	100.24	1.81
ol164-hm1	Kistufell	164	1	MI	48.09	0.90	15.75	0.05	9.15	0.15	11.37	12.77	1.64	0.08	0.06	0.19	0.0164	0.0036	100.21	1.68

Sample	Eruption	Ol	In	Phase	SiO ₂	TiO ₂	Al_2O_3	Cr ₂ O ₃	FeOt	MnO	MgO	CaO	Na ₂ O	K ₂ O	P_2O_5	SO_2	F	Cl	Total	PEC
ol165-hm1	Kistufell	165	1	MI	48.07	0.90	15.72	0.02	9.37	0.14	11.42	12.52	1.69	0.08	0.05	0.17	0.0039	0.0036	100.18	1.91
ol167-hm1	Kistufell	167	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
ol169-hm1	Kistufell	169	1	MI	47.85	0.88	15.98	0.06	9.07	0.16	11.35	13.04	1.51	0.04	0.06	0.21	0.0089	0.0043	100.22	2.17
ol170-hm1	Kistufell	170	1	MI	47.89	0.91	16.02	0.06	9.14	0.15	11.29	12.78	1.62	0.05	0.08	0.22	0.0018	0.0055	100.22	2.33

Table C.3 Major element data from Kistufell host olivines measured by EPMA. Ol - olivine number. All in wt%.

Sample	Eruption	Ol	Phase	SiO ₂	TiO ₂	$\mathrm{Al}_2\mathrm{O}_3$	Cr_2O_3	FeO	MnO	MgO	CaO	P_2O_5	NiO	Total
ol003	Kistufell	3	Ol	40.60	0.021	0.06	0.05	11.58	0.20	47.38	0.34	0.001	0.33	100.55
ol004	Kistufell	4	Ol	40.48	0.002	0.06	0.02	11.29	0.15	48.79	0.32	0.000	0.29	101.38
ol007	Kistufell	7	Ol	41.10	0.004	0.07	0.02	11.61	0.18	47.74	0.30	0.008	0.30	101.33
ol013	Kistufell	13	Ol	40.75	0.000	0.07	0.08	11.38	0.14	47.48	0.33	0.028	0.29	100.56
ol015	Kistufell	15	Ol	na	na	na	na	na	na	na	na	na	na	na
ol016	Kistufell	16	Ol	40.68	0.006	0.07	0.05	11.48	0.16	47.15	0.32	0.014	0.26	100.19
ol019	Kistufell	19	Ol	40.29	0.000	0.07	0.07	10.83	0.18	48.92	0.32	0.000	0.31	100.98
ol020	Kistufell	20	Ol	40.43	0.005	0.06	0.00	11.14	0.21	48.65	0.33	0.000	0.29	101.12
ol024	Kistufell	24	Ol	40.85	0.008	0.06	0.05	11.45	0.20	47.33	0.32	0.117	0.28	100.66
ol025	Kistufell	25	Ol	40.00	0.009	0.09	0.02	11.23	0.18	48.64	0.32	0.000	0.30	100.79
ol026	Kistufell	26	Ol	40.49	0.006	0.06	0.17	11.23	0.22	48.54	0.31	0.000	0.28	101.31
ol030	Kistufell	30	Ol	40.36	0.001	0.06	0.03	11.16	0.17	49.19	0.32	0.000	0.28	101.58
ol031	Kistufell	31	Ol	40.52	0.011	0.06	0.03	11.34	0.14	47.81	0.31	0.006	0.33	100.55
ol036	Kistufell	36	Ol	40.59	0.011	0.07	0.06	11.54	0.20	47.98	0.32	0.000	0.31	101.08
ol041	Kistufell	41	Ol	na	na	na	na	na	na	na	na	na	na	na
ol043	Kistufell	43	Ol	40.71	0.036	0.08	0.03	11.26	0.17	48.02	0.33	0.002	0.30	100.94
ol045	Kistufell	45	Ol	40.14	0.012	0.04	0.02	11.29	0.17	47.18	0.30	0.018	0.29	99.47
ol046	Kistufell	46	Ol	40.19	0.015	0.06	0.03	11.50	0.16	48.00	0.33	0.000	0.31	100.60
ol054	Kistufell	54	Ol	40.23	0.015	0.06	0.03	11.18	0.17	48.14	0.32	0.004	0.31	100.45
01056	Kistufell	56	OI	40.49	0.019	0.05	0.04	11.42	0.20	47.62	0.32	0.000	0.29	100.45
ol057	Kistufell	57	OI	41.23	0.010	0.06	0.02	11.50	0.15	47.92	0.30	0.000	0.28	101.46
ol059	Kistufell	59	Ol	39.64	0.007	0.05	0.03	11.21	0.21	48.75	0.32	0.000	0.29	100.52
01060	Kistufell	60	Ol	41.12	0.000	0.06	0.08	11.14	0.13	47.84	0.31	0.013	0.35	101.03
ol062	Kistufell	62	Ol	40.18	0.013	0.06	0.04	11.27	0.09	48.73	0.33	0.000	0.29	101.01
ol064	Kistufell	64	Ol	40.11	0.005	0.07	0.06	11.37	0.21	48.75	0.32	0.000	0.28	101.17
ol066	Kistufell	66	Ol	40.99	0.000	0.06	0.06	11.46	0.17	47.68	0.31	0.005	0.33	101.05
ol068	Kistufell	68	Ol	41.03	0.035	0.06	0.04	11.54	0.20	47.73	0.31	0.010	0.29	101.24
ol069	Kistufell	69	Ol	40.25	0.005	0.06	0.09	11.12	0.17	48.95	0.33	0.000	0.26	101.22
ol071	Kistufell	71	Ol	na	na	na	na	na	na	na	na	na	na	na
ol072	Kistufell	72	Ol	41.20	0.016	0.06	0.04	11.68	0.17	47.69	0.35	0.004	0.30	101.51
ol074	Kistufell	74	Ol	40.53	0.001	0.06	0.09	11.08	0.17	49.22	0.32	0.000	0.32	101.78
ol076	Kistufell	76	Ol	40.94	0.000	0.07	0.07	11.25	0.16	47.91	0.31	0.000	0.34	101.05
ol078	Kistufell	78	Ol	40.70	0.001	0.07	0.07	11.62	0.14	47.62	0.31	0.055	0.29	100.88
ol085	Kistufell	85	Ol	na	na	na	na	na	na	na	na	na	na	na
ol086	Kistufell	86	Ol	41.16	0.010	0.07	0.03	11.37	0.16	47.65	0.30	0.009	0.34	101.08
ol087	Kistufell	87	Ol	40.20	0.003	0.07	0.08	11.01	0.23	48.79	0.32	0.000	0.30	100.99
ol088	Kistufell	88	Ol	41.13	0.003	0.06	0.06	11.42	0.14	47.73	0.31	0.000	0.30	101.14
ol089	Kistufell	89	Ol	40.92	0.000	0.06	0.01	11.44	0.15	47.67	0.32	0.006	0.29	100.85
ol090	Kistufell	90	Ol	41.20	0.000	0.06	0.07	11.56	0.20	47.82	0.33	0.009	0.28	101.53
ol092	Kistufell	92	Ol	40.70	0.001	0.06	0.04	10.88	0.22	48.89	0.33	0.000	0.33	101.45
ol095	Kistufell	95	Ol	41.21	0.034	0.07	0.07	11.51	0.18	47.64	0.31	0.002	0.33	101.36
ol096	Kistufell	96	Ol	40.69	0.008	0.06	0.07	11.49	0.15	48.02	0.31	0.004	0.29	101.09
ol097	Kistufell	97	Ol	40.84	0.012	0.07	0.07	11.69	0.17	47.58	0.31	0.003	0.29	101.02
ol101	Kistufell	101	Ol	40.77	0.021	0.06	0.03	11.63	0.17	47.86	0.33	0.003	0.31	101.20
ol102	Kistufell	102	Ol	40.85	0.000	0.06	0.05	11.39	0.19	48.34	0.34	na	0.28	101.49
ol103	Kistufell	103	Ol	40.72	0.004	0.05	0.09	11.39	0.21	48.68	0.33	na	0.25	101.73
ol104	Kistufell	104	Ol	40.64	0.007	0.08	0.03	11.21	0.15	48.64	0.34	na	0.34	101.42
ol105	Kistufell	105	Ol	40.40	0.001	0.07	0.06	11.08	0.18	48.57	0.31	na	0.27	100.95
ol106	Kistufell	106	Ol	40.58	0.005	0.07	0.05	11.32	0.18	48.77	0.34	na	0.32	101.62
ol108	Kistufell	108	Ol	40.52	0.004	0.06	0.06	11.51	0.13	48.76	0.34	na	0.31	101.70
ol109	Kistufell	109	Ol	40.88	0.006	0.10	0.10	11.20	0.17	48.61	0.33	na	0.32	101.72
ol112	Kistufell	112	Ol	40.40	0.009	0.07	0.08	11.33	0.17	48.60	0.35	na	0.30	101.29
ol113	Kistufell	113	Ol	40.71	0.007	0.07	0.02	11.14	0.18	48.74	0.32	na	0.32	101.50
ol115	Kistufell	115	Ol	40.75	0.007	0.06	0.04	11.40	0.20	48.35	0.36	na	0.28	101.45
ol117	Kistufell	117	Ol	40.16	0.007	0.07	0.05	11.27	0.16	48.33	0.33	na	0.30	100.68
ol118	Kistufell	118	Ol	39.91	0.000	0.06	0.06	11.31	0.19	48.34	0.32	na	0.30	100.49
ol119	Kistufell	119	Ol	39.64	0.003	0.06	0.02	11.03	0.19	48.51	0.31	na	0.28	100.04
ol120	Kistufell	120	01	39.60	0.006	0.07	0.07	11 17	0.16	48 29	0.33	na	0.27	99.96

Sample	Eruption	Ol	Phase	SiO ₂	TiO ₂	Al_2O_3	Cr_2O_3	FeO	MnO	MgO	CaO	P_2O_5	NiO	Total
ol122	Kistufell	122	Ol	39.44	0.000	0.07	0.03	11.28	0.22	48.20	0.32	na	0.33	99.91
ol124	Kistufell	124	Ol	39.35	0.016	0.06	0.04	11.55	0.16	48.25	0.37	na	0.30	100.10
ol126	Kistufell	126	Ol	39.46	0.003	0.05	0.01	11.35	0.17	48.22	0.32	na	0.30	99.87
ol128	Kistufell	128	Ol	39.42	0.000	0.04	0.03	11.67	0.23	47.75	0.33	na	0.27	99.75
ol129	Kistufell	129	OI	38.99	0.000	0.05	0.04	11.37	0.18	48.11	0.33	na	0.26	99.34
ol130	Kistufell	130	Ol	39.83	0.001	0.05	0.06	11.08	0.13	48.18	0.34	na	0.31	99.99
ol131	Kistufell	131	OI	39.19	0.011	0.07	0.07	11.05	0.20	48.35	0.33	na	0.32	99.59
ol132	Kistufell	132	Ol	39.31	0.005	0.06	0.02	11.24	0.18	48.19	0.32	na	0.33	99.66
ol133	Kistufell	133	OI	39.90	0.009	0.08	0.04	11.05	0.23	48.41	0.34	na	0.32	100.39
ol134	Kistufell	134	Ol	39.95	0.005	0.06	0.06	11.27	0.15	48.57	0.32	na	0.30	100.68
ol135	Kistufell	135	Ol	40.30	0.004	0.06	0.06	11.26	0.14	48.60	0.33	na	0.29	101.02
ol137	Kistufell	137	Ol	39.94	0.006	0.05	0.05	11.24	0.16	48.51	0.33	na	0.27	100.57
ol139	Kistufell	139	Ol	39.98	0.012	0.06	0.05	11.86	0.20	48.01	0.33	na	0.19	100.68
ol141	Kistufell	141	Ol	40.19	0.004	0.06	0.03	11.25	0.23	48.40	0.31	na	0.30	100.77
ol142	Kistufell	142	Ol	40.16	0.005	0.06	0.05	11.09	0.16	48.37	0.33	na	0.26	100.49
ol143	Kistufell	143	Ol	40.22	0.000	0.06	0.03	11.44	0.18	48.24	0.34	na	0.32	100.84
ol144	Kistufell	144	Ol	39.90	0.000	0.06	0.04	11.27	0.19	48.30	0.32	na	0.31	100.39
ol145	Kistufell	145	Ol	39.58	0.003	0.09	0.10	11.43	0.13	48.08	0.33	na	0.31	100.05
ol146	Kistufell	146	Ol	40.60	0.000	0.06	0.07	11.06	0.18	48.61	0.33	na	0.32	101.23
ol147	Kistufell	147	Ol	39.66	0.003	0.06	0.01	11.17	0.22	48.19	0.34	na	0.25	99.91
ol148	Kistufell	148	OI	40.56	0.012	0.05	0.03	11.08	0.26	48.47	0.33	na	0.31	101.10
ol149	Kistufell	149	Ol	40.44	0.003	0.07	0.06	11.17	0.15	48.44	0.32	na	0.28	100.93
ol150	Kistufell	150	Ol	40.58	0.007	0.06	0.06	11.04	0.18	48.72	0.33	na	0.30	101.28
ol151	Kistufell	151	Ol	40.86	0.008	0.05	0.04	12.09	0.28	48.06	0.32	na	0.16	101.86
ol152	Kistufell	152	Ol	40.13	0.006	0.06	0.07	11.17	0.19	48.29	0.33	na	0.30	100.54
ol154	Kistufell	154	Ol	39.42	0.000	0.05	0.06	11.75	0.18	47.78	0.33	na	0.32	99.89
ol155	Kistufell	155	Ol	40.34	0.006	0.06	0.06	11.07	0.17	48.55	0.33	na	0.28	100.86
ol156	Kistufell	156	Ol	40.10	0.000	0.06	0.06	11.05	0.20	48.41	0.32	na	0.34	100.54
ol157	Kistufell	157	Ol	40.41	0.004	0.06	0.02	11.18	0.17	48.65	0.32	na	0.24	101.06
ol158	Kistufell	158	Ol	40.83	0.014	0.05	0.08	11.52	0.18	48.63	0.32	na	0.21	101.83
ol159	Kistufell	159	Ol	40.53	0.003	0.05	0.04	11.00	0.18	48.87	0.34	na	0.27	101.28
ol160	Kistufell	160	Ol	40.16	0.003	0.06	0.02	11.24	0.19	48.63	0.32	na	0.34	100.97
ol161	Kistufell	161	Ol	40.21	0.016	0.06	0.08	11.15	0.15	48.51	0.34	na	0.31	100.84
ol163	Kistufell	163	Ol	40.69	0.010	0.07	0.04	11.23	0.19	48.77	0.32	na	0.31	101.63
ol164	Kistufell	164	Ol	40.38	0.013	0.06	0.02	10.95	0.22	48.57	0.34	na	0.31	100.86
ol165	Kistufell	165	Ol	39.88	0.004	0.07	0.09	11.15	0.19	48.50	0.32	na	0.27	100.49
ol167	Kistufell	167	Ol	39.69	0.007	0.07	0.02	11.56	0.20	48.60	0.31	na	0.28	100.74
ol169	Kistufell	169	Ol	39.91	0.009	0.05	0.05	10.90	0.16	48.55	0.34	na	0.31	100.28
ol170	Kistufell	170	Ol	40.03	0.004	0.06	0.09	10.98	0.18	48.29	0.31	na	0.29	100.22

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Miðfell I	nt4	3 1	IW	450.5	0.065	2.35	64.5	217792	14.3	13.4	128	3793 10	8.2 14	.54 19).22 O.	34 3.1	02 0.6	2 2.16	5 0.49	2.99	1.48	0.55	2.46	1.60	0.35 2	2.73 0.	.60 1.5	57 0.2	1 1.48	0.20	
Miðfell 1	nt4	3 2	IM	513.9	0.063	2.32	67.4	215598	18.0	na	93	3853 5	8.7 14	.31 19	9.20 0.	21 2	33 0.4;	8 2.3	7 0.46	1.85	1.32	0.55	2.20	2.02	0.30 2	2.67 0.	.49 1.7	74 0.2	5 1.40	0.24	
Miðfell	nt4	4	IM	616.4	0.071	2.88	86.6	217983	187.7	240.7	1217	7086 15	8.1 16	.01 59	9.12 5.	96 36.	47 4.6	7 12.7	1 1.72	9.40	2.45	0.93	2.70	2.82	0.42 3	3.17 0.	.53 1.8	80 0.2	0 1.89	0.27	
Miðfell	nt4	4 ·	W ;	384.3	0.069	2.78	75.0	216574	137.6	126.8	. 197	5479 20	15 15 15 15	.98 40).42 3.	83 25. 50	10 3.4	8.4.	2 1.14	0.60	1.94	0.75	1.60	2.33	0.28 2	2.53 0. 27	. 55	74 0.3	1.6	0.28	
Miðfell r	CIII 51	~	IN IN	86.8 86.8	0.063	2.65	89.5	11c022	6.12 8.6	4.22 ha	807 807	1 0064 11 0064	68 16 68 16	51 66. 12 80	.0 0 0.	15 2.	04 0.9. 28 0.4:	2 2.L	10.0 C	474	1.20	0.71	c/-1	1.96	0.40	0 0 33 0	4 99	57 0.7 19	207 1	12.0	
Miðfell I	at5	4	IW	38.2	0.071	2.51	81.3	222653	21.5	29.4	196	1941 12	1.6 13	71 17	.9 06.7	77 6.0	61 0.62	2 2.6(5 0.51	3.15	1.19	0.61	2.38	1.61	0.39 2	2.86 0.	50 1.4	13 0.2	1.40	0.23	
Miðfell I	nt5	4 2	MI	142.0	0.069	2.38	78.2	221079	27.8	17.4	193 4	12 12	0.1 13	.51 18	3.11 0.	63 6.	11 0.99	9 2.8	1 0.58	3.43	1.44	0.52	1.84	1.87	0.43 2	2.25 0.	.47 1.8	33 0.2	4 1.29	0.15	
Miðfell I	nt5	4	IM	247.6	0.076	3.14	102.0	225494	560.6	572.0	3589 1-	1006 15	0.4 29	.43 142	2.35 24.	39 115.	30 17.1:	5 37.6:	5 4.82	22.05	4.74	1.70	8.62	4.38	0.97 5	5.65 1.	.19 3.	50 0.50	0 2.40	0.35	
Miðfell 1	nt5	5 1	III	485.7	0.071	2.62	90.8	220769	292.3	193.1	1747 -	4506 15	12.7 15	.02 44	t.37 9.	59 56.	83 7.3:	5 12.3	7 1.46	13.24	1.65	0.69	2.98	1.86	0.28 2	2.06 0.	59 1.	32 0.2	8 1.6	0.25	
Miðfell 1	nt5	6 1	IW	159.0	0.079	2.52	69.0	220700	16.2	13.8	54	4602 é	9.9 16	.15 25	5.13 0.	10 1.4	44 0.19	9 1.8.	3 0.33	3.44	1.65	0.81	1.79	2.51	0.48 2	2.84 0.	.59 1.7	72 0.2	3 1.85	0.24	
Miðfell 1	nt6	1	III	70.3	0.066	2.34	58.5	220923	17.5	11.0	33	3451 4	3.1 14	.57 15	5.13 0.	11 0.4	64 0.20	0 1.2-	4 0.46	0 2.14	1.21	0.53	1.51	1.85	0.33 2	2.26 0.	.53 1.0	6 0.2	0 1.7	0.23	
Miðfell 1	nt6	2 1	IM	966.7	0.063	2.64	87.6	217861	309.1	200.6	1586	3915 23	(2.0 19	.06 82	2.76 7.	35 49.	12 7.0	7 17.4	8 2.57	. 11.30	3.88	1.35	5.01	3.29	0.49 3	3.55 0.	.80 2.2	23 0.2	9 1.60	0.35	
Miðfell 1	nt6	2	IM	808.1	0.059	2.62	86.0	213030	277.0	391.5	1543	9166 22	1.5 17	.64 71	1.03 7.	20 49	31 7.0	7 17.6:	5 2.15	11.12	2.91	1.27	2.33	2.96	0.59 3	3.37 0.	.77 2.	9 0.2	7 2.3	0.27	
Miðfell 1	nt6	3 1	IM	421.3	0.067	2.28	83.6	211764	23.5	23.0	178 -	11 11	8.1 13	71 07.	7.61 0.	89 6.1	07 0.8;	8 2.9.	1 0.54	3.26	1.25	0.52	1.73	2.04	0.38 2	2.37 0.	53 1.	52 0.2	4 1.28	0.21	
Miðfell 1	nt6	4	IM	783.5	0.055	2.45	120.1	216984	80.3	91.3	479	5793 13	0.0 17	.04 29).74 1.	79 14.	46 2.1.	3 6.2:	5 0.97	. 5.96	1.88	0.64	2.16	2.21	0.34 3	3.35 0.	.71 1.8	87 0.2	4 2.3	0.25	
Miðfell 1	nt6	5 1	IW	317.4	0.067	2.30	77.5	215864	25.4	42.1	173 4	11 11 11	7.7 13	.91 16	5.64 0.	71 5.	77 1.0.	1 3.19	9 0.49	3.78	1.62	0.65	2.09	1.88	0.40 2	2.68 0.	.47 1.5	51 0.2	3 1.4	0.16	
Miðfell 1	nt6	6 1	IW	216.8	0.064	2.35	74.1	218380	26.4	18.3	186 -	1885 11	9.4 13	.96 17	7.78 0.	67 5.4	85 0.85	9 3.1.	3 0.56	4.04	1.69	0.68	1.95	1.22	0.30 2	2.45 0.	.59 1.8	80 0.13	8 1.55	0.18	
Miðfell 1	nt6	7 1	IW	445.9	0.065	2.47	74.5	222353	188.7	172.7	1214	5212 14	2.4 18	.35 55	5.11 6.	50 38.0	00 5.6:	5 12.6	8 1.78	: 8.01	1.95	0.61	2.72	1.72	0.38 3	3.15 0.	.63 1.9	94 0.2	7 1.72	1 0.27	
Miðfell 1	nt6	7 2	IM	706.2	0.064	2.27	79.0	216939	32.0	15.0	187 ,	4649 11	8.4 13	.35 18	3.10 0.	65 5.:	86 0.97	7 2.6	7 0.55	3.09	1.28	0.70	1.72	1.59	0.33 2	2.47 0.	.49 1.4	46 0.2	2 1.19	0.16	
Miðfell I	nt7	1	IM	193.2	0.070	2.49	66.3	222402	17.0	5.3	55	3713 7	6.8 17	.27 10).40 0.	05 1.7	71 0.2	4 0.98	8 0.27	1.78	1.04	0.62	2.42	2.02	0.46 3	3.01 0.	.62 1.6	55 0.3	2 1.40	0.24	
Miðfell I	nt7	1 2	IM	305.9	0.074	2.37	63.7	220405	13.5	12.0	57	3506 5	6.8 17	.25 9).33 O.	12 1.4	85 0.0	7 1.00	0.23	2.31	1.04	0.57	1.86	1.68	0.41 3	3.34 0.	.69 1.6	51 0.3	5 1.6	0.32	
Miðfell 1	nt7	1 3	IW	66.7	0.071	2.44	61.5	221739	7.3	11.4	4	3614 é	9.4 17	.73 8	3.90 0.	07 1.:	53 0.1:	3 0.6	5 0.23	1.67	1.39	0.56	2.11	1.79	0.38 3	3.24 0.	.73 2.0	9 0.2	7 1.74	0.20	
Miðfell 1	nt7	1 4	IM	124.8	0.062	2.63	72.8	226824	10.5	12.4	34	4144 4	3.9 18	.03 22	2.97 0.	0 60	56 0.20	9 1.20	0 0.34	2.13	1.23	0.62	3.06	2.14	0.53 3	3.46 0.	.65 2.2	27 0.2	5 1.87	0.34	
Miðfell 1	nt7	2 1	IM	802.9	0.060	2.33	88.4	214008	38.7	36.3	198 -	1391 11	7.3 13	.11 16	5.57 0.	76 6	26 0.72	8 2.8.	2 0.44	3.38	1.18	0.46	1.83	1.46	0.23 2	2.25 0.	57 1.2	20 0.2	4 1.1	0.15	
Miðfell 1	nt7	2	IW	761.3	0.063	2.33	83.1	210260	32.5	28.4	195 -	4497 12	0.5 12	.78 16	5.98 0.	69 5.	70 0.74	6 3.0:	5 0.43	3.62	1.30	0.62	1.39	1.74	0.30 2	2.06 0.	.56 1.4	13 0.2	4 1.60	0.15	
Miðfell 1	nt7	2	III	751.4	0.064	2.33	83.0	212273	31.0	25.3	180	te57 11	7.8 13	.31 16	5.41 0.	69 5.4	42 0.9	6 2.9.	3 0.47	2.82	1.19	0.56	1.72	1.69	0.36 2	2.48 0.	39 1.	57 0.17	7 1.54	1 0.21	
Miðfell 1	nt7	2	IM	811.9	0.065	2.33	78.1	216037	15.0	28.2	188 .	4437 11	4.0 13	.08 16	5.62 0.	60 5.:	51 0.97	7 2.7.	2 0.60	1 2.95	1.12	0.50	1.24	1.67	0.31 2	2.19 0.	.61 1.	34 0.2	3 1.40	0.18	
Miðfell 1	nt7	3 1	IM	368.3	0.061	3.06	105.9	222056	23.4	41.1	132	7454 5	5.4 19	.01 49).53 0.	41 2.	73 1.28	8 5.35	9 1.14	. 7.76	3.07	1.04	2.87	3.08	0.50 3	3.53 0.	.56 2.	0 0.3	2 2.0	0.33	
Miðfell 1	nt7	3 2	III	771.0	0.065	3.13	57.7	216824	226.6	247.6	1327	7077 15	3.4 18	.53 61	1.42 6.	96 43.0	65 6.1:	5 14.70	0 2.06	9.13	2.47	0.94	2.40	3.01	0.48 3	3.20 0.	.67 2.	l6 0.2	9 1.60	0.25	
Miðfell 1	nt8	1	IM	560.3	0.075	2.39	91.6	222529	9.2	10.1	214	4864 11	8.9 13	.25 17	7.67 0.	79 6.:	55 0.90	0 3.05	8 0.54	3.18	1.45	0.58	1.95	2.32	0.33 2	2.41 0.	.56 1.0	57 0.2	3 1.52	0.15	
Miðfell 1	nt8	2 1	IW	226.9	0.061	2.41	103.3	221577	3.6	2.6	201	5024 11	8.9 14	.23 18	3.00 0.	71 6.	70 0.8	6 2.8	9 0.53	5.19	1.60	0.72	2.16	2.27	0.32 2	0.09 0.	.63 1.8	85 0.2	2 1.6	0.27	
Miðfell 1	nt8	2	IM	477.8	0.058	2.67	86.5	221577	1.7	2.3	85	4931 10	01.0 15	.48 15	5.31 0.	30 2	27 0.7.	2 2.4	7 0.48	3.30	1.29	0.82	2.70	2.72	0.41 3	3.40 0.	.64 2.(9 0.2	5 1.28	0.22	
Miðfell 1	nt8	3 1	IW	1125.0	0.077	2.74	117.9	221577	56.4	53.8	5089	7067 17	5.8 17	.50 58	8.62 7.	43 69.	73 5.5:	5 12.6	5 1.77	76.6	2.33	0.98	2.88	3.81	0.56 3	3.45 0.	57 2.	1 0.2	5 1.89	0.28	
Miðfell 1	nt8	4	IW	47.0	0.055	2.28	82.7	221577	2.0	1.6	201	5014 12	1.7 13	.98 18	8.79 0.	86 6.	72 1.0	7 3.3.	2 0.59	3.49	1.72	0.69	2.73	2.77	0.38 3	3.19 0.	.62	59 0.2I	0 1.8(0.28	
Miðfell 1	nt9	1	IM	221.7	0.071	2.69	96.2	221861	51.8	46.0	. 8791	7626 28	\$0.5 17	.88 62	2.30 8.	85 64.4	84 7.5(0 17.00	5 2.10	9.47	2.49	0.98	1.07	3.32	0.54 3	3.03 0.	.70 1.9	9 0.2	8 2.2	0.25	
Miðfell 1	nt9	2	IM	90.7	0.062	2.46	59.9	231196	1.7	0.7	38	3843 5	0.5 13	.54 25	5.02 0.	07 0.5	93 0.3:	5 1.7(6 0.46	2.75	1.65	0.68	1.99	2.40	0.38 2	2.63 0.	.55 1.5	54 0.2	2 1.0	0.15	
Miðfell 1	nt9	2 2	IM	460.9	0.058	2.40	82.7	224400	4.9	2.6	185 :	5132 12	0.8 14	.57 19	3.0 0.	83 6.	12 0.9.	3 3.0	8 0.56	3.71	1.72	0.74	1.95	2.54	0.49 2	2.26 0.	.66 1.6	54 0.2	2 1.92	0.22	
Miðfell 1	nt9	4	IW	398.9	0.058	2.43	60.1	219291	60.3	62.1	2188 :	3487 17	9.9 15	.78 95	5.60 14.	08 73.4	03 10.8;	8 24.5	9 3.31	15.02	2.85	1.19	0.49	3.30	0.51 3	3.35 0.	.71 1.7	76 0.2	5 1.8	0.25	
Miðfell 1	nt9	4 2	IW	929.6	0.055	2.36	48.9	225405	8.5	8.9	425 (5348 12	0.2 16	.46 28	3.16 1.	95 13.:	50 2.0.	3 5.10	6 0.98	. 4.85	1.83	0.69	1.92	2.22	0.43 2	2.72 0.	.59 1.8	84 0.2	5 1.7	0.23	
Miðfell 1	nt9	6 1	IM	806.6	0.069	2.64	105.7	221577	4.6	3.8	208	4644 11	0.9 15	.69 20).86 0.	84 7	18 1.3(0 3.1;	8 0.61	3.62	1.48	0.58	2.03	1.86	0.40 2	2.30 0.	.56 2.(0.2	5 1.65	0.25	
Miðfell 1	nt9	7 1	IW	768.0	0.060	2.33	86.3	220682	5.0	1.4	191	4613 11	8.7 13	59 17	7.10 0.	81 6	31 0.7;	8 2.8	8 0.54	3.45	1.31	0.69	2.34	1.35	0.41 2	2.35 0.	.54 1.2	26 0.19	9 1.30	0.21	
Miðfell 1	nt9	7 2	IM	693.0	0.040	2.11	62.2	221577	2.8	0.8	57 -	11 686†	2.5 13	.78 12	2.33 0.	14 1	58 0.4	9 1.7;	8 0.42	3.40	1.58	0.58	1.72	1.91	0.40 2	2.38 0.	.66 1.7	77 0.2	3 1.1	0.15	
Miðfell 1	nt9	8	IM	128.2	0.055	2.40	81.5	227959	3.0	0.7	36	1669 5	3.1 17	.07 28	3.08 0.	07 0	59 0.20	6 1.9;	5 0.49	3.74	1.62	0.68	1.65	2.20	0.41 3	3.03 0.	.68 1.8	35 0.2	0 1.58	0.28	
Miðfell 1	nt9 1	10 1	IM	478.8	0.066	2.46	71.0	226990	28.2	29.4	1611	7284 14	17.2 15	.84 59	9.42 6.	05 35.1	68 4.7.	4 12.5	8 2.03	10.16	2.67	1.04	2.54	2.34	0.41 2	2.66 0.	.63 1.7	74 0.2	1 1.50	0.18	
Miðfell	nt9 1	1	IM	662.6	0.061	2.75	120.2	226243	6.2	3.1	206	5878 11	4.2 15	.88 25	5.17 0.	76 6.	97 1.0:	5 3.0	9 0.64	. 4.25	1.77	0.66	1.42	2.41	0.44	3.17 0.	.55 1.0	57 0.2	2 1.72	1 0.26	
Miðfell 1	nt9 1	12 1	IW	395.1	0.055	2.37	65.9	224766	35.9	29.3	1219	5962 14	5.2 15	.71 41	1.46 7.	03 36.	78 4.2	7 8.9.	2 1.28	5.12	1.55	0.68	1.43	1.98	0.40	2.86 0.	.52 1.7	75 0.2	7 1.4	0.24	

(continued)
C.4
Table

Γn	0.20	0.23	0.19	0.25	0.23	0.21	0.19	0.22	0.16	0.18	0.26	0.25	0.27	0.23	0.28	0.22	0.18	0.22	0.29	0.22	0.22	0.24	0.28	0.20	0.24	0.23	0.20	0.23	0.23	0.27	0.31
dY 1	1.62	5 1.46	1.47	ι 1.78	3 1.34	1.55	1.38	1.41	1.55	. 1.42	1 1.46	7 1.63	1.99	. 1.42	2.36	2.07	5 1.64	1.58	2.27	. 1.73	1.54	2.26	1.86	2.12	5 1.64	1.63	1.61	2.12	ι 1.88	1.72	2.17
. Tm	0.21	0.26	0.20	0.24	0.18	0.21	0.25	0.16	0.19	0.22	0.24	0.27	0.29	0.22	0.29	0.21	0.26	0.23	0.30	0.22	0.25	0.22	0.25	0.29	0.26	0.23	0.31	0.21	0.24	0.29	0.29
Er	1.39	1.46	1.95	1.66	1.72	1.77	1.54	1.59	1.69	1.78	2.21	1.72	1.84	1.49	2.32	1.74	1.87	1.47	2.33	1.91	1.69	1.73	1.96	1.78	2.02	1.66	1.54	1.65	2.12	2.10	2.63
Ho	0.43	0.59	0.62	0.61	0.47	. 0.55	0.58	0.39	0.43	0.63	0.61	0.63	0.66	0.53	. 0.73	0.68	0.63	0.71	0.70	0.60	0.59	0.60	0.79	0.74	0.53	0.69	0.54	0.56	0.73	0.68	0.79
Ď	5 2.38	3 2.36	5 2.63	2.82	3 1.92	3.14	3.10	5 2.75	3 2.25	3.35	5 2.80	5 2.29	2.95	1 2.78	5 4.24	2.75	2.66	3 2.92	3.41	3 2.79	2.51	3.56	2.98	5 3.62	3 2.45	3.16	3 2.73	7 3.41	3.40	3.18	3.43
μ μ	0 0.26	6 0.43	8 0.35	1 0.39	0 0.33	7 0.39	7 0.36	3 0.36	3 0.35	4 0.4(4 0.46	8 0.36	1 0.5(8 0.41	6 0.75	1 0.51	5 0.42	3 0.35	8 0.5(3 0.35	9 0.41	1 0.4(4 0.35	9 0.46	3 0.33	2 0.41	3 0.35	1 0.37	1 0.51	1 0.41	0 0.5(
d ¹⁵⁷ G	6 2.0	6 2.6	5 2.0	7 2.8	3 2.0	7 2.0'	6 1.9	4 1.9	8 1.5	5 2.0	6 2.5	7 2.5	4 2.1	5 1.8	1 4.0	7 2.2	9 2.1:	2 2.0	7 2.6	2 2.2	1 2.1	4 2.4	2 2.8	2 3.0	9 2.4	6 2.9	8 2.6	8 2.6	2 3.1	3 3.0	9 3.0
u ¹⁵⁶ G	7 1.3	4.1.4	8 2.0	9 2.3	1 1.5	7 2.3	3 1.9	7 2.4	4 1.0	5 2.4	1 2.3	2 1.9	2 1.8	1 2.2	0 2.9	3 2.2	8 2.3	3 2.0	5 1.1	7 2.2	8 2.1	5 1.6	4 2.8	3 2.9	9 2.1	9 2.7	8 2.2	0 1.7	3 2.2	7 2.7	5 2.3
ш	3 0.6	0.6	5 0.6	17 0.6	9 0.7	4 0.5	6 0.6	9 0.5	0.6	4 0.5	54 0.7	8 0.7	0 0.7	9 0.6	4 1.4	8 0.6	12 0.6	61 0.6	6 0.7	57 0.8	5 0.6	1 0.6	84 0.8	64 0.7	80 0.7	4 0.5	7 0.6	8 0.6	8 0.6	3 0.7	H 0.7
ld Si	7 1.6	11 1.0	0 1.1	1.4	1 1.3	3 1.4	8 1.5	3 1.3	6 1.3	1.1	9 1.5	35 1.6	1.9	1.1	0 3.7	1.1	60 1.4	0 1.6	4 2.6	2 1.6	9 1.5	1 1.8	54 1.8	2 1.6	1.3	84 1.4	1.1 68	8 1.6	4 1.7	7 1.7	2 1.4
Z	4.1	1 2.8	2.7	3.3	1.2.5	3.5	3.0	4 2.2	2.5	2.5	5.6	3.5	4.4	3.4	2 12.6	2.0	2.5	4.0	1 12.4	3.5	3.6	3.2	3.2.5	3.2	3.2	3.2.8	3.6	2.5	5.7	2.7	3.2
e D	4 0.57	2 0.4	5 0.45	2 0.53	3 0.51	3 0.47	3 0.32	3 0.2	4 0.5(3 0.33	9.0.62	0.66	0.71	0.51	8 2.72	3 0.27	5 0.3(5 0.55	3 2.94	1 0.55	3 0.6(8 0.42	5 0.35	3 0.37	5 0.45	3 0.25	5 0.62	0.39	2 1.28	4 0.37	7 0.58
Ŭ	3.02	2.82	2.45	2.62	2.93	2.23	1.78	1.05	2.74	1.73	2.79	3.1(4.2(3.0(20.78	0.8	0.85	3.46	23.03	3.21	2.8	2.48	1.25	1.6	1.96	1.23	2.96	2.29	10.02	1.5^{2}	2.67
La	0.88	0.83	0.66	0.98	0.82	0.50	0.35	0.17	1.09	0.53	0.87	1.05	1.47	1.18	8.57	0.23	0.22	0.91	10.46	1.09	1.25	0.74	0.20	0.35	0.52	0.32	0.97	0.68	5.08	0.38	0.84
Ba	6.25	5.90	5.25	6.11	5.81	4.08	2.61	0.37	5.90	1.40	4.93	6.42	6.26	6.34	62.33	2.00	1.86	7.42	74.46	6.46	6.82	3.57	0.58	1.20	1.34	3.40	6.55	3.46	64.61	0.48	3.32
qN	0.61	0.72	0.56	0.69	0.67	0.32	0.11	0.09	0.74	0.31	0.46	0.92	1.27	0.63	10.68	0.08	0.05	0.86	13.53	0.89	0.87	0.47	0.11	0.11	0.13	0.10	0.95	0.45	8.79	0.11	0.54
Zr	16.86	20.14	19.09	19.48	17.27	23.43	18.90	14.25	17.97	22.54	26.66	19.73	30.20	18.67	85.31	10.36	12.12	19.23	87.16	19.12	19.52	24.27	20.53	23.83	12.43	13.90	18.61	22.26	44.66	24.37	27.79
Y	14.28	14.95	13.85	17.31	13.56	15.71	14.99	14.71	13.77	15.40	16.00	15.34	15.26	13.95	20.40	15.73	15.47	15.84	19.42	15.42	15.33	17.57	17.09	15.16	14.82	16.13	14.17	15.80	15.38	16.77	19.07
Sr	115.1	108.7	101.9	109.8	115.4	100.3	112.5	31.3	114.6	57.5	130.7	125.1	51.1	124.5	187.2	98.9	92.8	135.2	219.2	134.0	132.4	94.9	46.7	62.0	105.4	128.4	124.2	87.8	142.2	47.4	79.0
Ħ	4881	4320	3907	5158	4747	4575	5891	3918	4848	4013	6730	5352	4422	5169	9760	4578	4574	5750	5796	5780	5557	4348	4136	4788	4315	4584	5067	3826	4833	4300	4614
ч	179	192	169	208	192	146	99	27	193	69	181	188	234	211	1826	62	55	219	2271	217	215	148	36	58	43	88	207	149	1995	36	163
³⁷ Cl	8.7	1.7	4.7	2.3	7.2	4.8	1.8	0.7	5.9	21.6	18.1	24.3	29.1	66.3	304.6	7.9	12.5	42.2	342.5	33.0	31.9	11.7	8.5	na	na	0.7	4.8	1.4	46.4	1.4	4.3
³⁵ Cl	3.6	1.7	2.8	4.0	11.6	3.8	1.9	2.1	3.1	17.3	33.3	38.9	55.9	34.2	281.7	16.4	9.4	51.2	401.5	34.6	27.7	16.8	12.2	1.7	1.7	2.2	4.4	2.1	44.0	2.2	4.4
Si	221577	227237	225965	227864	221577	226997	221577	229325	221604	228174	221577	225740	221577	224120	224121	225259	221577	254814	256048	254814	252512	253973	261521	221577	221577	221577	221577	221577	223151	221577	230221
F	89.6	71.1 2	76.9 2	85.2 2	86.8	65.2 2	99.4	67.0 2	91.1	59.4	75.2	103.6 2	100.2	101.3 2	111.7 2	72.6	58.1 2	102.4 2	84.2	97.0	96.8	95.3 2	116.7 2	71.4	143.3 2	115.8 2	103.1 2	95.4 2	6.111	122.4	133.2 2
Ľ	2.37	2.36	2.39	2.88	2.45	2.46	2.44	2.50	2.39	2.31	2.53	2.46	2.65	2.84	2.65	2.43	2.37	2.70	2.81	2.77	2.74	2.70	3.08	2.33	2.58	2.65	2.28	2.38	2.46	2.50	2.71
$\rm H_2O$	0.062	0.060	0.059	0.061	0.071	0.059	0.058	0.056	0.058	0.049	0.041	0.046	0.041	0.041	0.045	0.044	0.043	0.097	0.089	0.078	0.081	0.078	0.081	0.079	0.076	0.075	0.071	0.060	0.071	0.066	0.069
CO_2	740.7	653.7	599.7	896.4	836.7	418.7	154.1	144.9	570.8	268.6	396.5	348.9	852.8	079.3	701.8	316.4	165.3	737.3	507.1	799.2	841.5	776.7	209.2	317.1	310.8	84.9	275.5	333.7	714.1	95.1	323.8
hase	Б	Ē	Б		П	, D	Б	Б	Б	Б	П	E E		-1 -1 -1	Б	Б	Б	Б	- E		Б	D	E	Б	E E	Б	Б	Б	Ē	п	Б
In Pl	1 M	1 M	2 M	1 M	2 M	1 M	1 M	1	1 M	1 M	1 M	1 M	1 M	1 M	1 M	1 M	2 M	1 M	1 M	2 M	1 M	1 M									
ō	13	14	14	15	16	17	18	19	20	21	22	23	24	25	26	27	27	-	0	0	ŝ	4	5	9	-	6	ŝ	4	5	9	٢
Mt	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt9	mt10	mt11	mt11	mt11	mt11	mt11	mt11	mt11						
Eruption	diðfell	Vliðfell	Miðfell	Miðfell	Miðfell	Vliðfell	Miðfell	Vliðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell	Miðfell

Table C.5 Major and volatile element data from Miðfell melt inclusions measured by EPMA. Mt - mount number, Ol - olivine number, In - inclusion number, PEC - post-entrapment crystallisation correction (Ol%). All in wt%.

Eruption	Mt	Ol	In	Phase	SiO_2	TiO ₂	$\mathrm{Al}_2\mathrm{O}_3$	Cr_2O_3	FeO	MgO	CaO	MnO	Na ₂ O	K ₂ O	P_2O_5	F	Cl	SO_2	Fe ³⁺	σ	PEC
Miðfell	mid01a	2	-	Gl	47.64	0.83	15.33	0.02	9.40	9.63	14.10	0.19	1.53	0.010	0.028	0.017	0.002	0.070	0.1698	0.0003	na
Miðfell	mid01a	3	-	Gl	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt2	4	-	Gl	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt3	1	-	Gl	na	na	na	na	na	na	na	na	na	na	na	na	na	na	0.1709	0.0003	na
Miðfell	mt4	1	-	Gl	47.73	0.85	15.40	0.06	9.54	9.65	14.27	0.19	1.52	0.031	0.024	0.009	0.002	0.071	na	na	na
Miðfell Miðfell	mt4	3	-	GI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na 0.1674	na	na
Miðfell	mt5	4	-	GI	11a 47 70	0.83	15.43	0.06	0.46	0.63	14 03	0.17	1 50	0.028	0.024	0.006	0.002	0.071	0.1074	0.0003	na
Miðfell	mt6	7	-	Gl	-17.70 na	na	13.45 na	na	7.40 na	7.05 na	14.05 na	na	na	0.020 na	0.024 na	na	na	na	0.1729	0.0003	na
Miðfell	mt7	2	-	Gl	na	na	na	na	na	na	na	na	na	na	na	na	na	na	0.1710	0.0003	na
Miðfell	mt8	2	-	Gl	47.85	0.83	15.50	0.05	9.36	9.58	14.24	0.20	1.58	0.025	0.026	0.012	0.002	0.074	na	na	na
Miðfell	mt8	3	-	Gl	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt9	3	-	Gl	47.72	0.82	15.17	0.03	9.52	9.71	14.12	0.16	1.45	0.027	0.040	0.015	0.003	0.071	0.1676	0.0003	na
Miðfell	mt9	14	-	Gl	47.58	0.84	15.26	0.06	9.30	9.76	14.08	0.17	1.50	0.029	0.031	0.015	0.003	0.072	na	na	na
Miðfell	mt9	20	-	GI	47.60	0.85	15.29	0.05	9.44	9.67	14.03	0.20	1.51	0.045	0.024	0.013	0.004	0.072	na	na	na
Miðfell	mt9	22	-	GI	47.63	0.80	15.25	0.05	9.40	9.05	14.03	0.17	1.49	0.029	0.020	0.022	0.003	0.072	na	na	na
Miðfell	mt10	5	-	Gl	-17.05 na	na	na	na	7.42 na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt10	6	-	Gl	47.12	0.84	15.56	0.06	9.49	9.49	14.20	0.18	1.54	0.051	0.032	0.006	0.003	0.071	na	na	na
Miðfell	mt11	5	-	Gl	47.54	0.85	15.32	0.05	9.41	9.74	14.05	0.17	1.49	0.038	0.026	0.011	0.001	0.071	0.1725	0.0003	na
Miðfall	mid01a	1	1	мі	40.21	0.72	14.67	0.05	8 20	10.85	14.16	0.15	1 75	0.012	0.022	0.012	0.001	0.150	0.1854	0.0003	4.10
Miðfell	mid01a	1	2	MI	49.51	0.73	14.07	0.03	7.96	10.85	14.10	0.15	1.75	0.012	0.022	0.012	0.001	0.139	0.1854	0.0003	3.00
Miðfell	mid01a	1	3	MI	49.47	0.74	14.81	0.08	8.05	10.55	14.37	0.18	1.72	0.003	0.034	0.008	0.001	0.179	0.1900	0.0003	3.50
Miðfell	mid01a	2	1	MI	48.83	0.76	15.33	0.05	8.32	10.82	14.02	0.15	1.69	0.009	0.033	0.009	0.003	0.151	0.1843	0.0003	3.90
Miðfell	mid01a	2	2	MI	48.76	0.74	15.44	0.05	8.28	10.76	14.11	0.17	1.64	0.013	0.036	0.017	0.001	0.151	0.1885	0.0003	3.90
Miðfell	mid01a	2	3	MI	49.28	0.76	15.03	0.08	8.23	10.69	14.19	0.20	1.48	0.022	0.021	0.001	0.003	0.111	na	na	3.60
Miðfell	mid01a	2	4	MI	47.70	0.77	15.46	0.05	8.63	11.23	14.48	0.20	1.45	0.000	0.028	0.001	0.002	0.124	0.1852	0.0003	4.50
Miðfell	mid01b	1	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell Miðfell	mid01b	1	2	MI	na 47.07	na	na 15 20	na 0.04	na	na 10.75	na	na	na 1 24	na	na	na	na	na	na	na	na 4 20
Miðfell	mt1	1	2	MI	47.07	0.84	15.59	0.04	9.95	10.75	13.18	0.17	1.54	0.235	0.210	0.008	0.013	0.231	0 3720	0 0006	4.50
Miðfell	mt1	2	1	MI	48.90	0.88	15.25	0.07	8.37	10.51	14.05	0.15	1.77	0.024	0.034	0.003	0.001	0.119	0.1881	0.0003	3.80
Miðfell	mt1	2	2	MI	47.30	0.76	14.56	0.05	9.80	12.30	13.51	0.18	1.47	0.034	0.030	0.021	0.003	0.124	0.3720	0.0006	7.70
Miðfell	mt2	3	1	MI	48.46	0.63	14.31	0.18	9.44	10.91	14.21	0.20	1.60	0.026	0.034	0.015	0.009	0.140	0.1916	0.0003	5.00
Miðfell	mt2	4	1	MI	47.33	0.86	15.53	0.05	9.97	10.06	14.38	0.19	1.57	0.028	0.031	0.005	0.002	0.092	0.1681	0.0003	1.80
Miðfell	mt3	1	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	0.1902	0.0003	na
Miðfell	mt3	2	1	MI	na	na	na	na	na	na	na	na 0.17	na	na	na	na	na	na	na	na	na
Miðfall	mt4	1	1	MI	47.08	0.80	13.43	0.07	10.05	10.58	14.22	0.17	1.50	0.016	0.024	0.003	0.004	0.110	0.1776	0.0003	2.80
Miðfell	mt4	1	3	MI	48.22	0.76	14.85	0.00	10.23	10.75	13.54	0.15	1.39	0.130	0.107	0.017	0.008	0.193	0.1910	0.0003	5.20
Miðfell	mt4	1	4	MI	48.02	0.76	14.37	0.09	10.49	11.05	13.60	0.21	1.30	0.059	0.059	0.015	0.006	0.175	na	na	7.20
Miðfell	mt4	1	5	MI	47.88	0.83	14.47	0.09	10.55	11.12	13.41	0.20	1.32	0.074	0.059	0.014	0.005	0.104	0.1780	0.0003	6.10
Miðfell	mt4	2	1	MI	46.70	0.79	15.11	0.05	9.99	12.09	13.53	0.15	1.57	0.016	0.020	0.021	0.007	0.123	0.1792	0.0003	6.00
Miðfell	mt4	3	1	MI	48.28	0.69	14.91	0.05	8.61	10.84	14.86	0.17	1.54	0.016	0.030	0.013	0.002	0.136	0.1873	0.0003	3.80
Miðfell	mt4	3	2	MI	48.05	0.67	14.53	0.10	8.98	11.30	14.76	0.12	1.45	0.012	0.030	0.004	0.001	0.148	0.1870	0.0003	5.30
Miðfell Miðfell	mt4	4	1	MI	48.04	1.27	14.55	0.07	10.07	10.89	12.71	0.19	1.94	0.159	0.124	0.011	0.009	0.230	na 0.1807	na	4.40
Miðfell	mt5	4	2	MI	47.62	0.98	15.07	0.07	9.66	11.10	13.13	0.21	1.62	0.099	0.074	0.000	0.000	0.147	0.1897	0.0003	4.50
Miðfell	mt5	3	1	MI	48.23	0.71	14.65	0.06	10.45	10.84	13.23	0.15	1.66	0.000	0.019	0.007	0.001	0.111	0.1701 na	na	7.10
Miðfell	mt5	4	1	MI	48.14	0.87	15.48	0.06	9.65	9.76	14.32	0.16	1.50	0.019	0.033	0.004	0.001	0.080	na	na	2.00
Miðfell	mt5	4	2	MI	48.00	0.84	15.76	0.07	9.52	9.63	14.39	0.19	1.54	0.028	0.027	0.014	0.002	0.084	0.1749	0.0003	0.20
Miðfell	mt5	4	3	MI	48.12	2.12	13.96	0.04	10.75	10.87	10.97	0.18	2.17	0.451	0.364	0.018	0.028	0.275	0.2026	0.0002	11.10
Miðfell	mt5	5	1	MI	48.27	0.76	14.57	0.05	10.22	10.78	13.16	0.22	1.51	0.246	0.211	0.014	0.012	0.245	na	na	3.90
Miðfell	mt5	6	1	MI	48.13	0.76	14.76	0.10	9.13	11.29	13.97	0.15	1.67	0.013	0.021	0.013	0.001	0.166	0.1870	0.0003	5.60
Miðfell Miðfell	mt6	1	1	MI	48.67	0.60	15.17	0.08	8.31	10.92	14.41	0.16	1.66	0.006	0.015	0.010	0.002	0.131	na	na	3.90
Miðfell	mt6	2	2	MI	48.00	1.54	14.40	0.07	9.62	11.20	12.95	0.16	2.05	0.191	0.149	0.020	0.012	0.265	0.1912	0.0003	4.40 6.40
Miðfell	mt6	3	1	MI	47.12	0.86	15.51	0.07	9.90	10.34	14.50	0.10	1.47	0.029	0.028	0.027	0.001	0.232	0.1914 na	0.0003 na	2.30
Miðfell	mt6	4	1	MI	47.86	1.03	14.92	0.07	10.00	10.65	13.35	0.18	1.83	0.063	0.041	0.014	0.004	0.148	na	na	3.10
Miðfell	mt6	5	1	MI	47.14	0.79	14.87	0.06	10.04	11.33	14.15	0.17	1.42	0.019	0.018	0.015	0.003	0.103	na	na	6.90
Miðfell	mt6	6	1	MI	47.13	0.79	14.42	0.12	9.79	12.75	13.39	0.19	1.37	0.025	0.030	0.007	0.002	0.100	na	na	10.90
Miðfell	mt6	7	1	MI	48.21	0.91	14.89	0.08	8.58	11.48	13.78	0.20	1.63	0.139	0.121	0.018	0.009	0.211	0.1895	0.0003	4.60
Miðfell	mt6	7	2	MI	47.13	0.77	14.64	0.06	9.66	12.93	13.21	0.17	1.41	0.013	0.025	0.008	0.001	0.119	0.1811	0.0003	7.70
Miðfell	mt7	1	1	MI	48.87	0.66	15.22	0.06	8.26	11.13	13.85	0.14	1.79	0.003	0.018	0.016	0.001	0.111	0.1803	0.0003	4.00
Miðfall	mt7	1	2	MI	48.90	0.64	15.00	0.15	8.1/ 8.20	11.01	14.11	0.15	1.88	0.000	0.008	0.006	0.001	0.115	0.1810	0.0003	3.60
Miðfell	mt7	1	4	MI	49.12	0.68	14.89	0.05	8.35	11.23	13.72	0.17	1.74	0.017	0.022	0.007	0.001	0.173	0.1821	0.0003	4.60
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Eruption	Mt	Ol	In	Phase	SiO ₂	TiO ₂	Al_2O_3	Cr ₂ O ₃	FeO	MgO	CaO	MnO	Na ₂ O	K ₂ O	P_2O_5	F	Cl	SO ₂	Fe ³⁺	σ	PEC
Miðfell	mt7	2	1	MI	47.41	0.79	15.64	0.06	9.92	10.08	14.28	0.20	1.56	0.015	0.037	0.025	0.001	0.116	0.1818	0.0003	1.70
Miðfell	mt7	2	2	MI	46.95	0.84	15.70	0.07	10.08	10.24	14.33	0.16	1.57	0.025	0.014	0.013	0.002	0.120	na	na	1.50
Miðfell	mt7	2	3	MI	46.81	0.82	15.16	0.06	10.49	10.67	14.36	0.17	1.43	0.019	0.025	0.007	0.000	0.127	na	na	6.80
Miðfell	mt7	2	4	MI	47.53	0.81	15.84	0.10	9.75	9.91	14.31	0.19	1.52	0.024	0.017	0.004	0.002	0.116	0.1815	0.0003	0.90
Miðfell	mt7	3	1	MI	47.96	1.18	13.17	0.03	12.33	11.18	12.03	0.22	1.81	0.021	0.067	0.015	0.002	0.243	0.3909	0.0006	6.10
Miðfell	mt7	3	2	MI	47.16	1.17	13.05	0.01	12.83	11.63	11.74	0.24	1.90	0.160	0.116	0.019	0.009	0.235	0.3929	0.0006	6.90
Miðfell	mt8	1	1	MI	47.81	0.80	15.54	0.05	9.44	10.44	14.22	0.18	1.45	0.045	0.035	0.000	0.003	0.121	na	na	2.20
Miðfell	mt8	2	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt8	2	2	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt8	3	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt8	4	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt9	1	1	MI	47.82	1.26	13.83	0.02	11.77	10.69	12.13	0.19	1.94	0.226	0.132	0.024	0.017	0.369	0.1940	0.0003	4.20
Miðfell	mt9	2	1	MI	49.80	0.67	15.33	0.11	8.78	10.04	13.36	0.19	1.65	0.023	0.036	0.030	0.002	0.179	0.1781	0.0003	3.10
Miðfell	mt9	2	2	MI	47.82	0.62	15.44	0.06	9.62	10.98	13.30	0.21	1.81	0.016	0.122	0.006	0.001	0.131	na	na	8.00
Miðfell	mt9	4	1	MI	47.51	1.41	14.98	0.09	8.88	11.03	13.80	0.17	1.64	0.265	0.237	0.017	0.021	0.257	0.1900	0.0003	3.80
Miðfell	mt9	4	2	MI	48.54	1.10	14.83	0.06	8.71	10.82	14.15	0.13	1.55	0.065	0.059	0.002	0.002	0.151	0.1784	0.0003	3.10
Miðfell	mt9	6	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt9	7	1	MI	47.54	0.79	15.43	0.06	9.62	11.18	13.62	0.19	1.51	0.017	0.032	0.012	0.002	0.119	0.1782	0.0003	4.00
Miðfell	mt9	7	2	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Midfell	mt9	8	1	MI	48.78	0.77	14.53	0.08	7.90	11.26	15.04	0.11	1.49	0.009	0.034	0.020	0.003	0.197	na	na	6.90
Midfell	mt9	10	1	MI	48.54	1.26	14./1	0.06	9.30	10.36	13.70	0.17	1.64	0.135	0.119	0.020	0.010	0.240	0.1932	0.0003	8.80
Midfell	mt9	11	1	MI	48.18	1.10	14.58	0.06	10.29	10.52	13.27	0.22	1.61	0.033	0.064	0.014	0.001	0.145	na	na	7.10
Mioren	mt9	12	1	MI	48.54	0.88	14.85	0.09	8.05	11.57	13.39	0.15	1.42	0.234	0.223	0.024	0.015	0.213	0.1880	0.0003	5.40
Miðfall	mt9	13	1	MI	10 14	na 0.72	14 90	na	na	10.20	na	na	na 1 47	na	na	na	na	na	na	na	na 2 20
Miðfall	mt0	14	2	MI	49.14	0.72	14.60	0.07	9.14	10.50	14.10	0.20	1.47	0.017	0.042	0.008	0.002	0.127	0.1820	0.0005	2.50
Miðfall	mt0	14	1	MI	40.00	0.75	14.95	0.05	9.51	10.50	12.59	0.10	1.42	0.027	0.025	0.002	0.001	0.127	na	na	2.40
Miðfall	mt0	15	2	MI	48.80	0.72	14.57	0.28	9.42	10.94	15.56	0.15	1.47	0.050	0.020	0.015	0.001	0.142	0 1765	0.0003	5.10
Miðfell	mt0	17	1	MI	18 74	0.78	15.13	0.08	7.88	10.63	15.06	0.17	1.46	0.032	0.035	0.010	0.002	0.167	0.1705	0.0005	5 00
Miðfell	mt9	18	1	MI	40.74 na	0.70 na	10.15 na	na	7.00 na	10.05 na	15.00 na	na	na	0.052 na	0.055 na	na	0.002 na	na	na	na	5.00 na
Miðfell	mt9	19	1	MI	49.11	0.66	14 72	0.26	8 4 2	10.69	14 34	0.17	1 59	0.016	0.033	0.014	0.000	0.138	0 1836	0.0003	6.00
Miðfell	mt9	20	1	MI	47.61	0.78	14.69	0.06	9.80	11.70	13.73	0.20	1.37	0.025	0.028	0.015	0.001	0.117	0.1841	0.0003	6.60
Miðfell	mt9	21	1	MI	49.16	0.67	15.15	0.09	7.82	10.95	14 44	0.18	1 44	0.040	0.047	0.000	0.002	0.145	0 1811	0.0003	3 30
Miðfell	mt9	22	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt9	23	1	MI	47.63	0.78	13.99	0.03	9.91	12.61	13.56	0.11	1.31	0.038	0.030	0.012	0.007	0.103	na	na	14.50
Miðfell	mt9	24	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt9	25	1	MI	48.37	0.73	14.96	0.04	8.60	11.26	14.48	0.13	1.38	0.042	0.021	0.000	0.010	0.131	0.1825	0.0003	4.10
Miðfell	mt9	26	1	MI	48.46	1.49	13.69	0.02	11.16	10.64	11.94	0.16	2.02	0.243	0.185	0.014	0.014	0.266	0.1954	0.0002	4.80
Miðfell	mt9	27	1	MI	48.60	0.72	15.66	0.09	7.86	10.99	14.35	0.17	1.49	0.047	0.016	0.013	0.001	0.075	0.1768	0.0003	4.00
Miðfell	mt9	27	2	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	0.1798	0.0003	na
Miðfell	mt10	1	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt10	2	1	MI	48.26	0.86	14.91	0.05	8.49	11.59	13.76	0.12	1.53	0.233	0.184	0.015	0.017	0.263	0.1937	0.0003	5.20
Miðfell	mt10	2	2	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt10	3	1	MI	47.61	0.84	15.60	0.08	9.66	10.41	14.08	0.17	1.50	0.022	0.028	0.000	0.001	0.121	0.1792	0.0002	2.10
Miðfell	mt10	4	1	MI	47.20	0.78	14.10	0.06	9.76	13.38	13.09	0.23	1.35	0.030	0.019	0.007	0.001	0.124	0.1829	0.0003	11.70
Miðfell	mt10	5	1	MI	48.77	0.67	14.51	0.06	10.32	10.61	13.19	0.16	1.71	0.005	0.014	0.007	0.001	0.097	na	na	6.10
Miðfell	mt10	6	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt11	1	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt11	2	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt11	3	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt11	4	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt11	5	1	MI	47.66	0.74	14.58	0.12	10.64	10.90	13.51	0.18	1.32	0.149	0.199	0.014	0.010	0.185	na	na	6.70
Miðfell	mt11	6	1	MI	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Miðfell	mt11	7	1	MI	49.46	0.76	15.02	0.07	7.22	11.39	14.28	0.14	1.59	0.018	0.052	0.021	0.004	0.173	0.1835	0.0003	4.10

Emution	M	01	Dhasa	5:0	TO	41.0	C= 0	NIO	E-O	MaQ	C=O	MaQ
Eruption	WIt	01	rnase	3102	1102	Al ₂ O ₃	CI203	NIO	reo	MgO	CaO	Willo
Miðfell	mid01a	1	01	39.77	0.007	0.10	0.09	0.31	10.34	48.26	0.36	0.15
Miðfell	mid01a	2		40.48	0.008	0.11	0.10	0.30	10.40	48.22	0.30	0.17
Miðfell	mid01a	3	01	40.11	0.008	0.07	0.08	0.28	11.09	47.75	0.36	0.19
Miðfell	mid01b	1	OI	na	na	na	na	na	na	na	na	na
Miðfell	mt1	1	Ol	40.61	0.008	0.06	0.05	0.24	12.18	47.00	0.35	0.21
Miðfell	mt1	2	Ol	40.54	0.005	0.09	0.09	0.28	10.74	48.06	0.37	0.18
Miðfell	mt2	3	Ol	39.86	0.010	0.09	0.08	0.29	11.41	46.93	0.38	0.17
Miðfell	mt2	4	01	39.82	0.007	0.05	0.03	0.21	12.80	45.97	0.37	0.20
Miðfell Miðfell	mt2	4		39.82	0.007	0.05	0.03	0.21	12.80	45.97	0.37	0.20
Miðfell	mt3	2		na	na	na	na	na	na	na	na	na
Miðfell	mt4	1	01	39.76	0.006	0.07	0.04	0.25	12.36	46.40	0.34	0.19
Miðfell	mt4	1	OI	39.76	0.006	0.07	0.04	0.25	12.36	46.40	0.34	0.19
Miðfell	mt4	2	Ol	39.91	0.004	0.08	0.06	0.31	11.00	47.49	0.37	0.17
Miðfell	mt4	3	Ol	40.37	0.008	0.08	0.08	0.30	10.61	47.58	0.38	0.16
Miðfell	mt4	3	Ol	40.37	0.008	0.08	0.08	0.30	10.61	47.58	0.38	0.16
Miðfell	mt4	4	01	39.94	0.008	0.06	0.05	0.26	12.13	46.80	0.35	0.18
Miðfell	mt4	4		39.94	0.008	0.06	0.05	0.26	12.13	46.80	0.35	0.18
Miðfell	mt5	3		39.65	0.008	0.07	0.00	0.20	12 50	46.25	0.30	0.20
Miðfell	mt5	4	01	39.49	0.010	0.05	0.04	0.24	12.74	45.97	0.34	0.20
Miðfell	mt5	5	OI	39.22	0.007	0.06	0.04	0.24	12.31	46.25	0.36	0.21
Miðfell	mt5	6	Ol	40.05	0.008	0.09	0.07	0.29	10.80	47.60	0.39	0.19
Miðfell	mt6	1	Ol	40.19	0.012	0.10	0.08	0.35	10.24	47.97	0.36	0.18
Miðfell	mt6	2	Ol	40.88	0.010	0.06	0.05	0.28	11.82	47.18	0.37	0.17
Miðfell	mt6	3	01	39.64	0.003	0.07	0.03	0.23	12.46	46.40	0.37	0.21
Miðfell	mt6	4		39.23	0.010	0.05	0.03	0.22	12.27	46.51	0.36	0.20
Miðfell	mt6	6	01	39.29	0.003	0.07	0.00	0.30	10.35	48.04	0.35	0.19
Miðfell	mt6	7	01	39.05	0.003	0.09	0.08	0.34	10.07	48.06	0.35	0.16
Miðfell	mt6	7	OI	39.05	0.003	0.09	0.08	0.34	10.07	48.06	0.35	0.16
Miðfell	mt7	1	Ol	39.88	0.005	0.09	0.09	0.35	10.02	48.13	0.35	0.17
Miðfell	mt7	2	Ol	39.75	0.017	0.06	0.04	0.22	12.76	46.21	0.34	0.22
Miðfell	mt7	3	Ol	39.61	0.011	0.06	0.02	0.21	13.99	45.25	0.35	0.22
Miðfell	mt8	1	01	40.52	0.004	0.10	0.05	0.28	11.92	47.07	0.36	0.18
Miðfell	mt8	2		na	na	na	na	na	na	na	na	na
Miðfell	mt8	4	01	40.24	0.004	0.05	0.04	0.24	12.41	46.54	0.38	0.21
Miðfell	mt9	1	OI	40.26	0.007	0.06	0.01	0.20	14.10	45.63	0.36	0.21
Miðfell	mt9	2	Ol	40.59	0.008	0.10	0.06	0.27	11.59	47.23	0.34	0.18
Miðfell	mt9	3	Ol	41.04	0.006	0.08	0.09	0.33	10.42	48.36	0.37	0.17
Miðfell	mt9	4	01	40.91	0.009	0.07	0.08	0.31	10.87	48.10	0.37	0.17
Miðfell	mt9	0		40.60	0.007	0.07	0.05	0.22	12.78	46.40	0.35	0.17
Miðfell	mt9	8	01	41.02	0.011	0.10	0.09	0.30	9.60	47.23	0.39	0.20
Miðfell	mt9	10	01	40.78	0.008	0.05	0.05	0.26	11.89	47.19	0.37	0.20
Miðfell	mt9	11	Ol	40.48	0.011	0.06	0.04	0.22	12.78	46.64	0.35	0.19
Miðfell	mt9	12	Ol	40.93	0.005	0.08	0.06	0.32	10.31	48.33	0.34	0.15
Miðfell	mt9	13	Ol	40.62	0.008	0.08	0.07	0.24	12.54	46.56	0.36	0.20
Miðfell	mt9	14	01	40.69	0.005	0.09	0.07	0.27	11.70	47.00	0.36	0.20
Miðfell	mt9	14		40.69	0.005	0.09	0.07	0.27	11.70	47.00	0.30	0.20
Miðfell	mt9	17	01	40.72	0.009	0.11	0.08	0.32	9.99	48.13	0.38	0.15
Miðfell	mt9	18	Ol	40.48	0.005	0.06	0.07	0.25	12.19	46.61	0.36	0.20
Miðfell	mt9	19	Ol	40.56	0.006	0.10	0.10	0.33	10.52	47.70	0.38	0.15
Miðfell	mt9	20	Ol	40.76	0.007	0.11	0.09	0.29	11.13	47.41	0.35	0.18
Miðfell	mt9	20	Ol	40.76	0.007	0.11	0.09	0.29	11.13	47.41	0.35	0.18
Miðfell	mt9	21	01	41.09	0.007	0.09	0.08	0.34	9.68	48.32	0.36	0.14
Miðfall	mt9 mt0	22		40.86	0.008	0.07	0.04	0.26	11.79	47.30	0.36	0.20
Miðfell	mt9	23	01	40.92	0.008	0.07	0.04	0.20	10.61	48.08	0.36	0.20
Miðfell	mt9	23	OI	40.92	0.007	0.09	0.07	0.31	10.61	48.08	0.36	0.20
Miðfell	mt9	24	Ol	40.53	0.012	0.07	0.05	0.22	12.71	46.25	0.37	0.19
Miðfell	mt9	25	Ol	40.88	0.003	0.07	0.08	0.31	10.29	48.03	0.39	0.15
Miðfell	mt9	26	Ol	40.47	0.011	0.05	0.01	0.17	13.44	45.79	0.36	0.23
Miðfell	mt9	27	01	41.06	0.009	0.10	0.10	0.38	9.78	48.74	0.36	0.16
Miðtell Miðfell	mt10	1		40.21	0.009	0.10	0.07	0.26	12.43	46.36	0.35	0.17
Miðfell	mt10	23	01	40.20	0.009	0.08	0.08	0.50	7.83 12.06	46 34	0.55	0.17
Miðfell	mt10	4	01	40.70	0.007	0.11	0.11	0.34	9.82	47.96	0.37	0.16

 $\textbf{Table C.6} \ \text{Major element data} \ (\text{wt\%}) \ \text{from Midfell olivines measured by EPMA. Ol-olivine number.}$

Eruption	Mt	Ol	Phase	SiO_2	${\rm TiO}_2$	$\mathrm{Al}_2\mathrm{O}_3$	$\mathrm{Cr}_2\mathrm{O}_3$	NiO	FeO	MgO	CaO	MnO
Miðfell	mt10	5	Ol	40.15	0.012	0.08	0.05	0.23	12.54	46.04	0.36	0.21
Miðfell	mt10	5	Ol	40.15	0.012	0.08	0.05	0.23	12.54	46.04	0.36	0.21
Miðfell	mt10	6	Ol	40.52	0.003	0.09	0.09	0.32	10.02	47.98	0.37	0.14
Miðfell	mt10	6	Ol	40.52	0.003	0.09	0.09	0.32	10.02	47.98	0.37	0.14
Miðfell	mt11	1	Ol	40.28	0.005	0.08	0.09	0.27	11.54	46.88	0.39	0.19
Miðfell	mt11	2	Ol	40.73	0.006	0.08	0.04	0.27	11.44	47.41	0.39	0.18
Miðfell	mt11	3	Ol	40.47	0.007	0.06	0.04	0.26	12.11	47.04	0.37	0.21
Miðfell	mt11	4	Ol	41.07	0.006	0.09	0.10	0.31	9.71	48.77	0.39	0.17
Miðfell	mt11	5	Ol	40.61	0.009	0.06	0.03	0.27	12.72	46.55	0.36	0.20
Miðfell	mt11	5	Ol	40.61	0.009	0.06	0.03	0.27	12.72	46.55	0.36	0.20
Miðfell	mt11	6	Ol	40.82	0.006	0.10	0.09	0.30	10.93	47.98	0.39	0.18
Miðfell	mt11	7	Ol	41.39	0.007	0.11	0.14	0.39	8.83	49.60	0.36	0.14

Table C.7 CO_2 standards for Kistufell and Miðfell SIMS sessions. No. - analysis number. CO_2 known in ppm, isotopes in cps.

Standard	Month	Day	No.	²⁴ Mg/2	¹² C	²⁸ Si/2	⁴⁰ Ca/2	³⁰ Si	CO ₂ known
17-2	Apr14	1	1	1342	246.98	736.3	na	67191	2552
17-2	Apr14	1	2	1231	221.39	633.8	na	57606	2552
BIR	Apr14	1	1	1498	3.75	596.3	na	57027	1
S2-3	Apr14	1	1	1463	52.19	714.1	na	60795	636
S4-13	Apr14	1	1	1323	105.64	768.0	na	70129	922
S5-14	Apr14	1	1	1303	130.74	685.1	na	66343	1284
17-2	Apr14	2	3	1675	275.65	836.1	na	78538	2552
17-2	Apr14	2	4	1689	267.25	919.3	na	78906	2552
BIR	Apr14	2	2	2002	2.24	755.5	na	75605	1
S2-3	Apr14	2	2	1867	64.56	935.1	na	81887	636
S4-13	Apr14	2	2	1576	129.35	851.0	na	83137	922
S5-14	Apr14	2	2	1668	155.44	926.6	na	88019	1284
DID	414	1		2000	1.55		64500	75200	1
17.0	Aug 14	1	1	2090	4.55	na	54950	75298	2552
17-2	Aug14	1	1	1008	229.25	na	54852	155/8	2552
84-13	Aug14	1	1	1690	111.83	na	63089	85445	922
\$5-14	Aug14	1	1	1666	132.61	na	62368	85969	1284
NIST610	Jan15	1	1	8.93	2.40	na	19856	65293	0
17-2	Jan15	1	1	1730	260.84	na	63317	83707	2552
BIR	Jan15	1	1	2024	5.86	na	74146	83281	1
BIR	Jan15	1	2	1937	7.24	na	69111	78085	1
S4-13	Jan15	1	1	1550	110.68	na	64005	83959	922
S5-14	Jan15	1	1	741	72.45	na	27781	36857	1284
S5-14	Jan15	1	2	1549	136.26	na	64240	83601	1284
Araldite	Jan15	1	1	0.28	18366.29	na	3.1	2.5	na
S4-13	Jan15	3	2	1733	119.65	na	77079	99568	922
S5-14	Jan15	3	3	1630	142.16	na	68705	90646	1284
17-2	Jan15	3	2	1830	268.53	na	65545	87564	2552
BIR	Jan15	3	2	2079	4.21	na	73911	84208	1
ST1	Jan15	3	1	1462	49.30	na	61442	81878	691
ST3	Jan15	3	1	1635	162.58	na	74877	94895	1394
BIR	Jun15	1	1	2115	6.83	na	63775	79457	1
BIR	Jun15	1	2	1831	4 4 3	na	53684	67036	1
17-2	Jun15	1	1	1265	223.46	na	38897	58151	2552
ST1	Jun15	1	1	1173	46.91	na	37578	54822	691
ST3	Jun15	1	1	11/7	133.58	na	33727	48808	130/
\$4-13	Jun15	1	1	1010	87.80	na	30302	56601	022
\$5.14	Jun15	1	1	1351	135 31	na	40004	71073	1284
Araldite	Jun15	1	1	0.28	23030.24	na	85	206	1204
RIP	Jun15	2	3	1745	23730.24	na	50/08	63566	110
17.2	Jun 15	2	2	1/40	2.77	114	20621	50740	1551
17-2 64 12	Jun 15	2	2	1421	119.60	na	47504	60052	2352
S5 14	Jun15	2	2	1339	122.79	na	4/304	60045	1284
\$2.2	Jun 15	2	1	1309	51.60	na	40320	51004	1204
52-5 ST1	Jun 15	2	1	1411	J1.09	na	27260	54560	na
ST1 6T2	Jun 15	2	2	11//	42.70	na	20664	45215	1204
S15 0T4	Jun15	2	2	1143	131.43	na	50004	43313	1394
514	Jun 15	2	1	1524	12.31	na	52591	15/54	8/1
St2	Jun15	2	1	1339	97.54	na	45449	64661	1136

Table ppm.	C.8.	Vola	utile	and	trace	eler	nent	stand	lards 1	for K	istufí	ell a	nd N	liðfe	II SIV	MS s	essio	ns. I	0V	anal	lysis	unu	ber.]	H ₂ O	in w	t%, 8	all of	her e	leme	ints i	ц
Standard	Aonth	Day	No.	H ₂ O	ц	ц	Si	³⁵ CI	³⁷ CI	х	Ħ	Sr	7	Zr	qN	Ba	La	c	Pr	PN	Sm	Eu ¹⁵⁶ ,	Gd ¹⁵⁷	'Gd	Tb	Dy I	Чo	Er T	,E	ъ I	.3
ST1 /	Apr14	3	-	3.04	na	114.5	233720	21.2	na	1458	4363	327	15.3	41.3	1.20 8	34.5 3	3.27 8.	.78 1	.26 7	.81 1.	.95 0.	.52 0.	.60 2	2.20 0	.33 2	.55 0.	59 1.	48 0.2	24 1.8	39 0.2	56
ST3 /	Apr14	ŝ	-	1.37	na	108.7	233720	25.3	na	1420	4355	330	15.1	41.5	1.00 8	34.0 3	3.55 8.	.74 1	.45 6	.97 1.	.81 0.	.53 0.	.88 2	2.24 0	.45 2	.40 0.	47 1.	45 0.2	22 1.7	0.0	25
BCR-2G	Apr14	Э	-	na	na	430.3	253820	34.3	na	11327	11586	285	30.7 1	72.5 1	1.53 51	1.8 21	1.73 44	.54 5	.48 25	.23 6.	.00	.64 6.	.72 6	5.59 0	.93 5	.52 1.	25 3.	33 0.4	44 2.8	37 0.5	53
GSD-1G	Apr14	ŝ	-	-0.01	na	29.7	233720	78.2	na	18864	6146	57	37.1	39.6 3	8.27 5	52.1 33	3.89 33.	.24 36	.78 39	.45 39.	.90 34.	.51 41.	25 42	2.03 39	.37 42	.35 42.	73 34.	49 42.5	55 44.9	8 45.9	96
ST1	Apr14	4	0	2.69	na	123.3	233720	26.0	na	1599	4500	347	15.6	42.7	1.18 5	92.1	3.58 9.	.12	.45 8	.93 2	.17 0	.78 1.	35 2	2.92 C	40	.0	62 1.	71 0.2	20 1.6	20 22	27
ST3 /	Apr14	4	0	1.33	na	122.3	233720	25.4	na	1571	4477	35	15.5	42.9	1.21 5	6.06	3.52 9	1	.46 8	.24 2	.27 0	.67 1	.48	2.25 C	.43	.29 0.	.60 1.	61 0.2	20 1.8	.0 .0	20
BCR-2G	Apr14	4	0	0.03	na	536.8	253350	42.5	na	11465	11715	283	31.2 1	73.3 1	1.86 51	17.3 22	2.40 43.	.97 5	.35 26	.36 5.	.68 1	.22 6.	24 6	5.40 C	.94 5	.23 1.	22 3.	72 0.1	51 3.1	14	49
GSD-1G	Apr14	4	0	0.00	na	31.0	233720	87.6	na	19960	6266	58	36.6	38.7 3	9.41 5	52.8 35	5.12 34	.85 37	.61 39	.92 39.	.34 35	.73 39.	.74 40	0.12 41	.23 46	.06 44.	21 34.	70 44.	22 47.3	33 48.	13
ST1 /	Apr14	5	З	2.93	na	134.5	233720	20.0	na	1518	4378	330	15.1	43.6	1.10 8	35.2 3	3.61 8.	.96 1	.26 8	.78 2.	.08	.69 1.	.11 2	2.07 0	.33 2	.33 0.	52 1.	75 0.2	21 2.0	.0 .7	21
ST3 /	Apr14	5	ю	1.43	na	122.0	233720	24.7	na	1445	4354	327	15.0	41.4	1.31 8	33.4 3	3.25 8.	.59 1	.27 7	.64 2	.08 0.	.53 0.	.25 2	2.47 0	44.	.01 0.	38 1.	35 0.2	22 1.8	33 0.2	21
BCR-2G	Apr14	5	б	0.02	na	476.2	253350	35.7	na	11290	11371	278	29.7 1	67.7 1	1.56 45	4.5 2 0	0.75 43.	.28 5	.51 23	.78 5.	.39 1	.59 6.	.11 6	5.19 6	.97 6	.15 1.	.14 3.	29 0.4	45 2.9		4
GSD-1G	Apr14	2	б	0.00	na	37.9	233720	94.8	na	19001	6147	57	36.4	38.9 3	8.33 5	50.1 34	4.11 34.	.11 37	.59 39	.39 39.	.61 34	.36 37.	57 38	3.28 38	.96	.81 42.	44 33.	35 42.0	57 46.0)6 46.0	80
KL-2G /	Apr14	5	-	0.00	na	169.7	233720	21.9	na	2999	13008	297	23.1 1	42.2 1	2.88 8	87.1 1)	1.73 27	.60 3	.82 19	20 4	.87 1	.60 3	38 5	5.25 C	.69	.52 0.	95 2.	38 0.	30 2.(8 0.	31
ST1 /	Aug14	2	1	2.95	na	103.1	233720	78.4	na	1758	4524	366	15.7	44.7	1.21 9	9.2 3	1.88 10.	20 1.	50 8.	.86 2.	.11 0.	.67 2.	69 2	76 0	.47 3	.03 0.	.62 2.	19 0.2	28 1.7	75 0.2	25
ST3	Aug14	6	-	1.42	na	103.0	233720	78.4	na	1735	4472	364	15.9	43.9	1.24 9	8 .3 3	1.85 9.	53 1.	41 7.	.52 2.	.19 0.	.74 1.	48 2	27 0	.55 3	.13 0.	66 1.	82 0.2	22 1.5	56 0.2	25
GSD-1G	Aug14	6	-	0.03	na	32.9	248680	295.5	na	24078	6664	67	40.9	43.4 4	3.79 6	3.3 38	1.71 37.	52 41	43 43	.94 44.	.98 40.	.96 52.	36 53	3.12 45	.96 52	.12 50.	33 37.	91 50.0	06 52.4	17 55.7	72
VIST610 4	Aug14	6	-	-0.01	na	522	327670	1622	na	490	438	514	448	438	487	430	463 4	49 4	133 4	147 4	157 4	182 4	138 4	438 3	354	38 3	60 3	32 33	35 34	-61 	49
BCR-2G	Aug14	6	-	0.04	na	412.4	253820	114.5	na	13583	11880	309	32.3 1	78.0 1	2.48 59	04.8 23	1.35 47.	.15 5.	81 26	.27 6.	31 2.	.17 6.	85 6	6.87 1	.07 6	.53 1.	34 3.	95 0.5	56 3.3	37 0.5	55
GSD-1G	Aug14	3	6	0.02	na	27.8	248680	274.8	na	25267	6775	69	40.4	42.5 4	6.39 6	5.3 39	0.78 40.	.11 42	.85 43.	.37 46.	.95 41.	.60 55.	69 54	1.38 48	.05 52	59 51.	70 40.	18 51.8	89 55.3	30 58.4	43
VIST610 4	Aug14	3	0	0.00	na	425	327670	1533	na	562	447	565	470	460	520	503	510 4	4 4	171 4	176 5	503 5	537 4	181	505 4	7 7	175 5	11 4	80	75 50	86	88
ST1 /	Aug14	4	0	2.88	na	94.8	233720	73.5	na	1835	4557	372	16.0	45.0	1.02 10	3.8 3	3.96 10.	.33 1.	.46 8	.72 2.	.49 0.	.86 1.	33 2	2.77 0	.47 3	.13 0.	59 1.	66 0.2	25 1.8	34 0.2	25
ST3 /	Aug14	4	7	1.45	na	114.9	233720	96.8	na	1778	4506	362	15.2	45.1	1.41 5	9.9 3	3.86 9.	26 1	.48 7	.92 2.	.36 0.	.77 2.	.13 2	2.13 0	.49 2	.91 0.	52 1.	60 0.	22 1.6	51 0.2	26
GSD-1G	Aug14	4	3	0.01	na	35.4	248680	291.7	na	26136	6868	69	41.1	44.2 4	5.05 6	57.5 38	3.93 38.	.81 42	.35 44	.68 47.	.32 42.	.41 57.	.49 59	P.36 47	.77 51	.02 50.	78 40.	75 50.9	95 55.5	58 57.8	88
NIST610 /	Aug14	4	б	0.00	na	410	327670	1640	na	554	446	565	471	462	517	499	501 4	485 4	167 4	479 4	194 £	531 4	173	500	181	174 4	99 4	65 4.	59 49	24	73
S4-13 J	an15	2	-	3.54	7.60 1	1111.2	233720	2701.5	2750.0	14224	4435	657	23.3 1.	39.8 1	9.50 77	1.4 44	1.39 85.	53 9.	.40 39	.35 7.	.28 0.	.99 5.	40 7	7.21 0	.86 4	.89 1.	07 2.	40 0.	38 2.5	55 0.3	36
S5-14 J	an15	6	-	2.75	8.64	818.4	233720	1805.4	1709.9	14182	4420	644	21.9 1	36.0 1	9.45 74	16.8 41	.80 80	8 60.	.99 39	.56 7.	.50 1.	.04 5.	.44	5.19 0	.88	.01 0.	91 2.	49 0.3	32 1.6	58 O.	32
17-Feb J	an15	0	-	3.63	7.08	36.9	233720	82.0	84.1	13699	4304	617	22.3 1	30.4 1	9.53 72	36.1 43	3.40 79.	20 8	.89 37	51 7.	.06 0.	.9 6.	.78 6	5.20 0	.80 4	.01 0.	.98 2.	43 0.3	37 2.2	26 O.	37
NIST610 J	an15	0	-	0.00	344	172	233720	679	2323	368	316	387	334	328	365	329	344 3	339 3	321 2	335 3	341 3	359 3	326	340	327	322 3	32 3	12 3	10 32	25 32	54
GSD-1G	an15	0	-	0.01	34.20	32.6	233720	327.5	429.1	19926	6288	55	36.0	37.3 3	8.57 5	50.0 31	1.11 30	.63 35	.09 34	.78 36.	00 33	.58 41	.60 45	5.01 37	.06	.00 39.	92 30.	82 40.	57 43.2	21 43.9	16
u gl-USD	cl ne	2 1	2 1	10.0	33.11	C-87	233720	316.9	3030	20392	1650	80	30.9	5/.5 5	: 00°	15 017	<u>55</u> 55.1	.45 36 26	12 20 12 21	. 25 28 20	د <i>د</i> ود. ۲۰	55. 14 14	47 47	7.0 VC	-24 	-01 41.	28 31. 20 31.	92 42.0	5.14 C0	204 50 204 50	67 92
L OLUTCINI	an15	1 4	1 0	3.58	100 737	000 2	07/007	2335.8	20503	11950	4006	100	1 8 66	1 276	12 00 8	83 47	- 67 10 1	- 020 34 0	36 40	, 5 40	, 00 1	35 7		0 223	- 71 C	5 - 0	1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1	56	10 01		5 8
S5-14 J	an15	4	10	2.98	8.47	780.8	233720	1661.4	1544.2	13299	4349	627	22.7	35.4 1	8.79 72	8.7 42	2.84 80.	585	28 ± 28 ±	28 6	- 1 - 2	25 4		5.15 0	6.6	26 0.	79 2	23	34 1.8	8	1 00
ST1 J	an15	4	-	3.00	4.48	122.2	233720	81.9	48.8	1572	4475	339	15.3	43.0	1.16 8	37.7 3	.44	96	33 7.	.45 1.	.65 0.	.67 2.	51 2	21 0	37 2	50 0.	64 1.	52 0.2	26 1.4	9	27
ST3 J	an15	4	-	1.32	4.42	107.2	233720	84.8	56.0	1571	4422	340	14.9	43.3	1.26 8	38.8	1.51 8.	74 1	43 7	.45 2.	.05 0.	.83 3.	.16 2	2.45 0	.45	.58 0.	52 1.	58 0.2	24 1.4	15 0.2	25
f 019LSIN	an15	4	ю	0.00	346	175	233720	1051	2375	347	310	368	321	314	346	303	327 3	320 3	307 3	321 3	325 3	341 3	307	321 3	316	305 3	21 3	04 3(30 31	8	10
GSD-1G J	an15	4	б	0.01	33.78	30.3	233720	300.8	404.0	19363	6344	58	36.5	37.9 3	8.41 5	51.5 34	1.01 32	.98 36	.70 37	.81 39.	.07 34.	.31 43.	21 44	1.80 39	.38 41	.17 42.	26 32.	80 42.2	22 46.3	9 47.4	41
S4-13 J	an15	5	ŝ	3.54	7.52	117.1	233720	2748.5	2789.6	14443	4510	657	23.4 1	39.2 2	0.29 77	73.5 44	1.05 82.	.82 9	.62 40	.33 7.	.62 1.	.00	.62 5	5.71 0	.86 4	.97 0.	97 2.	60 0.	38 2.2	24 0.	35
ST1 J	an15	5	0	2.88	4.43	120.7	233720	103.1	91.0	1609	4521	349	15.3	45.1	1.16 8	39.9 3	3.62 8.	.66 1	.49 8	.15 2.	.20 0	.72 2.	22 2	2.30 G	.41	.61 0.	57 1.	41 0.2	23 1.6	57 0.2	25
ST3 J	an15	5	7	1.41	4.42	119.1	233720	113.2	76.5	1611	4519	339	14.8	42.9	1.26 8	88.1 3	3.64 9.	47 1	.48 7	31 2	.13 0	.66 0.	.80	2.02 G	.36 1	.91 0.	57 1.	90	23 1.7	72 0.2	26
NIST610	an15	ŝ	4	0.00	349	117	233720	862	2166	373	330	389	338	334	377	329	343	339	322	343 3	345	360	313	328	331	319 3	36 3	17 3	16 33	20 20	2
GSD-1G	an15	ŝ	4	0.01	34.07	36.8	233720	321.3	473.8	20497	6419	57	37.2	39.0 4	1.26	53.1 35	3.42 33	.81 36	36 39	59 39	.07 35	.27 42	59 45	5.06 35	. 81	.73 43.	01 33.	75 42.0	53 42.5	56 47.	21
T-1G	an15	5	-	0.11	14.34	177.3	233720	190.1	179.8	10680	3066	193	17.8 1	15.3	6.78 24	13.3 45	.50 83	.70 8	.24 28	.71 4	0 06 ⁻	.74 6	E 69.	J. 98 (. 67 3	.54 0.	62 2.	02 0.	32 1.2	0.02	53
GSD-1G	un15	6	-	-0.21	33.20	30.5	234000	42.9	56.5	19500	6110	55	35.5	37.0 3	7.70 4	19.6 32	2.80 32.	40 35.	30 37	20 35.	60 32	.00 38.	80 38	37	80 39	50 39.	80 31.	10 40.2	20 43.1	10 44.0	20

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Standard	Month	Day	No.	H_2O	Ľ	Ч	Si	³⁵ CI	³⁷ CI	K	Ξ	Sr	Y	Zr	ЧN	Ba	La	ő	Pr	ΡN	Sm	Eu	156Gd 1	57Gd	Tb	Dy	Но	Er	$\mathbf{T}_{\mathbf{m}}$	Yb	Lu
ST1	Jun15	7	-	2.73	4.48	137.0	234000	15.0	12.1	1840	4540	351	15.2	44.7	1.40	97.1	3.61	9.64	1.43	8.38	1.97	0.76	0.87	2.99	0.35	2.59	0.60	1.69	0.25	.19	.30
ST3	Jun15	2	-	1.38	4.41	144.0	234000	15.8	19.1	1800	4460	345	14.8	44.3	1.53	95.9	4.01	10.10	1.53	6.72	1.91	0.71	0.24	2.69	0.47	2.78	0.57	1.63	0.21	.58 (0.26
GSD-1G	Jun15	б	7	0.01	32.20	33.7	234000	48.0	65.6	21000	6190	58	36.5	39.1 4	40.60	52.8	34.40	35.00 2	37.00	39.00	39.80	35.50	49.80	45.60	41.60 4	14.10	14.30 3	4.60 4	5.00 48	5 06.3	.40
STI	Jun15	б	6	3.17	4.34	132.0	234000	14.2	12.2	1620	4390	332	14.9	43.4	1.23	90.3	3.80	10.00	1.25	8.61	2.26	0.79	-0.88	3.88	0.26	2.44	0.69	1.75	0.25	.88	.32
ST3	Jun15	б	0	1.56	4.27	145.0	234000	16.0	17.5	1620	4420	336	15.7	42.8	1.19	90.7	3.79	9.16	1.60	8.24	1.98	0.95	1.18	2.37	0.49	2.96	0.83	1.65	0.27	.67	.30
ST4	Jun15	б	-	1.43	4.23	157.0	234000	10.7	8.6	1760	4540	359	15.8	46.5	1.27	102.0	4.13	10.20	1.50	9.85	2.63	0.75	2.08	2.73	0.46	3.12	0.59	1.56	0.30	.85 (.34
ST2	Jun15	б	-	2.76	4.31	152.0	234000	15.6	10.3	1820	4660	372	16.6	48.2	1.52	104.0	4.53	10.30	1.77	9.33	2.43	0.91	1.73	3.04	0.49	3.07	0.65	1.96	0.28	.92	.29
GSD-1G	Jun15	4	ŝ	0.01	33.50	28.5	234000	38.0	50.3	19500	6010	57	36.9	38.0 2	38.80	50.3	33.50	33.70 5	36.90 4	40.40	40.70	34.90	40.00	42.80	40.70 4	14.10	12.70 3	3.40 4	3.80 45	5.50 48	.90
T-1G	Jun15	4	-	0.00	14.50	165.0	234000	25.7	30.7	11000	3010	195	18.2 1	15.0	6.81	256.0 :	51.90	87.10	8.56	31.70	4.57	1.08	2.76	3.61	0.56	3.95	0.75	2.16	0.33	.94	.27
NIST610	Jun15	4	ŝ	0.00	346	357	234000	171	335	347	310	368	321	314	346	303	327	320	307	321	325	341	290	298	316	303	321	301	299	317	307
STI	Jun15	4	ŝ	2.90	4.35	122.0	234000	11.8	6.0	1560	4320	333	15.4	42.7	1.14	87.5	3.58	8.73	1.42	8.52	2.18	0.66	1.43	2.68	0.52	2.09	0.66	1.42	0.26	.55 (.25
ST3	Jun15	4	ŝ	1.41	4.27	118.0	234000	14.9	20.9	1570	4290	336	15.2	43.2	1.13	89.6	3.65	9.43	1.45	9.01	2.18	0.86	1.41	2.20	0.31	2.10	0.59	1.41	0.23	.98	.24
GSD-1G	Jun15	5	4	0.01	33.60	33.9	234000	332.0	407.0	19700	6220	58	37.5	39.0 4	40.00	51.5	34.40	34.70 §	38.10	10.70	39.20	35.50	42.60	43.00	41.40 4	15.70 4	13.60 3	4.50 4	4.00 4	80 48	3.70
T-1G	Jun15	5	0	0.00	14.20	173.0	234000	193.0	189.0	11100	3110	198	18.4 1	18.0	7.20	254.0	52.10	90.30	8.63	31.80	4.39	1.19	5.15	4.58	0.57	3.18	0.69	2.34	0.27	.65 (0.28
ST1	Jun15	5	4	2.87	4.46	125.0	234000	117.0	77.3	1560	4500	340	15.1	45.2	1.28	90.5	4.01	9.31	1.62	7.99	2.10	0.73	2.18	2.56	0.44	2.56	0.55	1.65	0.22	.86	.25
ST3	Jun15	5	4	1.32	4.21	121.0	234000	104.0	98.7	1520	4430	334	15.2	42.6	1.23	85.2	3.58	9.47	1.44	7.25	2.39	0.84	1.92	2.35	0.40	2.81	0.57	1.68	0.21	. 16	0.30
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Table C.9 Major element standards for basaltic glass and olivine analyses - Miðfell. All in wt%.

tandard	Phase	SiO_2	TiO_2	Al_2O_3	Cr_2O_3	FeO	MgO	CaO	MnO	NiO	$\mathrm{Na_2O}$	$\rm K_2O$	P_2O_5	Total
113716	ß	51.49	1.35	14.97	0.08	9.15	8.02	11.06	0.16	na	2.63	0.07	0.14	99.34
8113716	G	51.35	1.36	15.00	0.02	9.10	8.01	11.11	0.21	na	2.62	0.09	0.14	99.26
113716	Ū	51.28	1.34	14.91	0.05	9.30	7.99	11.18	0.20	na	2.55	0.07	0.13	99.24
113716	Ū	50.78	1.31	14.99	0.08	9.19	7.93	11.01	0.23	na	2.76	0.06	0.15	98.72
113716	G	51.30	1.40	15.02	0.01	9.22	7.90	11.06	0.21	na	2.73	0.09	0.13	99.29
111240	G	50.35	1.93	13.54	0.03	11.67	6.74	10.78	0.29	na	2.65	0.22	0.23	98.78
111240	Ū	50.34	1.89	13.58	0.01	11.70	6.88	10.78	0.24	na	2.59	0.20	0.21	98.78
8111240	G	49.96	1.85	13.69	0.01	11.24	6.96	10.72	0.17	na	2.73	0.21	0.22	98.11
111240	Ū	50.50	1.87	13.62	0.00	11.32	6.92	10.66	0.23	na	2.71	0.20	0.20	98.59
111240	ß	50.58	1.90	13.63	0.02	11.38	6.79	10.82	0.26	na	2.64	0.16	0.20	98.72
anC-OI	Ю	41.08	0.00	0.04	0.02	9.76	48.62	0.11	0.24	0.39	na	na	na	100.25
anC-Ol	Ō	41.12	0.01	0.03	0.02	9.86	48.74	0.10	0.20	0.38	na	na	na	100.45
anC-Ol	ō	40.88	0.01	0.03	0.05	9.70	48.95	0.11	0.17	0.37	na	na	na	100.28
anC-Ol	Ō	40.79	0.00	0.04	0.02	9.79	48.45	0.25	0.15	0.35	na	na	na	99.85
anC-Ol	Ō	41.13	0.01	0.04	0.01	9.80	48.66	0.11	0.14	0.35	na	na	na	100.25
anC-Ol	ō	41.15	0.01	0.03	0.01	9.73	48.72	0.11	0.13	0.38	na	na	na	100.27
anC-Ol	Ō	41.20	0.00	0.04	0.05	9.63	48.77	0.11	0.12	0.40	na	na	na	100.32
anC-Ol	ō	41.05	0.01	0.03	0.06	LL_{0}	48.94	0.11	0.20	0.39	na	ц	na	100 55

Table C.10Volatile and trace element data from CO2 reconstructed melt inclusions measured by X-ray tomography, Ramam spectroscopy andSIMS at WHOI. Glass samples are from mount 9 (Table C.5). MI - melt inclusion, GI - glass, CO2 - melt inclusion glass CO2 content, CO2rec vapour bubble and melt inclusion reconstructed CO₂ content.

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Sample	Phase	CO_2	CO ₂ rec	H_2O	н	s	G	Li	К	μ	Sr	Y	Zr	ЧN	Ba	La	Ce	Pr	S PN	m	Eu (L PC	Tb I	y F	I oF	Er T	m J	b I	3
Ol-W1-MI-1	IM	75	na	600	33	880	0.76	2.23	19	4843	52.4	18.32	27.24	0.16	0.88	0.32	1.75 0	.46	3.69 1.	86 0	39 2.	42 0.4	44 3.2	35 0.1	71 2.5	57 0.3	31 2.0	10.2	8
Ol-W4-MI-1	IW	330	na	640	72	066	117	2.51	1048	10025	132.8	16.43	81.55	14.36 6	59.53 1	0.91 2.	2.10 2	.81 15	3.25 3.	000	.75 3.	65 0.¢	54 3.	53 0.7	71 2.1	18 0.3	30 2.3	5 0.2	50
OI-W5-MI-1	IW	70	na	660	49	730	0.8	2.03	18	4906	48.5	16.83	23.04	0.14	0.65	0.30	1.26 0	4	3.13 1.	61 0	39 2.	28 0.5	50 3.:	58 0.	77 1.5	0.0	30 2.	0.0	37
OI-W5-MI-2	IW	99	na	069	57	640	0.7	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na i	na	na	na	na	na 1	1a	na
Ol-W5-MI-3	IW	99	na	620	46	700	0.71	2.28	25	5407	64.9	16.81	22.02	0.08	1.76	0.56	1.39 0	4.	4.71 1.	56 0	.37 2.	21 0.4	47 3.1	28 0.	71 1.5	38 0.2	25 2.(0.0	25
OI-W5-MI-4	IM	67	na	660	52	700	0.73	2.05	16	4850	52.7	14.41	20.74	-0.08	0.78	0.16	1.41 0	.32	1.96 1.	58 0	.22 1.	36 0.2	26 2.(D6 0.:	57 0.9	.0 76	20 1.	1.0.1	81
OI-W5-MI-5	IW	64	na	630	47	700	0.71	2.00	18	5068	52.7	18.19	22.30	0.12	0.63	0.33	1.09 0	36 2	2.82 1.	70 0	.36 2.	80 0.4	47 2.5	90 0.t	69 I.(58 0.2	22 1.9	2 0.2	4
Ol-W8-MI-1	IW	830	na	720	56	1480	7.3	2.43	93	5729	105.3	18.44	23.44	0.94	5.97	1.03	3.53 0	171 4	4.61 1.	44 0	.38 3.	07 0.4	49 3.:	56 0.	70 2.5	.0 66	38 2.3	0.0	22
OI-W8-MI-2	IM	950	na	650	57	710	7.9	2.43	110	6230	111.4	19.11	25.90	1.10	6.95	1.24	3.85 0	: 66	3.83 1.	89 0	47 2	95 0.4	48 3.	36 0.	77 2.2	34 0.3	35 2.2	4 0.3	36
OI-W12-MI-1	IW	1160	2540	640	47	950	95	2.27	866	5876	196.6	17.22	77.49	12.06 4	14.89	9.39 2	0.63 2	31 8	3.56 2.	03 0	.31 2.	60 0.4	45 2.5	97 0.4	67 2.1	0.0	25 1.8	0.0	20
Ol-W14-MI-2	IW	1020	2160	670	59	1200	103	2.27	956	6843	245.8	16.49	76.22	12.20 6	52.23	9.67 1	9.73 2	42 1	1.28 2.	70 0	.61 3.	35 0.5	57 2.5	90 0	76 2.1	10 0.	39 1.9	0.3	38
OI-W17-MI-1	IW	400	na	630	4	1120	109	2.05	1016	7384	150.2	15.16	78.55	13.78 7	70.98 1-	0.48 2.	2.66 2	72 1	2.89 3.	65 0	.64 2.	84 0.5	59 3.:	56 1.(00 2.1	11 0.	34 2.4	5 0.5	53
OI-W22-MI-1	IW	1030	2930	570	60	800	21	2.51	206	6188	102.2	18.08	45.69	2.47 1	12.39	2.93	7.87 1	.42	5.92 1.	92 0	.41 2.	45 0.4	44 3.:	56 0.7	75 2.(0.0	27 2.	6 0.3	57
OI-W23-MI-2	IW	1080	3900	580	63	1640	51	2.52	477	6706	108.6	17.27	42.09	7.29 2	24.81	3.86	8.48 1	.32	7.04 3.	29 0	.49 2.	41 0.5	39 3.2	54 0.0	63 1.t	56 0.3	30 1.5	0.0	31
OI-W27-MI-1	IW	1050	2800	630	73	880	31	2.48	275	5152	122.6	16.74	46.64	4.42 1	19.31	3.21	8.49 1	.33 (5.07 1.	86 0	.38 2.	22 0.4	40 2.8	80 0.t	69 2.8	87 0.3	31 1.7	3 0.2	53
OI-W27-MI-2	IW	890	2070	630	68	790	36	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na i	na	na	na 1	na	na 1	la I	na
Ol-W30-MI-1	IW	830	1880	620	62	1230	95	2.30	882	7006	204.0	18.19	69.99	11.49 5	55.52	8.67 1	5.14 1	3 26.	8.24 2.	44	.47 1.	97 0.4	45 2.4	52 O.t	63 1.5	80 0.2	26 1.5	33 0.2	25
OI-W38-MI-1	IW	890	1400	600	72	960	57	2.51	555	7450	149.8	17.89 .	58.01	7.31 3	37.15	5.05 1	1.43 1	5 6L.	 3.13 3. 	10 0	.59 2.	78 0.5	50 3.4	52 0.7	72 2.1	12 0.3	39 2.	6 0.4	1 3
OI-W38-MI-2	IW	096	2520	630	75	950	61	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na 1	na	na	na i	na	na 1	Ia I	na
Ol-W38-MI-3	IW	720	1340	620	73	1060	53	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na 1	na	na	na 1	na	na 1	1	na
Ol-W40-MI-1	IW	1130	4550	740	60	1600	58	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na 1	na	na	na 1	na	na 1	a I	na
Ol-W41-MI-1	IW	820	1730	560	37	660	22	2.02	208	4325	67.5	14.90	23.05	2.93 1	12.96	2.02	4.34 0	.63	3.41 1.	59 0	.36 2.	38 0.4	46 3.	37 0.:	58 2.(0.7	24 1.0	8 0.3	8
OI-W42-MI-1	III	930	2330	570	68	830	55	2.38	609	5518	108.8	14.19	42.86	5.74 3	37.61	3.77	8.96 1	.30	7.62 2.	08 0	.42 1.	94 0.2	29 2.(54 0.	70 1.1	.0 67	20 1.8	5 0.3	38
mt9ol03gl	ū	na	na	na	na	na	na	2.29	200	5740	117	13.8	19.2	0.91	7.85	0.93	3.16 0	.68	3.78 1.	36 0.	32 2.	03 0.3	37 2.5	50 0.0	61 1.6	55 0.2	24 1.7	6 0.2	12
mt9ol14g1	G	na	na	na	na	na	na	2.35	170	5980	122	14.0	18.3	0.76	8.59	0.93	3.22 0	.55 2	3.07 1.	54 0	.32 1.	88 0.5	38 2.t	55 0.:	56 1.7	74 0.	17 1.3	88 0.2	53
mt9ol20gl	G	na	na	na	na	na	na	2.46	223	6300	125	14.7	19.2	1.15	7.88	0.99	3.13 0	09.	3.34 1.	86 0	.36 2.	30 0.4	43 2.(54 0.:	52 1.(56 0.3	30 1.5	32 0.3	33
mt9ol22gl	ß	na	na	na	na	na	na	2.30	166	6110	124	14.7	18.9	1.11	8.31	0.91	3.42 0	.66	3.80 1.	50 0	.33 2.	26 0.4	46 2.:	54 0.4	61 1.(59 0.2	27 1.5	3 0.1	4
mt9ol23gl	G	na	na	na	na	na	na	2.50	167	6490	132	15.4	19.3	1.16	7.17	1.03	3.14 0	.63	3.96 1.	68 0	.31 2.	37 0.5	38 2.	75 0.:	51 1.1	0.77	22 1.6	0.0	4

Table C.11 Major element data (wt%) from experimental glass (Gl). No. - analysis number.

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MgO	CaO	MnO	NiO	Na ₂ O	K ₂ O	P_2O_5	SO ₂	Total
mid02	07/10/14	0.001	1310	Gl	1	48.31	0.82	15.33	0.17	9.58	11.43	13.82	0.14	0.05	1.03	0.027	0.031	0.006	100.74
mid02	07/10/14	0.001	1310	Gl	2	47.34	0.85	15.06	0.13	9.49	11.18	13.66	0.20	0.00	0.96	0.026	0.039	0.000	98.93
mid02	07/10/14	0.001	1310	Gl	3	47.54	0.82	15.11	0.09	9.83	11.16	13.66	0.20	0.04	1.07	0.016	0.024	0.000	99.56
mid02	07/10/14	0.001	1310	Gl	4	48.14	0.83	15.14	0.10	9.69	11.12	13.65	0.17	0.00	1.05	0.026	0.011	0.000	99.94
mid02	07/10/14	0.001	1310	GI	5	48.23	0.81	15.19	0.15	9.57	11.08	14.01	0.16	0.01	1.01	0.047	0.021	0.000	100.23
mid02	09/10/14	0.001	1290	Gl	1	47.62	0.79	15.30	0.13	9.64	10.91	13.81	0.21	0.00	1.16	0.013	0.035	0.000	99.64
mid03	09/10/14	0.001	1290	Gl	2	48.23	0.82	15.37	0.14	9.82	10.99	13.88	0.17	0.10	1.09	0.006	0.024	0.004	100.64
mid03	09/10/14	0.001	1290	Gl	3	48.19	0.81	15.23	0.13	10.09	11.22	13.90	0.19	0.04	1.04	0.031	0.021	0.005	100.91
mid03	09/10/14	0.001	1290	Gl	4	48.25	0.83	15.29	0.15	9.60	11.27	13.82	0.21	0.02	1.18	0.029	0.045	0.000	100.68
mid03	09/10/14	0.001	1290	GI	5	48.25	0.83	15.19	0.16	9.91	11.28	13.74	0.17	0.04	1.06	0.038	0.010	0.000	100.66
mid05	10/10/14	0.001	1290	Gl	1	47.98	0.81	15.11	0.15	9.74	11.32	13.63	0.10	0.03	1.08	0.022	0.045	0.000	100.05
mid05	10/10/14	0.001	1270	Gl	2	48.52	0.80	15.10	0.12	9.71	11.02	13.85	0.16	0.04	1.15	0.038	0.040	0.000	100.56
mid05	10/10/14	0.001	1270	Gl	3	48.54	0.82	15.26	0.13	9.42	11.27	13.67	0.13	0.09	1.10	0.013	0.013	0.000	100.45
mid05	10/10/14	0.001	1270	Gl	4	48.34	0.79	15.12	0.17	9.84	11.09	13.96	0.22	0.01	1.17	0.041	0.033	0.011	100.81
mid05	10/10/14	0.001	1270	GI	5	48.32	0.82	15.31	0.11	9.67	11.24	13.95	0.19	0.03	1.20	0.032	0.037	0.014	100.93
mid05	10/10/14	0.001	1250	Gl	1	47.53	0.81	14.96	0.18	9.53	10.81	13.96	0.17	0.04	1.37	0.044	0.017	0.005	99.40
mid06	10/10/14	0.001	1250	Gl	2	48.07	0.83	15.30	0.10	9.44	10.95	13.80	0.18	0.07	1.39	0.032	0.052	0.000	100.20
mid06	10/10/14	0.001	1250	Gl	3	47.73	0.82	15.23	0.11	9.90	10.91	13.94	0.19	0.05	1.36	0.014	0.037	0.002	100.29
mid06	10/10/14	0.001	1250	Gl	4	47.50	0.82	15.11	0.11	9.72	10.99	13.82	0.14	0.03	1.37	0.042	0.030	0.005	99.69
mid06	10/10/14	0.001	1250	GI	5	48.20	0.81	15.23	0.10	9.40	11.14	13.70	0.16	0.05	1.35	0.029	0.024	0.000	100.32
mid16	27/10/14	0.001	1230	Gl	2	47.73	0.82	15.09	0.13	9.22	10.78	13.73	0.19	0.05	1.35	0.038	0.009	0.000	99.04
mid16	27/10/14	0.001	1245	Gl	3	47.45	0.83	15.08	0.11	9.14	10.57	14.01	0.18	0.00	1.32	0.029	0.007	0.000	98.73
mid16	27/10/14	0.001	1245	Gl	4	47.60	0.82	15.12	0.10	9.49	10.55	13.49	0.15	0.05	1.23	0.033	0.048	0.000	98.67
mid16	27/10/14	0.001	1245	Gl	5	47.79	0.81	15.07	0.06	9.48	10.43	13.65	0.17	0.05	1.32	0.037	0.021	0.000	98.90
mid16	27/10/14	0.001	1245	GI	6	47.26	0.84	15.15	0.13	9.51	10.64	13.88	0.16	0.02	1.31	0.034	0.033	0.001	98.96
mid16	27/10/14	0.001	1245	Gl	10	47.40	0.85	15.10	0.11	9.95	10.49	13.81	0.19	0.00	1.24	0.039	0.045	0.000	99.15
mid16	27/10/14	0.001	1245	Gl	12	47.65	0.84	15.27	0.10	9.67	10.61	13.76	0.16	0.00	1.39	0.024	0.016	0.000	99.50
mid16	27/10/14	0.001	1245	Gl	17	47.34	0.82	15.20	0.14	9.51	10.62	13.57	0.15	0.03	1.30	0.038	0.032	0.007	98.77
mid16	27/10/14	0.001	1245	Gl	18	47.37	0.82	15.20	0.11	9.44	10.61	13.60	0.18	0.00	1.32	0.006	0.016	0.012	98.69
mid16	27/10/14	0.001	1245	GI	19 20	47.83	0.83	15.17	0.09	9.46	10.65	13.61	0.17	0.07	1.29	0.018	0.034	0.000	99.20
mid10	21/10/14	0.001	1240	Gl	1	47.44	0.85	15.32	0.10	9.41	10.39	13.85	0.23	0.05	1.31	0.052	0.022	0.001	99.05
mid11	21/10/14	0.001	1240	Gl	2	48.17	0.83	15.27	0.07	9.20	10.25	13.81	0.24	0.04	1.39	0.033	0.032	0.005	99.34
mid11	21/10/14	0.001	1240	Gl	3	47.86	0.83	15.58	0.09	9.33	10.07	13.91	0.17	0.07	1.40	0.062	0.038	0.000	99.41
mid11	21/10/14	0.001	1240	Gl	5	48.08	0.82	15.40	0.13	9.60	10.27	13.94	0.17	0.00	1.34	0.074	0.023	0.004	99.85
mid11	21/10/14	0.001	1240	GI	7	47.87	0.82	15.42	0.10	9.08	10.30	13.81	0.17	0.03	1.30	0.030	0.020	0.001	98.88 99.07
mid11	21/10/14	0.001	1240	Gl	8	47.58	0.84	15.49	0.09	9.36	10.24	13.99	0.20	0.05	1.32	0.049	0.034	0.002	99.23
mid11	21/10/14	0.001	1240	Gl	9	48.13	0.83	15.40	0.09	9.30	10.18	13.78	0.23	0.01	1.40	0.034	0.021	0.006	99.41
mid11	21/10/14	0.001	1240	Gl	10	48.02	0.84	15.29	0.11	9.54	9.97	14.19	0.19	0.04	1.36	0.041	0.040	0.014	99.65
mid11	21/10/14	0.001	1240	GI	11	48.27	0.85	15.35	0.15	9.15	10.10	13.77	0.17	0.00	1.35	0.055	0.012	0.014	99.24
mid11	21/10/14	0.001	1240	Gl	14	47.51	0.83	15.35	0.12	9.19	10.25	13.84	0.17	0.03	1.33	0.062	0.025	0.000	98.72
mid11	21/10/14	0.001	1240	Gl	16	47.59	0.84	15.39	0.09	9.76	10.40	13.90	0.19	0.00	1.31	0.038	0.028	0.004	99.54
mid11	21/10/14	0.001	1240	Gl	17	47.48	0.84	15.41	0.11	9.25	10.25	13.96	0.16	0.07	1.27	0.034	0.021	0.000	98.86
mid11	21/10/14	0.001	1240	Gl	18	47.57	0.86	15.31	0.08	9.60	10.18	13.97	0.17	0.02	1.37	0.032	0.036	0.014	99.21
mid11	21/10/14	0.001	1240	GI	19 20	47.71	0.83	15.28	0.12	9.10	10.21	13.92	0.26	0.04	1.30	0.037	0.010	0.009	98.83
mid07	11/10/14	0.001	1230	Gl	1	47.90	0.86	15.43	0.13	9.54	10.03	14.09	0.11	0.07	1.33	0.024	0.023	0.000	99.52
mid07	11/10/14	0.001	1230	Gl	2	48.07	0.83	15.58	0.11	9.25	10.17	14.24	0.20	0.01	1.26	0.022	0.047	0.000	99.82
mid07	11/10/14	0.001	1230	Gl	3	48.51	0.85	15.48	0.12	9.53	9.93	14.01	0.21	0.00	1.32	0.025	0.031	0.008	100.02
mid07	11/10/14	0.001	1230	Gl	4	47.98	0.86	15.52	0.11	9.24	10.05	13.87	0.21	0.00	1.31	0.015	0.037	0.000	99.21
mid07	11/10/14	0.001	1230	GI	5 6	48.19 48.37	0.84	15.51	0.15	9.01 9.74	10.06 9.81	14.11	0.15	0.01	1.30	0.019	0.035	0.000	100.04
mid07	11/10/14	0.001	1230	Gl	7	48.43	0.83	15.31	0.11	9.60	9.98	14.20	0.18	0.03	1.24	0.021	0.027	0.000	99.96
mid07	11/10/14	0.001	1230	Gl	8	47.79	0.86	15.49	0.15	9.44	9.94	14.05	0.17	0.02	1.33	0.026	0.011	0.000	99.27
mid07	11/10/14	0.001	1230	Gl	9	47.63	0.82	15.49	0.11	9.53	10.03	14.05	0.14	0.04	1.27	0.028	0.029	0.002	99.20
mid07	11/10/14	0.001	1230	GI	10	47.91	0.87	15.42	0.12	9.40 0.47	10.05	14.19	0.18	0.00	1.36	0.021	0.023	0.000	99.54
mid07	11/10/14	0.001	1230	Gl	12	48.01	0.83	15.65	0.11	9.47 9.45	9.71	13.93	0.19	0.05	1.33	0.025	0.012	0.007	99.55
mid07	11/10/14	0.001	1230	Gl	13	47.93	0.85	15.66	0.10	9.40	9.67	14.20	0.20	0.01	1.33	0.027	0.035	0.000	99.42
mid07	11/10/14	0.001	1230	Gl	14	47.80	0.84	15.57	0.09	9.56	9.98	14.08	0.17	0.06	1.28	0.044	0.022	0.006	99.50
mid07	11/10/14	0.001	1230	Gl	15	47.60	0.84	15.45	0.06	9.52	9.76	14.31	0.18	0.03	1.26	0.024	0.017	0.000	99.07
mid07	11/10/14	0.001	1230	GI	16 17	47.97 47.62	0.84	15.56	0.08	9.27	9.54	14.26	0.18	0.06	1.31	0.020	0.021	0.000	99.11 00 17
mid07	11/10/14	0.001	1230	Gl	18	47.75	0.84	15.61	0.10	9.15	9.83	14.29	0.20	0.01	1.35	0.017	0.021	0.014	99.20

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al_2O_3	Cr ₂ O ₃	FeO	MgO	CaO	MnO	NiO	Na ₂ O	K ₂ O	P_2O_5	SO ₂	Total
mid07	11/10/14	0.001	1230	Gl	19	47.85	0.86	15.60	0.06	9.76	9.98	13.98	0.14	0.02	1.40	0.048	0.027	0.000	99.72
mid07	11/10/14	0.001	1230	Gl	20	48.10	0.86	15.58	0.11	9.46	9.85	14.07	0.23	0.06	1.36	0.017	0.024	0.000	99.72
mid18	30/10/14	0.001	1225	Gl	1	47.93	0.85	15.66	0.08	9.27	9.62	13.96	0.17	0.00	1.41	0.041	0.028	0.000	99.02
mid18	30/10/14	0.001	1225	Gl	2	48.23	0.85	15.70	0.11	9.07	9.61	14.10	0.17	0.02	1.45	0.035	0.034	0.012	99.38
mid18	30/10/14	0.001	1225	GI	3	47.69	0.85	15.48	0.14	9.19	9.58	14.13	0.18	0.01	1.34	0.040	0.031	0.010	98.69
mid18	30/10/14	0.001	1225	GI	4	47.88	0.83	15.55	0.10	9.47	9.66	13.98	0.15	0.02	1.51	0.048	0.027	0.004	99.23
mid18	30/10/14	0.001	1225	GI	7	47.45	0.85	15.54	0.06	9.48	9.70	14.10	0.21	0.05	1.41	0.059	0.007	0.005	98.89
mid18	30/10/14	0.001	1225	GI	8	47.35	0.82	15.55	0.11	9.50	9.50	14 20	0.14	0.03	1 33	0.050	0.031	0.007	98 71
mid18	30/10/14	0.001	1225	Gl	15	47.57	0.82	15.55	0.14	9.14	9.87	13.84	0.17	0.05	1.36	0.074	0.019	0.000	98.59
mid18	30/10/14	0.001	1225	Gl	16	48.02	0.84	15.62	0.07	9.25	9.67	14.03	0.17	0.01	1.35	0.045	0.030	0.014	99.12
mid13	23/10/14	0.001	1220	Gl	1	48.21	0.87	15.27	0.09	9.46	9.51	14.03	0.22	0.00	1.27	0.013	0.038	0.000	98.99
mid13	23/10/14	0.001	1220	Gl	2	47.91	0.89	15.33	0.15	9.32	9.47	14.11	0.22	0.03	1.29	0.011	0.016	0.006	98.75
mid13	23/10/14	0.001	1220	Gl	3	48.09	0.87	15.48	0.09	9.67	9.51	14.20	0.20	0.07	1.30	0.006	0.046	0.001	99.53
mid13	23/10/14	0.001	1220	Gl	4	48.56	0.86	15.33	0.07	9.47	9.52	14.05	0.20	0.03	1.40	0.013	0.032	0.000	99.54
mid13	23/10/14	0.001	1220	GI	5	48.11	0.87	15.43	0.13	9.39	9.37	14.26	0.17	0.00	1.35	0.027	0.016	0.001	99.12
mid12	23/10/14	0.001	1220	GI	7	40.00	0.80	15.56	0.10	9.00	9.55	13.94	0.15	0.02	1.55	0.011	0.014	0.000	99.03
mid13	23/10/14	0.001	1220	GI	8	48.10	0.84	15.42	0.09	9.33	9.49	14.12	0.12	0.01	1.55	0.024	0.028	0.000	99.18
mid13	23/10/14	0.001	1220	Gl	9	48.28	0.87	15.50	0.15	9.15	9.30	14.13	0.16	0.02	1.37	0.013	0.033	0.004	99.00
mid13	23/10/14	0.001	1220	Gl	10	48.16	0.85	15.43	0.12	9.30	9.55	14.06	0.13	0.03	1.29	0.007	0.017	0.001	98.96
mid08	14/10/14	0.001	1210	Gl	1	48.17	0.97	14.69	0.07	9.94	8.93	14.41	0.18	0.01	1.37	0.023	0.011	0.003	98.78
mid08	14/10/14	0.001	1210	Gl	2	48.76	0.93	14.72	0.12	9.71	9.12	14.33	0.18	0.00	1.46	0.011	0.029	0.021	99.39
mid08	14/10/14	0.001	1210	Gl	3	48.73	0.95	14.82	0.08	9.97	9.10	14.45	0.22	0.03	1.55	0.022	0.027	0.014	99.95
mid08	14/10/14	0.001	1210	Gl	4	48.30	0.94	14.76	0.10	9.97	9.11	14.25	0.16	0.00	1.41	0.038	0.036	0.006	99.07
mid08	14/10/14	0.001	1210	GI	5	47.96	0.97	14.87	0.12	9.78	9.14	14.36	0.20	0.00	1.44	0.006	0.035	0.001	98.88
mid08	14/10/14	0.001	1210	GI	0	48.30	0.95	15.09	0.10	9.77	9.17	14.59	0.21	0.00	1.41	0.026	0.054	0.005	99.71
mid08	14/10/14	0.001	1210	GI	8	48.91	0.97	14.02	0.07	9.43	9.21	14 24	0.21	0.02	1.44	0.012	0.026	0.000	99.43
mid08	14/10/14	0.001	1210	Gl	9	48.71	0.92	14.55	0.05	9.82	9.19	14.31	0.24	0.00	1.49	0.023	0.027	0.011	99.34
mid08	14/10/14	0.001	1210	Gl	10	48.74	0.94	14.80	0.09	9.42	9.03	14.24	0.18	0.00	1.45	0.004	0.023	0.000	98.92
mid24	05/11/14	0.001	1205	Gl	1	48.73	0.95	14.32	0.13	10.23	8.99	14.63	0.20	0.05	1.42	0.016	0.027	0.013	99.72
mid24	05/11/14	0.001	1205	Gl	2	48.40	0.98	14.17	0.12	10.11	9.01	14.39	0.15	0.06	1.39	0.029	0.031	0.000	98.85
mid24	05/11/14	0.001	1205	Gl	3	48.70	0.97	14.21	0.10	10.06	9.08	14.38	0.19	0.04	1.37	0.028	0.032	0.016	99.17
mid24	05/11/14	0.001	1205	Gl	5	48.39	1.00	14.42	0.13	9.76	8.97	14.48	0.13	0.03	1.41	0.002	0.034	0.010	98.76
mid24	05/11/14	0.001	1205	GI	6	49.01	0.95	14.39	0.09	9.73	9.05	14.50	0.17	0.06	1.48	0.013	0.028	0.000	99.47
mid24	05/11/14	0.001	1205	GI	8	48.09	0.90	14.27	0.11	10.21	9.01	14.52	0.24	0.00	1.40	0.007	0.034	0.020	99.27
mid24	05/11/14	0.001	1205	GI	9	48.82	0.99	14.20	0.10	9.83	9.02	14 35	0.24	0.00	1.40	0.044	0.025	0.000	99.24
mid24	05/11/14	0.001	1205	Gl	10	48.31	0.96	14.33	0.08	10.12	9.14	14.27	0.20	0.01	1.43	0.021	0.025	0.011	98.90
mid24	05/11/14	0.001	1205	Gl	11	49.00	1.00	14.25	0.11	9.88	8.97	14.24	0.17	0.00	1.38	0.027	0.039	0.011	99.07
mid24	05/11/14	0.001	1205	Gl	12	48.99	0.98	14.20	0.10	10.26	9.08	14.55	0.18	0.03	1.37	0.024	0.032	0.002	99.80
mid24	05/11/14	0.001	1205	Gl	13	49.00	0.96	14.34	0.10	10.37	9.00	14.39	0.22	0.00	1.46	0.041	0.033	0.007	99.91
mid24	05/11/14	0.001	1205	Gl	14	48.54	0.97	14.26	0.06	9.97	9.04	14.47	0.18	0.02	1.47	0.053	0.024	0.006	99.08
mid24	05/11/14	0.001	1205	Gl	15	48.95	0.97	14.26	0.10	10.19	8.78	14.32	0.22	0.00	1.43	0.044	0.020	0.014	99.31
mid24 mid20	05/11/14	0.001	1205	GI	16	48.54	1.07	14.37	0.09	0.72	8.94	14.52	0.25	0.02	1.38	0.028	0.025	0.005	99.47
mid20	01/11/14	0.001	1200	GI	2	48.78	1.07	13.89	0.14	10.35	8.69	14.39	0.15	0.00	1.40	0.021	0.055	0.000	98.00
mid20	01/11/14	0.001	1200	GI	3	49.47	1.06	14.02	0.09	9.93	8.75	14.44	0.20	0.03	1.48	0.042	0.023	0.000	99.53
mid20	01/11/14	0.001	1200	Gl	4	49.03	1.06	13.99	0.07	10.43	8.81	14.38	0.23	0.03	1.44	0.031	0.030	0.006	99.53
mid20	01/11/14	0.001	1200	Gl	7	48.78	1.07	13.97	0.10	10.15	8.61	14.40	0.20	0.01	1.44	0.030	0.026	0.009	98.78
mid20	01/11/14	0.001	1200	Gl	8	49.51	1.03	14.02	0.11	10.41	8.73	14.46	0.23	0.01	1.42	0.044	0.033	0.006	100.01
mid20	01/11/14	0.001	1200	Gl	9	49.43	1.04	14.06	0.12	10.35	8.66	14.53	0.26	0.02	1.51	0.019	0.039	0.012	100.05
mid20	01/11/14	0.001	1200	Gl	10	49.43	1.05	14.12	0.09	10.20	8.94	14.57	0.22	0.05	1.49	0.041	0.044	0.006	100.25
mid20	01/11/14	0.001	1200	Gl	11	49.03	1.04	14.13	0.09	10.10	8.71	14.40	0.30	0.02	1.41	0.038	0.045	0.013	99.32
mid20	01/11/14	0.001	1200	GI	12	49.23	1.05	14.08	0.10	9.73	8.09	14.20	0.18	0.02	1.46	0.025	0.055	0.000	99.00
mid20	01/11/14	0.001	1200	GI	14	49.05	0.98	14.04	0.00	9.98	8 84	14 51	0.15	0.02	1.43	0.006	0.037	0.000	99.16
mid20	01/11/14	0.001	1200	Gl	15	49.47	1.03	14.07	0.08	10.22	8.79	14.41	0.24	0.00	1.46	0.002	0.037	0.000	99.81
mid20	01/11/14	0.001	1200	Gl	16	48.97	1.02	13.92	0.10	10.39	8.75	14.57	0.22	0.00	1.49	0.043	0.061	0.000	99.54
mid20	01/11/14	0.001	1200	Gl	17	49.08	1.03	13.89	0.13	10.33	8.91	14.63	0.21	0.04	1.52	0.008	0.024	0.011	99.81
mid20	01/11/14	0.001	1200	Gl	18	49.05	1.03	13.97	0.14	9.95	8.73	14.48	0.21	0.02	1.49	0.023	0.032	0.010	99.14
mid20	01/11/14	0.001	1200	Gl	19	48.86	1.03	14.06	0.12	10.24	8.80	14.39	0.20	0.03	1.47	0.030	0.028	0.000	99.26
mid20	01/11/14	0.001	1200	Gl	20	48.42	1.05	14.04	0.06	10.39	8.96	14.40	0.22	0.01	1.43	0.002	0.038	0.000	99.02
mid09	15/10/14	0.001	1190	GI	1	48.79	1.10	14.63	0.09	10.83	/.86	14.44	0.22	0.04	1.65	0.044	0.049	0.000	99.74
mid00	15/10/14	0.001	1100	GI	э л	40.09	1.11	12.90	0.10	10.00	0.2J 8 14	14.12	0.10	0.05	1.30	0.037	0.030	0.000	70.03 00.07
mid09	15/10/14	0.001	1190	Gl	5	49.04	1.12	13.87	0.12	10.73	8.44	14.21	0.16	0.04	1.75	0.014	0.036	0.000	99.38
mid09	15/10/14	0.001	1190	Gl	6	48.90	1.12	13.90	0.04	10.92	8.06	14.27	0.26	0.00	1.58	0.046	0.062	0.000	99.16
mid09	15/10/14	0.001	1190	Gl	7	49.06	1.13	13.91	0.08	10.95	8.15	14.27	0.24	0.02	1.60	0.054	0.058	0.002	99.51
mid09	15/10/14	0.001	1190	Gl	8	48.98	1.12	14.00	0.06	11.05	8.26	14.05	0.21	0.00	1.89	0.062	0.039	0.000	99.71

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MgO	CaO	MnO	NiO	Na ₂ O	K ₂ O	P_2O_5	SO ₂	Total
mid09	15/10/14	0.001	1190	Gl	9	49.06	1.05	12.86	0.19	10.39	9.42	14.91	0.24	0.03	1.57	0.086	0.026	0.000	99.83
mid09	15/10/14	0.001	1190	Gl	10	49.25	1.19	13.92	0.08	11.00	8.26	14.32	0.24	0.00	1.63	0.028	0.041	0.004	99.96
mid09	15/10/14	0.001	1190	Gl	11	49.58	1.06	12.89	0.17	10.44	9.02	14.92	0.23	0.01	1.47	0.016	0.035	0.013	99.85
mid09	15/10/14	0.001	1190	Gl	12	48.87	1.12	13.94	0.11	11.25	8.26	14.16	0.22	0.04	1.66	0.067	0.049	0.001	99.76
mid12	22/10/14	0.001	1170	Gl	1	48.82	1.31	13.52	0.05	11.68	7.89	13.57	0.25	0.06	1.56	0.049	0.061	0.010	98.83
mid12 mid12	22/10/14	0.001	1170	GI	2	49.14	1.27	13.66	0.06	11.68	7.84	13.03	0.28	0.04	1.50	0.038	0.038	0.019	99.26
mid12	22/10/14	0.001	1170	Gl	4	49.15	1.18	14.11	0.07	11.19	7.79	14.12	0.23	0.01	1.00	0.040	0.059	0.003	99.65
mid12	22/10/14	0.001	1170	Gl	5	48.78	1.20	14.35	0.07	11.51	7.70	13.80	0.20	0.02	1.63	0.034	0.043	0.008	99.35
mid31	01/12/14	5	1250	Gl	1	48.20	0.96	16.35	0.15	6.30	8.93	10.08	0.11	0.03	7.31	0.042	0.028	0.063	98.54
mid31	01/12/14	5	1250	GI	2	48.11	0.95	16.59	0.14	6.43	0.00 8 51	10.12	0.10	0.00	7.41	0.031	0.034	0.040	98.23
mid31	01/12/14	5	1250	Gl	4	47.96	0.96	16.56	0.06	6.23	8.74	10.34	0.06	0.00	7.26	0.056	0.037	0.048	98.31
mid31	01/12/14	5	1250	Gl	5	48.13	0.93	16.62	0.10	6.28	8.72	10.41	0.09	0.00	7.37	0.031	0.009	0.056	98.75
mid31	01/12/14	5	1250	Gl	6	48.22	0.94	16.59	0.15	6.17	8.78	10.36	0.08	0.01	7.26	0.021	0.033	0.049	98.66
mid31	01/12/14	5	1250	Gl	7	48.10	0.92	16.70	0.11	6.30	8.77	10.61	0.07	0.00	7.16	0.051	0.023	0.070	98.88
mid31	01/12/14	5	1250	Gl	8	48.15	0.94	16.59	0.12	6.15	8.80	10.31	0.07	0.03	7.45	0.039	0.038	0.054	98.74
mid31	01/12/14	5	1250	GI	10	48.38	0.91	16.61	0.11	6.06	8.72	10.14	0.10	0.00	7.38	0.031	0.019	0.048	98.51
mid31	01/12/14	5	1250	Gl	11	48.62	0.94	16.58	0.10	6.30	8.69	10.03	0.09	0.04	7.29	0.047	0.014	0.059	99.13
mid31	01/12/14	5	1250	Gl	12	47.85	0.93	16.71	0.10	6.53	8.63	10.28	0.06	0.00	7.46	0.043	0.034	0.053	98.68
mid31	01/12/14	5	1250	Gl	13	48.31	0.92	16.70	0.16	6.35	8.77	10.41	0.07	0.02	7.45	0.029	0.029	0.061	99.27
mid31	01/12/14	5	1250	Gl	14	48.85	0.95	16.79	0.11	5.99	8.82	10.38	0.05	0.00	7.40	0.033	0.033	0.046	99.45
mid32	03/12/14	5	1230	Gl	1	50.25	1.08	18.12	0.08	4.35	8.06	9.04	0.05	0.00	7.73	0.061	0.041	0.044	98.91
mid32	03/12/14	5	1230	Gl	2	49.64	1.07	18.05	0.12	4.31	8.06	9.00	0.06	0.02	7.79	0.062	0.039	0.044	98.27
mid32	03/12/14	5	1230	GI	3	50.05	1.09	18.08	0.11	4.47	8.21	9.17	0.07	0.00	7.70	0.065	0.019	0.038	99.13
mid32	03/12/14	5	1230	Gl	5	50.11	1.07	18.01	0.07	4.45	8.01	9.10	0.08	0.00	7.66	0.054	0.032	0.040	98.68
mid32	03/12/14	5	1230	Gl	6	49.97	1.06	18.00	0.10	4.70	8.18	9.42	0.03	0.02	7.77	0.050	0.022	0.044	99.35
mid32	03/12/14	5	1230	Gl	7	50.08	1.07	17.93	0.11	4.44	8.17	9.20	0.00	0.00	7.70	0.056	0.047	0.044	98.84
mid32	03/12/14	5	1230	Gl	8	49.28	1.05	17.76	0.07	4.61	7.93	9.36	0.07	0.03	7.58	0.055	0.017	0.046	97.84
mid32	03/12/14	5	1230	Gl	9	49.96	1.03	17.79	0.06	4.77	8.12	9.45	0.06	0.00	7.64	0.027	0.042	0.047	98.99
mid32	03/12/14	5	1230	GI	10	49.91	1.04	18.06	0.07	4.66	8.25	9.30	0.05	0.05	7.65	0.044	0.040	0.039	99.18
mid35	05/12/14	5	1210	GI	2	48.41	0.90	17.20	0.07	5.88 6.02	7.60	11.93	0.11	0.00	6.62	0.100	0.044	0.057	99.00 99.48
mid35	05/12/14	5	1210	Gl	3	48.32	0.97	17.13	0.09	6.34	7.60	11.75	0.10	0.00	6.43	0.103	0.004	0.053	98.89
mid35	05/12/14	5	1210	Gl	4	48.34	0.96	17.26	0.10	5.76	7.66	11.74	0.09	0.00	6.60	0.103	0.045	0.060	98.71
mid35	05/12/14	5	1210	Gl	5	49.31	0.96	16.98	0.09	6.11	7.45	11.65	0.07	0.00	6.58	0.101	0.021	0.046	99.37
mid35	05/12/14	5	1210	Gl	6	49.47	1.01	17.12	0.09	5.79	7.57	11.41	0.07	0.00	6.54	0.099	0.022	0.063	99.25
mid35	05/12/14	5	1210	Gl	7	48.92	0.96	17.07	0.10	6.14	7.45	12.09	0.12	0.00	6.62	0.111	0.016	0.052	99.66
mid35	05/12/14	5	1210	GI	9	49.17	0.97	16.79	0.13	6.68	7.52	12.02	0.14	0.04	6.04	0.089	0.024	0.050	99.41
mid35	05/12/14	5	1210	Gl	10	47.73	0.94	17.11	0.13	6.58	7.49	12.87	0.14	0.00	5.93	0.074	0.050	0.052	99.09
mid35	05/12/14	5	1210	Gl	11	47.45	0.94	18.57	0.08	6.01	6.66	12.40	0.04	0.00	6.47	0.116	0.057	0.042	98.84
mid35	05/12/14	5	1210	Gl	12	47.59	0.91	18.49	0.06	5.91	6.75	12.32	0.17	0.04	6.24	0.137	0.040	0.051	98.71
mid35	05/12/14	5	1210	Gl	13	47.89	0.96	18.46	0.06	5.82	6.74	12.22	0.11	0.01	6.77	0.091	0.061	0.050	99.24
mid35	05/12/14	5	1210	Gl	14	48.30	0.94	18.48	0.09	5.57	6.80	11.96	0.12	0.00	6.48	0.116	0.062	0.038	98.95
mid34	05/12/14	5	1210	GI	15	48.21	0.96	18.35	0.05	5.48 10.36	6.45	12.09	0.09	0.00	0.89	0.152	0.045	0.045	99.10
mid34	05/12/14	5	1190	Gl	2	46.43	0.82	16.01	0.06	10.30	6.74	13.09	0.20	0.04	4.73	0.158	0.022	0.074	98.54
mid34	05/12/14	5	1190	Gl	4	46.45	0.78	16.15	0.07	10.46	6.40	12.84	0.21	0.05	4.89	0.143	0.019	0.073	98.53
mid34	05/12/14	5	1190	Gl	5	46.47	0.82	15.90	0.08	10.34	6.47	13.13	0.17	0.03	4.73	0.154	0.004	0.082	98.37
mid34	05/12/14	5	1190	Gl	6	46.31	0.81	16.09	0.09	10.24	6.53	13.33	0.16	0.00	4.67	0.139	0.027	0.069	98.47
mid34	05/12/14	5	1190	Gl	10	46.52	0.80	15.98	0.12	10.37	6.62	13.16	0.21	0.01	4.70	0.166	0.011	0.068	98.73
m1d34	05/12/14	5	1190	GI	11	46.32	0.77	16.14	0.08	10.09	6.68	13.42	0.17	0.02	4.68	0.146	0.025	0.073	98.61
mid21	02/11/14	10	1412	Gl	1	47.05	0.78	14.96	0.06	9.38	10.96	14.19	0.19	0.00	1.51	0.051	0.025	0.060	99.21
mid21	02/11/14	10	1412	Gl	1	47.26	0.78	14.94	0.13	9.48	10.70	14.24	0.20	0.00	1.58	0.017	0.033	0.070	99.44
mid21	02/11/14	10	1412	Gl	1	46.62	0.80	14.70	0.12	9.45	10.86	14.37	0.14	0.00	1.51	0.038	0.019	0.080	98.70
mid21	02/11/14	10	1412	Gl	2	46.94	0.82	14.95	0.13	9.58	10.88	14.07	0.15	0.00	1.48	0.005	0.035	0.084	99.11
mid21	02/11/14	10	1412	GI	2	47.59	0.85	14.84 14.74	0.15	9.40 9.56	10.05	14.59	0.10	0.00	1.54	0.01/	0.026	0.068	99.48 90 20
mid21	02/11/14	10	1412	Gl	3	47.12	0.79	15.11	0.13	9.82	10.93	13.98	0.21	0.02	1.40	0.023	0.029	0.070	99.67
mid21	02/11/14	10	1412	Gl	3	47.03	0.77	15.07	0.09	9.49	10.61	14.09	0.16	0.03	1.51	0.030	0.022	0.072	98.98
mid21	02/11/14	10	1412	Gl	4	47.55	0.81	15.03	0.10	9.61	10.70	14.44	0.15	0.00	1.48	0.037	0.027	0.076	100.01
mid21	02/11/14	10	1412	Gl	4	47.31	0.81	15.06	0.16	9.48	10.69	13.96	0.20	0.01	1.54	0.020	0.031	0.075	99.34
mid21	02/11/14	10	1412	Gl	4	46.46	0.81	14.52	0.15	9.68	11.07	14.32	0.19	0.00	1.52	0.033	0.018	0.083	98.85
mid21	02/11/14	10	1412	Gl	5	47.35	0.83	14.98	0.13	9.54	10.85	14.22	0.09	0.03	1.51	0.008	0.014	0.071	99.63
mid21 mid21	02/11/14	10	1412	GI	5	40.59	0.80	14.71	0.13	9.49	10.92	14.33	0.16	0.06	1.45	0.035	0.027	0.075	98.77 100.49

Table C.11 (continued)

Experiment	Date	Р	Т	Phase	No.	SiO_2	TiO ₂	$Al_2O_3\\$	Cr_2O_3	FeO	MgO	CaO	MnO	NiO	Na ₂ O	K ₂ O	P_2O_5	SO_2	Total
mid21	02/11/14	10	1412	Gl	6	46.57	0.82	14.68	0.12	9.72	10.68	14.24	0.20	0.00	1.52	0.032	0.033	0.069	98.68
mid21	02/11/14	10	1412	Gl	8	47.05	0.81	14.70	0.13	9.13	10.99	14.11	0.20	0.00	1.49	0.038	0.028	0.071	98.74
mid21	02/11/14	10	1412	Gl	9	46.61	0.79	14.77	0.15	9.45	10.90	14.15	0.18	0.03	1.48	0.007	0.034	0.069	98.62
mid21	02/11/14	10	1412	Gl	10	47.20	0.80	14.82	0.11	9.45	10.75	14.08	0.20	0.01	1.52	0.016	0.020	0.077	99.03
mid21	02/11/14	10	1412	GI	13	47.17	0.79	14.89	0.10	9.54	10.07	14.00	0.20	0.00	1.46	0.000	0.024	0.008	98.94
mid21	19/11/14	10	1320	Gl	1	48.17	0.81	15.06	0.11	9.65	11.03	13.69	0.12	0.01	1.40	0.015	0.042	0.094	100.32
mid28	19/11/14	10	1320	Gl	2	47.31	0.81	15.14	0.15	9.46	11.11	13.42	0.18	0.01	1.51	0.023	0.017	0.070	99.21
mid28	19/11/14	10	1320	Gl	3	47.99	0.81	15.20	0.11	9.69	11.36	13.29	0.16	0.00	1.46	0.037	0.026	0.075	100.20
mid28	19/11/14	10	1320	Gl	4	47.74	0.78	15.11	0.15	9.84	10.89	13.46	0.15	0.06	1.54	0.028	0.020	0.066	99.86
mid28	19/11/14	10	1320	Gl	5	48.05	0.81	15.12	0.15	9.70	11.29	13.51	0.20	0.01	1.52	0.023	0.031	0.071	100.47
mid28	19/11/14	10	1320	Gl	6	47.56	0.79	14.92	0.09	9.66	11.06	13.50	0.16	0.00	1.53	0.045	0.027	0.064	99.41
mid28	19/11/14	10	1320	GI	8	47.70	0.80	15.11	0.15	9.49	11.19	13.45	0.14	0.06	1.48	0.022	0.027	0.085	99.68
mid28	19/11/14	10	1320	Gl	9	47.93	0.81	15.11	0.13	9.59	11.06	13.25	0.21	0.00	1.46	0.016	0.022	0.078	99.68
mid28	19/11/14	10	1320	Gl	10	48.26	0.78	15.00	0.14	9.45	10.99	13.36	0.22	0.00	1.46	0.015	0.024	0.091	99.78
mid25	08/11/14	10	1310	Gl	1	47.36	0.90	17.10	0.02	10.81	8.99	12.57	0.14	0.00	1.84	0.037	0.009	0.110	99.89
mid25	08/11/14	10	1310	Gl	2	47.22	0.93	17.21	0.05	10.80	8.83	12.69	0.17	0.00	1.84	0.026	0.021	0.095	99.88
mid25	08/11/14	10	1310	Gl	3	47.04	0.92	16.86	0.04	10.97	8.70	12.63	0.17	0.02	1.70	0.037	0.037	0.123	99.25
mid25	08/11/14	10	1310	Gl	4	47.85	0.91	17.31	0.00	10.90	9.00	12.70	0.14	0.01	1.88	0.028	0.031	0.106	100.86
mid25	08/11/14	10	1310	GI	5	47.30	0.91	16.93	0.02	10.61	8.71	12.39	0.16	0.00	1.85	0.057	0.025	0.088	99.02
mid25	08/11/14	10	1310	Gl	7	47.58	0.91	17.21	0.03	10.81	8.70	12.40	0.14	0.01	1.96	0.023	0.032	0.087	100.14
mid25	08/11/14	10	1310	Gl	8	47.43	0.92	17.17	0.03	10.57	8.77	12.61	0.16	0.00	1.83	0.044	0.034	0.088	99.65
mid25	08/11/14	10	1310	Gl	9	47.33	0.94	17.03	0.04	10.63	8.86	12.71	0.20	0.01	1.90	0.044	0.021	0.090	99.80
mid25	08/11/14	10	1310	Gl	10	47.47	0.90	16.77	0.06	10.47	8.74	12.55	0.16	0.04	1.85	0.013	0.030	0.094	99.15
mid10	21/10/14	10	1300	Gl	1	47.08	0.84	15.09	0.15	9.39	10.76	14.29	0.17	0.00	1.41	0.037	0.056	0.087	99.36
mid10	21/10/14	10	1300	Gl	2	47.14	0.82	15.11	0.14	9.68	10.61	14.34	0.21	0.02	1.45	0.038	0.010	0.084	99.65
mid10	21/10/14	10	1300	GI	3	47.07	0.83	15.12	0.12	9.51	10.81	14.25	0.22	0.02	1.3/	0.040	0.007	0.095	99.46
mid10	21/10/14	10	1300	GI	5	40.09	0.81	15.10	0.11	9.40	10.70	14.19	0.20	0.05	1.39	0.031	0.000	0.078	99.75
mid10	21/10/14	10	1300	Gl	6	46.55	0.81	15.02	0.13	9.36	10.90	14.20	0.18	0.00	1.47	0.041	0.007	0.089	98.76
mid10	21/10/14	10	1300	Gl	7	47.18	0.82	15.05	0.12	9.57	10.61	14.12	0.23	0.00	1.47	0.051	0.007	0.081	99.30
mid10	21/10/14	10	1300	Gl	8	46.93	0.81	14.93	0.12	9.47	10.62	14.23	0.12	0.00	1.51	0.046	0.004	0.078	98.87
mid10	21/10/14	10	1300	Gl	9	47.09	0.80	14.96	0.09	9.45	10.90	14.23	0.18	0.03	1.46	0.040	0.019	0.063	99.32
mid10	21/10/14	10	1300	Gl	11	47.39	0.80	15.06	0.13	9.59	10.78	14.22	0.16	0.00	1.41	0.052	0.032	0.095	99.72
mid10	21/10/14	10	1300	GI	12	47.29	0.80	15.15	0.11	9.50	10.74	14.39	0.18	0.02	1.50	0.064	0.030	0.084	99.82
mid10	21/10/14	10	1300	GI	14	47.03	0.79	15.24	0.13	9.41	10.45	14.23	0.23	0.01	1.45	0.053	0.037	0.091	99.37
mid10	21/10/14	10	1300	Gl	16	47.58	0.80	15.19	0.12	9.43	10.65	14.23	0.18	0.00	1.51	0.038	0.016	0.096	99.84
mid10	21/10/14	10	1300	Gl	17	47.52	0.81	15.10	0.09	9.19	10.18	14.21	0.13	0.00	1.43	0.059	0.017	0.089	98.83
mid10	21/10/14	10	1300	Gl	18	47.54	0.82	15.34	0.12	9.25	10.53	14.18	0.16	0.03	1.39	0.052	0.020	0.084	99.52
mid10	21/10/14	10	1300	Gl	19	47.97	0.80	15.16	0.13	9.25	10.35	14.17	0.19	0.00	1.48	0.044	0.032	0.082	99.65
mid10	21/10/14	10	1300	Gl	20	47.35	0.78	15.20	0.15	9.23	10.49	14.25	0.14	0.00	1.42	0.031	0.018	0.072	99.12
mid22	03/11/14	10	1290	GI	2	40.19	0.84	15.92	0.03	9.72	10.27	13.03	0.19	0.04	1.55	0.029	0.040	0.091	98.77
mid22	03/11/14	10	1290	Gl	3	46.59	0.79	15.80	0.05	9.67	10.27	13.63	0.10	0.02	1.50	0.032	0.032	0.074	98.85
mid22	03/11/14	10	1290	Gl	4	46.48	0.82	16.92	0.05	9.44	9.86	13.84	0.21	0.03	1.58	0.036	0.039	0.124	99.45
mid22	03/11/14	10	1290	Gl	5	46.07	0.80	17.07	0.10	9.18	9.77	13.77	0.21	0.00	1.66	0.038	0.019	0.124	98.79
mid22	03/11/14	10	1290	Gl	6	46.96	0.82	15.41	0.09	9.22	10.64	14.15	0.15	0.00	1.43	0.027	0.032	0.081	99.01
mid22	03/11/14	10	1290	Gl	7	47.26	0.78	15.23	0.12	9.14	10.82	14.21	0.15	0.03	1.51	0.033	0.010	0.083	99.39
mid22	03/11/14	10	1290	Gl	8	47.16	0.79	15.31	0.12	9.28	10.73	14.10	0.16	0.00	1.45	0.021	0.008	0.088	99.22
mid22 mid22	03/11/14	10	1290	GI	10	40.83	0.77	15.45	0.09	9.69	10.67	14.10	0.24	0.00	1.44	0.030	0.019	0.098	99.41
mid22	03/11/14	10	1290	Gl	11	46.64	0.78	15.13	0.10	9.35	10.82	14.16	0.22	0.00	1.32	0.035	0.024	0.080	98.66
mid22	03/11/14	10	1290	Gl	12	47.06	0.80	15.14	0.12	9.18	10.87	14.03	0.19	0.01	1.51	0.016	0.027	0.087	99.03
mid22	03/11/14	10	1290	Gl	13	47.44	0.78	15.32	0.16	9.08	10.96	14.24	0.17	0.00	1.43	0.052	0.028	0.082	99.75
mid22	03/11/14	10	1290	Gl	14	47.21	0.79	15.20	0.14	9.56	11.11	14.23	0.18	0.03	1.47	0.036	0.012	0.081	100.05
mid22	03/11/14	10	1290	Gl	15	47.13	0.79	15.22	0.10	9.23	10.93	14.26	0.20	0.00	1.35	0.020	0.017	0.087	99.35
mid14	24/10/14	10	1280	Gl	1	47.10	0.98	16.39	0.06	11.55	8.92	12.07	0.21	0.00	1.74	0.064	0.010	0.086	99.17
mid14	20/11/14	10	1280	GI	1	47.43	0.97	16.36	0.04	11.80	9.32	12.06	0.18	0.00	1.87	0.025	0.020	0.090	100.18
mid29	24/10/14 20/11/14	10	1280	GI	2	47.40 47.56	0.99	16.33	0.06	11.04	0.80 9 NQ	12.13	0.20	0.04	1.83	0.041	0.010	0.1091	99.23 99.73
mid14	24/10/14	10	1280	Gl	3	47.95	0.98	16.70	0.03	11.30	8.84	12.34	0.27	0.00	1.94	0.037	0.036	0.094	100.52
mid29	20/11/14	10	1280	Gl	3	47.64	0.99	16.44	0.06	11.59	8.99	12.03	0.18	0.02	1.89	0.048	0.027	0.106	100.01
mid14	24/10/14	10	1280	Gl	4	47.30	0.98	16.59	0.02	11.52	8.73	12.05	0.24	0.01	1.92	0.053	0.053	0.087	99.56
mid29	20/11/14	10	1280	Gl	4	47.18	0.98	16.44	0.04	11.98	9.30	12.07	0.24	0.04	1.90	0.053	0.022	0.085	100.33
mid14	24/10/14	10	1280	Gl	5	47.14	1.01	16.49	0.03	11.59	8.93	12.16	0.23	0.03	1.90	0.059	0.010	0.089	99.66
mid14	20/11/14	10	1280	GI	5	47.33	0.95	16.34	0.08	11.66	9.14	12.16	0.25	0.00	1.92	0.040	0.032	0.099	100.02
mid29	24/10/14	10	1280	Gl	0 6	47.50	0.97	16.36	0.06	11.28	0.84 9.11	12.01	0.21	0.02	1.94	0.065	0.040	0.104	99.17 100.01
			- 200		0				0.07	0		/ /	/						

Table C.11 (continued)

	D .	D		DI	N	a: 0	TO		0.0	E O		0.0	14.0	NEO	NO	W O	D O	60	75 × 1
Experiment	Date	Р	Т	Phase	NO.	\$10 ₂	1102	AI_2O_3	Cr_2O_3	FeO	MgO	CaO	MnO	NiO	Na ₂ O	K ₂ O	P ₂ O ₅	SO ₂	Total
mid14	24/10/14	10	1280	Gl	7	47.53	0.98	16.40	0.06	11.67	9.00	12.19	0.22	0.00	1.82	0.052	0.047	0.091	100.05
mid29	20/11/14	10	1280	Gl	7	47.27	1.01	16.32	0.08	11.71	9.06	12.06	0.12	0.00	1.79	0.048	0.036	0.109	99.62
mid14	24/10/14	10	1280	Gl	8	47.35	0.98	16.51	0.07	11.47	8.94	12.01	0.14	0.00	1.92	0.042	0.040	0.101	99.57
mid29	20/11/14	10	1280	Gl	8	47.23	1.00	16.37	0.08	11.70	9.10	12.14	0.17	0.01	1.80	0.023	0.018	0.102	99.76
mid14	24/10/14	10	1280	Gl	9	47.38	0.95	16.52	0.08	11.62	8.88	12.11	0.22	0.00	1.94	0.045	0.048	0.090	99.88
mid29	20/11/14	10	1280	Gl	9	47.19	1.00	16.46	0.03	11.66	9.02	11.81	0.19	0.04	1.82	0.061	0.014	0.096	99.38
mid14	24/10/14	10	1280	Gl	10	47.11	0.98	16.45	0.07	11.17	8.95	12.07	0.21	0.03	1.92	0.062	0.018	0.087	99.13
mid29	20/11/14	10	1280	Gl	10	47.41	0.98	16.43	0.04	11.57	9.29	11.88	0.19	0.04	1.90	0.026	0.024	0.105	99.89
mid23	05/11/14	10	1240	Gl	1	47.28	1.35	15.82	0.01	14.22	7.56	11.15	0.19	0.01	1.94	0.024	0.025	0.140	99.71
mid23	05/11/14	10	1240	Gl	2	46.69	1.33	15.85	0.05	14.46	7.72	11.14	0.23	0.04	2.04	0.052	0.014	0.129	99.74
mid23	05/11/14	10	1240	Gl	3	46.84	1.29	15.66	0.03	13.60	9.18	11.61	0.26	0.00	1.72	0.034	0.038	0.129	100.38
mid23	05/11/14	10	1240	Gl	4	46.55	1.39	15.86	0.00	14.17	7.69	11.28	0.26	0.02	2.05	0.054	0.050	0.135	99.51
mid23	05/11/14	10	1240	Gl	5	47.02	1.34	15.98	0.02	14.32	7.68	11.17	0.24	0.02	2.05	0.044	0.047	0.141	100.08
mid23	05/11/14	10	1240	Gl	6	46.60	1.33	15.98	0.02	13.93	7.66	11.37	0.25	0.00	1.95	0.053	0.045	0.133	99.32
mid23	05/11/14	10	1240	Gl	7	46.93	1.33	16.03	0.02	14.11	7.56	11.26	0.23	0.00	2.02	0.056	0.035	0.135	99.73
mid23	05/11/14	10	1240	Gl	8	46.91	1.33	16.00	0.03	14.25	7.74	11.26	0.24	0.02	1.97	0.056	0.055	0.128	100.00
mid23	05/11/14	10	1240	Gl	9	47.21	1.31	15.95	0.04	13.98	7.64	11.23	0.25	0.00	2.01	0.045	0.023	0.129	99.81
mid23	05/11/14	10	1240	Gl	10	46.94	1.32	15.92	0.07	14.19	7.89	11.26	0.26	0.08	2.02	0.056	0.053	0.135	100.19

 Table C.12 Major element data (wt%) from experimental olivine (Ol). No. - analysis number.

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al_2O_3	Cr ₂ O ₃	FeO	MgO	CaO	MnO	NiO	P_2O_5	Total
mid16	27/10/14	0.001	1245	Ol	1	39.01	0.017	0.10	0.08	11.99	46.31	0.44	0.17	0.29	0.001	98.41
mid16	27/10/14	0.001	1245	Ol	2	39.75	0.028	0.39	0.11	12.01	46.57	0.60	0.22	0.26	0.008	99.94
mid16	27/10/14	0.001	1245	Ol	3	39.78	0.017	0.10	0.06	12.26	46.69	0.44	0.25	0.27	0.009	99.88
mid16	27/10/14	0.001	1245	Ol	4	39.40	0.018	0.10	0.13	12.14	46.89	0.46	0.21	0.27	0.024	99.64
mid16	27/10/14	0.001	1245	Ol	5	39.95	0.023	0.10	0.12	12.44	47.17	0.47	0.20	0.27	0.014	100.76
mid16	27/10/14	0.001	1245	Ol	8	39.92	0.019	0.09	0.05	12.30	47.15	0.47	0.18	0.25	0.001	100.45
mid16	27/10/14	0.001	1245	Ol	9	40.06	0.015	0.09	0.11	12.29	47.01	0.47	0.19	0.25	0.008	100.50
mid16	27/10/14	0.001	1245	Ol	10	39.87	0.017	0.10	0.06	12.34	46.90	0.46	0.23	0.25	0.005	100.24
mid16	27/10/14	0.001	1245	Ol	11	39.47	0.027	0.07	0.12	12.11	46.62	0.49	0.17	0.24	0.009	99.33
mid16	27/10/14	0.001	1245	Ol	12	39.19	0.042	0.10	0.06	11.97	46.68	0.48	0.22	0.27	0.022	99.04
mid16	27/10/14	0.001	1245	Ol	14	39.45	0.035	0.10	0.08	12.31	46.50	0.53	0.18	0.24	0.007	99.44
mid16	27/10/14	0.001	1245	Ol	15	39.42	0.026	0.10	0.13	12.34	47.06	0.46	0.19	0.25	0.004	99.98
mid16	27/10/14	0.001	1245	Ol	16	39.79	0.012	0.09	0.07	11.87	47.07	0.43	0.14	0.25	0.021	99.74
mid16	27/10/14	0.001	1245	Ol	17	39.37	0.041	0.37	0.13	12.19	46.05	0.88	0.13	0.23	0.006	99.40
mid16	27/10/14	0.001	1245	Ol	18	39.17	0.028	0.08	0.07	12.07	46.90	0.40	0.23	0.28	0.007	99.24
mid16	27/10/14	0.001	1245	Ol	19	39.42	0.019	0.08	0.08	12.27	46.54	0.50	0.17	0.22	0.005	99.29
mid16	27/10/14	0.001	1245	Ol	20	39.49	0.003	0.10	0.11	12.42	46.83	0.50	0.23	0.26	0.018	99.95
mid16	27/10/14	0.001	1245	Ol	13a	39.50	0.004	0.09	0.11	12.38	46.69	0.46	0.23	0.22	0.023	99.71
mid16	27/10/14	0.001	1245	Ol	13b	39.27	0.013	0.11	0.10	12.07	46.66	0.41	0.17	0.23	0.011	99.04
mid16	27/10/14	0.001	1245	Ol	13c	39.40	0.011	0.09	0.12	12.05	46.67	0.45	0.14	0.24	0.008	99.18
mid16	27/10/14	0.001	1245	Ol	6a	39.94	0.009	0.07	0.09	12.20	47.05	0.44	0.16	0.28	0.005	100.24
mid16	27/10/14	0.001	1245	Ol	6b	39.84	0.010	0.08	0.09	12.18	46.83	0.44	0.24	0.25	0.000	99.94
mid16	27/10/14	0.001	1245	Ol	6c	39.68	0.013	0.07	0.09	12.38	46.75	0.44	0.17	0.26	0.001	99.84
mid16	27/10/14	0.001	1245	Ol	7a	40.07	0.033	0.21	0.11	12.24	46.44	0.67	0.21	0.26	0.010	100.25
mid16	27/10/14	0.001	1245	Ol	7b	40.17	0.023	0.12	0.11	12.18	46.78	0.57	0.23	0.25	0.012	100.43
mid11	21/10/14	0.001	1240	Ol	1	39.64	0.023	0.09	0.07	12.06	46.71	0.43	0.13	0.23	0.005	99.40
mid11	21/10/14	0.001	1240	Ol	2	39.43	0.023	0.09	0.12	12.35	46.55	0.46	0.21	0.26	0.005	99.50
mid11	21/10/14	0.001	1240	Ol	4	39.57	0.015	0.08	0.08	12.37	46.69	0.51	0.23	0.25	0.004	99.81
mid11	21/10/14	0.001	1240	Ol	5	40.00	0.017	0.10	0.08	12.02	46.50	0.43	0.16	0.27	0.017	99.60
mid11	21/10/14	0.001	1240	Ol	6	39.40	0.015	0.08	0.12	12.34	46.68	0.54	0.14	0.25	0.003	99.57
mid11	21/10/14	0.001	1240	Ol	7	39.38	0.010	0.09	0.09	12.04	46.70	0.43	0.17	0.26	0.013	99.18
mid11	21/10/14	0.001	1240	Ol	8	39.70	0.015	0.08	0.12	12.37	46.58	0.48	0.24	0.26	0.009	99.85
mid11	21/10/14	0.001	1240	Ol	9	39.97	0.010	0.08	0.08	12.53	46.39	0.47	0.17	0.23	0.010	99.95
mid11	21/10/14	0.001	1240	Ol	12	39.60	0.005	0.08	0.16	12.55	46.34	0.47	0.18	0.24	0.008	99.64
mid11	21/10/14	0.001	1240	Ol	10a	40.14	0.001	0.08	0.09	12.07	46.89	0.41	0.21	0.28	0.008	100.18
mid11	21/10/14	0.001	1240	Ol	10b	40.00	0.007	0.07	0.08	12.31	46.97	0.43	0.15	0.26	0.009	100.28
mid11	21/10/14	0.001	1240	Ol	11a	40.17	0.003	0.08	0.07	12.13	47.23	0.40	0.24	0.29	0.011	100.62
mid11	21/10/14	0.001	1240	Ol	11b	40.21	0.015	0.09	0.07	12.09	47.08	0.42	0.16	0.28	0.023	100.45
mid11	21/10/14	0.001	1240	Ol	3a	39.37	0.028	0.11	0.09	12.03	47.13	0.45	0.23	0.27	0.015	99.72
mid11	21/10/14	0.001	1240	Ol	3b	39.49	0.020	0.09	0.05	12.03	47.48	0.46	0.21	0.26	0.016	100.10
mid07	11/10/14	0.001	1230	Ol	1	39.36	0.002	0.07	0.12	12.74	45.81	0.53	0.23	0.23	0.004	99.10
mid07	11/10/14	0.001	1230	Ol	2	39.53	0.050	0.37	0.10	12.73	45.47	0.88	0.18	0.24	0.000	99.55
mid07	11/10/14	0.001	1230	Ol	3	39.69	0.019	0.06	0.10	12.76	46.76	0.49	0.22	0.23	0.001	100.33
mid07	11/10/14	0.001	1230	Ol	5	39.53	0.024	0.21	0.18	12.65	46.82	0.48	0.21	0.24	0.009	100.34
mid07	11/10/14	0.001	1230	Ol	6	39.26	0.027	0.17	0.17	12.49	46.67	0.52	0.21	0.20	0.005	99.71

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al_2O_3	Cr_2O_3	FeO	MgO	CaO	MnO	NiO	P_2O_5	Total
mid07	11/10/14	0.001	1230	Ol	7	39.97	0.036	0.13	0.11	12.71	46.97	0.53	0.15	0.21	0.000	100.81
mid07	11/10/14	0.001	1230	Ol	8	39.80	0.013	0.07	0.11	12.54	47.09	0.53	0.16	0.22	0.008	100.54
mid07	11/10/14	0.001	1230	Ol	10	39.53	0.028	0.27	0.17	12.63	46.25	0.58	0.15	0.21	0.005	99.81
mid07	11/10/14	0.001	1230	Ol	11	39.95	0.011	0.08	0.11	12.92	46.31	0.51	0.23	0.26	0.004	100.37
mid07	11/10/14	0.001	1230	Ol	12	39.56	0.012	0.07	0.09	12.84	46.41	0.48	0.23	0.22	0.004	99.90
mid07	11/10/14	0.001	1230	Ol	14	39.95	0.000	0.07	0.16	12.93	46.15	0.45	0.17	0.25	0.000	100.13
mid07	11/10/14	0.001	1230	Ol	15	40.12	0.016	0.08	0.11	12.73	46.17	0.47	0.20	0.26	0.004	100.17
mid07	11/10/14	0.001	1230	OI OI	17	39.79	0.016	0.06	0.09	13.01	46.16	0.48	0.20	0.24	0.000	100.05
mid07	11/10/14	0.001	1230		18	40.09	0.023	0.09	0.11	12.91	40.23	0.48	0.18	0.24	0.012	100.30
mid07	11/10/14	0.001	1230	0	20	39.79	0.021	0.07	0.09	12.85	46.16	0.48	0.21	0.21	0.007	99.92
mid07	11/10/14	0.001	1230	01	13a	39.83	0.014	0.07	0.06	12.95	46.55	0.48	0.22	0.22	0.002	100.40
mid07	11/10/14	0.001	1230	Ol	13b	39.57	0.014	0.09	0.12	13.00	46.32	0.47	0.11	0.23	0.011	99.93
mid07	11/10/14	0.001	1230	Ol	4a	39.58	0.021	0.08	0.10	12.70	46.71	0.46	0.21	0.24	0.000	100.10
mid07	11/10/14	0.001	1230	Ol	4b	40.14	0.025	0.07	0.07	12.56	47.15	0.46	0.20	0.21	0.000	100.88
mid07	11/10/14	0.001	1230	Ol	9a	39.70	0.010	0.09	0.12	12.77	46.43	0.41	0.23	0.27	0.000	100.03
mid07	11/10/14	0.001	1230	Ol	9b	39.86	0.014	0.10	0.09	12.71	46.09	0.44	0.27	0.25	0.000	99.83
mid18	30/10/14	0.001	1225	Ol	1	39.92	0.006	0.08	0.07	12.84	46.52	0.49	0.27	0.23	0.006	100.43
mid18	30/10/14	0.001	1225	OI	2	40.28	0.039	0.57	0.12	13.35	45.92	0.79	0.27	0.23	0.018	101.58
mid18	30/10/14	0.001	1225		3	39.75 20.78	0.022	0.08	0.09	12.06	46.02	0.51	0.25	0.21	0.007	100.05
mid18	30/10/14	0.001	1225		5	39.78	0.020	0.31	0.09	13.09	46.12	0.59	0.10	0.23	0.005	100.24
mid18	30/10/14	0.001	1225	Ol	6	39.55	0.014	0.10	0.11	12.86	46.66	0.50	0.23	0.21	0.005	100.25
mid18	30/10/14	0.001	1225	Ol	7	39.34	0.012	0.08	0.07	12.67	45.77	0.48	0.21	0.22	0.002	98.84
mid18	30/10/14	0.001	1225	Ol	8	40.24	0.009	0.08	0.11	13.06	46.33	0.55	0.24	0.24	0.010	100.88
mid18	30/10/14	0.001	1225	Ol	9	39.55	0.022	0.07	0.13	12.99	46.54	0.53	0.28	0.24	0.009	100.34
mid18	30/10/14	0.001	1225	Ol	10	39.76	0.030	0.11	0.10	12.95	46.25	0.54	0.25	0.25	0.010	100.23
mid18	30/10/14	0.001	1225	Ol	11	39.90	0.024	0.09	0.07	12.94	46.25	0.49	0.22	0.25	0.009	100.22
mid18	30/10/14	0.001	1225	Ol	12	39.42	0.023	0.50	0.33	13.13	46.11	0.48	0.21	0.22	0.004	100.43
mid18	30/10/14	0.001	1225	Ol	13	39.75	0.020	0.16	0.11	12.93	46.43	0.54	0.15	0.24	0.006	100.33
mid18	30/10/14	0.001	1225		14	40.03	0.018	0.10	0.05	13.02	46.47	0.55	0.24	0.24	0.007	100.73
mid18	30/10/14	0.001	1225		15	39.99	0.027	0.20	0.10	13.00	45.40	0.61	0.14	0.22	0.000	100.21
mid18	30/10/14	0.001	1225	01	17	39.36	0.027	0.21	0.21	13.02	46.16	0.59	0.13	0.23	0.011	99.97
mid18	30/10/14	0.001	1225	Ol	18	39.64	0.031	0.10	0.08	12.87	46.07	0.54	0.20	0.24	0.014	99.78
mid18	30/10/14	0.001	1225	Ol	19	39.62	0.010	0.09	0.08	12.95	46.52	0.54	0.22	0.24	0.007	100.27
mid13	23/10/14	0.001	1220	Ol	2	39.80	0.013	0.06	0.07	13.34	46.18	0.50	0.18	0.24	0.006	100.38
mid13	23/10/14	0.001	1220	Ol	3	39.60	0.029	0.09	0.09	13.27	46.03	0.53	0.22	0.19	0.006	100.06
mid13	23/10/14	0.001	1220	Ol	4	39.63	0.010	0.06	0.08	13.34	46.11	0.48	0.19	0.22	0.004	100.12
mid13	23/10/14	0.001	1220	Ol	6	39.34	0.019	0.07	0.09	13.55	45.90	0.52	0.21	0.24	0.006	99.94
mid13	23/10/14	0.001	1220	OI	7	40.17	0.022	0.11	0.10	13.47	46.12	0.48	0.23	0.24	0.008	100.94
mid13	23/10/14	0.001	1220		8	39.40	0.011	0.09	0.07	12.11	45.79	0.45	0.19	0.20	0.000	99.31
mid13	23/10/14	0.001	1220		10	39.86	0.027	0.37	0.18	13.38	44.95	0.51	0.17	0.21	0.002	100.39
mid24	05/11/14	0.001	1205	Ol	1	39.64	0.030	0.06	0.11	14.80	44.93	0.53	0.24	0.19	0.000	100.55
mid24	05/11/14	0.001	1205	Ol	2	39.79	0.012	0.05	0.10	14.51	44.74	0.57	0.22	0.18	0.022	100.21
mid24	05/11/14	0.001	1205	Ol	3	39.88	0.035	0.11	0.05	14.42	44.86	0.56	0.29	0.21	0.002	100.41
mid24	05/11/14	0.001	1205	Ol	4	39.59	0.041	0.08	0.09	14.81	44.72	0.60	0.20	0.19	0.001	100.31
mid24	05/11/14	0.001	1205	Ol	5	39.28	0.033	0.17	0.13	14.67	44.90	0.60	0.24	0.17	0.007	100.20
mid24	05/11/14	0.001	1205	Ol	6	39.53	0.030	0.09	0.09	14.56	45.21	0.62	0.25	0.20	0.001	100.57
mid24	05/11/14	0.001	1205	OI	7	39.37	0.022	0.09	0.13	14.55	44.92	0.61	0.22	0.17	0.012	100.10
mid24	05/11/14	0.001	1205	01	8	39.44	0.030	0.06	0.08	14.73	44.82	0.56	0.28	0.18	0.002	100.18
mid24	05/11/14	0.001	1205		10	39.26	0.035	0.07	0.09	14.63	44.54	0.57	0.31	0.17	0.008	99.00
mid24	05/11/14	0.001	1205	01	11	39.50	0.018	0.00	0.10	14.05	44 59	0.55	0.32	0.18	0.011	99.57
mid24	05/11/14	0.001	1205	Ol	12	40.07	0.021	0.06	0.06	14.75	44.86	0.55	0.20	0.22	0.001	100.80
mid24	05/11/14	0.001	1205	Ol	13	39.77	0.020	0.06	0.09	14.91	44.87	0.59	0.19	0.18	0.003	100.70
mid24	05/11/14	0.001	1205	Ol	14	39.69	0.028	0.07	0.07	14.68	44.97	0.56	0.19	0.22	0.000	100.48
mid24	05/11/14	0.001	1205	Ol	15	39.98	0.040	0.08	0.10	14.55	44.74	0.60	0.24	0.17	0.000	100.50
mid24	05/11/14	0.001	1205	Ol	16	39.63	0.010	0.10	0.12	14.80	44.81	0.55	0.22	0.18	0.001	100.42
mid24	05/11/14	0.001	1205	Ol	17	40.09	0.027	0.08	0.10	14.81	45.23	0.56	0.22	0.19	0.005	101.31
mid24	05/11/14	0.001	1205	Ol	18	40.00	0.028	0.06	0.09	14.83	44.80	0.56	0.24	0.17	0.003	100.79
mid20	01/11/14	0.001	1200	OI	1	39.01	0.018	0.06	0.09	15.39	44.53	0.57	0.29	0.17	0.003	100.14
mid20	01/11/14	0.001	1200		2	39.12	0.022	0.07	0.09	15.42	44.43	0.59	0.24	0.16	0.000	100.15
mid20	01/11/14	0.001	1200	01	3 1	39.40 30.62	0.043	0.15	0.00	15.57	45.97 44.40	0.77	0.24	0.10	0.000	100.19
mid20	01/11/14	0.001	1200	0	- 5	39.40	0.009	0.05	0.10	15.02	44 12	0.58	0.20	0.19	0.000	100.01
mid20	01/11/14	0.001	1200	OI	6	40.10	0.041	0.13	0.16	15.42	44.15	0.61	0.22	0.18	0.004	101.00
mid20	01/11/14	0.001	1200	Ol	7	39.98	0.020	0.09	0.10	15.55	44.33	0.61	0.25	0.18	0.000	101.10
mid20	01/11/14	0.001	1200	Ol	8	39.51	0.039	0.16	0.14	15.10	44.21	0.59	0.24	0.17	0.009	100.17
mid20	01/11/14	0.001	1200	Ol	9	39.76	0.017	0.08	0.09	15.02	44.67	0.59	0.23	0.19	0.002	100.66

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MgO	CaO	MnO	NiO	P_2O_5	Total
mid20	01/11/14	0.001	1200	Ol	10	39.93	0.034	0.07	0.07	14.98	44.22	0.60	0.26	0.18	0.010	100.35
mid20	01/11/14	0.001	1200	Ol	11	39.41	0.015	0.09	0.12	14.62	44.44	0.64	0.31	0.18	0.007	99.83
mid20	01/11/14	0.001	1200	Ol	12	39.62	0.015	0.05	0.12	15.23	44.69	0.55	0.19	0.21	0.006	100.68
mid20	01/11/14	0.001	1200	Ol	13	39.08	0.035	0.11	0.16	15.29	44.68	0.61	0.24	0.17	0.023	100.39
mid20	01/11/14	0.001	1200		14	38.96	0.061	0.15	0.19	15.32	44.46	0.59	0.21	0.18	0.012	100.15
mid20 mid20	01/11/14	0.001	1200		15	39.45 39.47	0.025	0.05	0.08	15.45	44.27	0.59	0.24	0.16	0.004	100.31
mid20	01/11/14	0.001	1200	Ol	17	39.48	0.011	0.05	0.14	15.37	44.19	0.59	0.19	0.18	0.008	100.21
mid20	01/11/14	0.001	1200	Ol	18	39.49	0.024	0.06	0.09	15.48	44.24	0.57	0.26	0.19	0.004	100.43
mid09	15/10/14	0.001	1190	Ol	2	39.50	0.032	0.09	0.08	16.44	42.77	0.62	0.28	0.17	0.005	100.00
mid09	15/10/14	0.001	1190	Ol	3	39.34	0.018	0.07	0.08	17.10	42.93	0.58	0.35	0.16	0.000	100.63
mid09	15/10/14	0.001	1190	Ol	4	39.37	0.007	0.07	0.10	16.74	42.75	0.55	0.25	0.19	0.009	100.04
mid09	15/10/14	0.001	1190		5	39.10	0.021	0.08	0.09	16.48	42.96	0.59	0.26	0.17	0.005	99.80
mid09	15/10/14	0.001	1190	Ol	8	39.30	0.032	0.09	0.10	16.67	43.14	0.50	0.28	0.18	0.000	100.29
mid09	15/10/14	0.001	1190	Ol	9	39.10	0.027	0.12	0.14	16.85	42.80	0.57	0.24	0.19	0.000	100.04
mid12	22/10/14	0.001	1170	Ol	1	39.01	0.041	0.07	0.09	18.50	41.65	0.61	0.29	0.15	0.004	100.41
mid12	22/10/14	0.001	1170	Ol	2	39.40	0.039	0.59	0.06	18.37	40.54	0.91	0.26	0.16	0.010	100.34
mid12	22/10/14	0.001	1170	Ol	4	38.96	0.048	0.22	0.04	18.20	41.45	0.74	0.31	0.16	0.005	100.14
mid12	22/10/14	0.001	1170		5	39.01	0.029	0.08	0.08	18.46	41./1	0.59	0.27	0.17	0.005	100.40
mid12	22/10/14	0.001	1170		7	39.14	0.030	0.00	0.04	18.51	41.85	0.62	0.23	0.15	0.002	100.03
mid12	22/10/14	0.001	1170	Ol	, 9	39.03	0.010	0.05	0.09	18.71	41.63	0.57	0.31	0.17	0.010	100.57
mid12	22/10/14	0.001	1170	Ol	12	39.27	0.040	0.05	0.07	18.97	41.81	0.60	0.30	0.18	0.003	101.28
mid12	22/10/14	0.001	1170	Ol	14	38.85	0.033	0.06	0.06	18.65	41.63	0.62	0.25	0.16	0.002	100.31
mid12	22/10/14	0.001	1170	Ol	15	39.02	0.031	0.09	0.03	18.52	41.34	0.58	0.31	0.19	0.000	100.11
mid12	22/10/14	0.001	1170	Ol	16	39.21	0.037	0.08	0.13	18.47	41.72	0.56	0.31	0.15	0.001	100.68
mid12	22/10/14	0.001	1170		20	38.78	0.032	0.14	0.08	18.36	40.96	0.68	0.23	0.15	0.007	99.41
mid12	22/10/14	0.001	1170		10a 10b	39.42	0.015	0.18	0.03	18.80	41.40	0.69	0.29	0.18	0.005	101.13
mid12	22/10/14	0.001	1170	Ol	c18	38.72	0.027	0.08	0.05	18.27	41.19	0.51	0.26	0.15	0.004	99.26
mid12	22/10/14	0.001	1170	Ol	r18	38.89	0.029	0.07	0.10	18.56	40.92	0.54	0.35	0.13	0.002	99.58
mid31	01/12/14	5	1250	Ol	1	41.58	0.016	0.07	0.14	9.50	49.65	0.38	0.14	0.00	0.001	101.48
mid31	01/12/14	5	1250	Ol	2	41.33	0.023	0.07	0.11	10.78	48.58	0.42	0.17	0.01	0.005	101.48
mid31	01/12/14	5	1250	Ol	3	40.95	0.028	0.07	0.10	10.69	48.69	0.44	0.19	0.04	0.004	101.20
mid31	01/12/14	5	1250	Ol	4	41.17	0.018	0.07	0.12	10.04	48.26	0.43	0.10	0.00	0.010	100.23
mid31	01/12/14	5	1250	Ol	6	41.25	0.028	0.08	0.10	10.06	48.69	0.48	0.17	0.03	0.006	100.89
mid31	01/12/14	5	1250		8	41.45	0.032	0.06	0.14	9.81	49.03	0.46	0.15	0.00	0.007	101.14
mid31	01/12/14	5	1250	Ol	10	41.46	0.043	0.06	0.13	9.65	49.16	0.40	0.11	0.01	0.001	101.10
mid31	01/12/14	5	1250	Ol	5a	41.07	0.030	0.07	0.16	10.71	48.42	0.44	0.13	0.00	0.015	101.04
mid31	01/12/14	5	1250	Ol	5b	40.77	0.017	0.06	0.10	10.63	48.45	0.44	0.18	0.00	0.018	100.66
mid31	01/12/14	5	1250	Ol	5c	41.46	0.009	0.06	0.14	9.77	49.09	0.40	0.11	0.01	0.006	101.06
mid31	01/12/14	5	1250	Ol	9b	41.43	0.025	0.08	0.11	10.44	48.49	0.45	0.12	0.07	0.000	101.22
mid32	03/12/14	5	1230	OI	1	41.12	0.020	0.08	0.10	8.61	49.74	0.44	0.13	0.02	0.002	100.26
mid32	03/12/14	5	1230		3 4	41.80	0.016	0.06	0.11	7.50	50.75 50.94	0.39	0.15	0.01	0.010	100.85
mid32	03/12/14	5	1230	OI	5	41.46	0.036	0.07	0.09	7.63	50.83	0.38	0.09	0.00	0.002	100.59
mid32	03/12/14	5	1230	Ol	6	41.13	0.021	0.07	0.07	8.10	50.42	0.46	0.08	0.02	0.004	100.36
mid32	03/12/14	5	1230	Ol	7	41.41	0.050	0.13	0.14	8.48	49.82	0.49	0.12	0.07	0.006	100.72
mid32	03/12/14	5	1230	Ol	8	42.06	0.027	0.07	0.07	7.15	51.32	0.39	0.02	0.00	0.007	101.12
mid32	03/12/14	5	1230	Ol	9	41.11	0.035	0.13	0.11	9.35	49.06	0.43	0.14	0.07	0.004	100.44
mid32 mid32	03/12/14	5	1230		10	41.26	0.023	0.07	0.12	8.29	50.20	0.47	0.19	0.02	0.009	100.66
mid32	03/12/14	5	1230	01	12	41.53	0.015	0.06	0.08	7.61	50.24	0.42	0.07	0.03	0.005	100.08
mid32	03/12/14	5	1230	Ol	13	41.75	0.027	0.06	0.09	7.57	50.74	0.42	0.10	0.03	0.000	100.79
mid32	03/12/14	5	1230	Ol	15	41.36	0.033	0.05	0.13	7.93	50.39	0.37	0.12	0.03	0.006	100.42
mid32	03/12/14	5	1230	Ol	14a	40.99	0.021	0.05	0.15	8.22	49.95	0.40	0.06	0.02	0.008	99.86
mid32	03/12/14	5	1230	Ol	14b	41.09	0.026	0.07	0.12	7.82	50.46	0.35	0.12	0.03	0.014	100.10
mid35	05/12/14	5	1210	OI	1	40.88	0.021	0.07	0.05	9.90	48.12	0.50	0.13	0.02	0.018	99.71
mid35	05/12/14	5	1210		2	41.04	0.023	0.06	0.09	9.51	48.97	0.46	0.12	0.00	0.004	100.28
mid35	05/12/14	5	1210	01	5 4	40.90	0.037	0.07	0.09	9.42 9.79	48 52	0.50	0.17	0.00	0.000	100.23
mid35	05/12/14	5	1210	Ol	5	41.21	0.043	0.09	0.05	9.73	48.63	0.58	0.17	0.02	0.017	100.53
mid35	05/12/14	5	1210	Ol	6	40.79	0.035	0.09	0.11	10.21	48.44	0.55	0.14	0.00	0.005	100.38
mid35	05/12/14	5	1210	Ol	7	40.97	0.023	0.08	0.10	10.54	47.99	0.56	0.18	0.00	0.007	100.44
mid35	05/12/14	5	1210	Ol	8	40.55	0.025	0.07	0.11	11.61	47.06	0.54	0.23	0.04	0.009	100.23
mid35	05/12/14	5	1210	Ol	9	40.70	0.037	0.06	0.09	10.78	47.48	0.55	0.17	0.02	0.000	99.89
mid35	05/12/14	5	1210		10	40.34	0.035	0.07	0.11	12.29	46.68	0.52	0.19	0.09	0.005	100.33
muss	03/12/14	2	1210	01	11	+1.00	0.023	0.07	0.09	10./1	+/.00	0.33	0.1.3	0.05	0.000	100.46

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al_2O_3	Cr_2O_3	FeO	MgO	CaO	MnO	NiO	P_2O_5	Total
mid35	05/12/14	5	1210	Ol	12	40.58	0.039	0.09	0.10	12.62	46.02	0.54	0.19	0.09	0.009	100.28
mid35	05/12/14	5	1210	Ol	13	40.86	0.028	0.06	0.08	10.94	47.62	0.55	0.17	0.00	0.012	100.31
mid35	05/12/14	5	1210	Ol	14	40.63	0.030	0.09	0.06	10.52	48.12	0.55	0.24	0.00	0.000	100.24
mid35	05/12/14	5	1210	Ol	15	40.33	0.034	0.08	0.13	12.48	46.62	0.51	0.21	0.03	0.004	100.43
mid35	05/12/14	5	1210	Ol	16	40.34	0.035	0.09	0.11	11.29	47.09	0.59	0.18	0.03	0.000	99.75
mid35	05/12/14	5	1210	Ol	17	40.73	0.028	0.06	0.05	11.49	47.28	0.58	0.18	0.01	0.003	100.41
mid35	05/12/14	5	1210	Ol	18	40.77	0.026	0.05	0.11	10.96	47.61	0.58	0.15	0.02	0.003	100.28
mid35	05/12/14	5	1210	OI	19	40.90	0.029	0.07	0.09	10.59	47.17	0.57	0.14	0.00	0.007	99.58
mid35	05/12/14	5	1210		20	40.28	0.023	0.05	0.10	16.82	47.12	0.58	0.18	0.01	0.001	99.97
mid34	05/12/14	5	1190		2	39.40	0.022	0.08	0.05	19.08	42.55	0.62	0.20	0.01	0.000	100.05
mid34	05/12/14	5	1190	01	3	38.78	0.045	0.06	0.12	21.56	39.27	0.53	0.28	0.02	0.004	100.67
mid34	05/12/14	5	1190	Ol	4	39.05	0.033	0.07	0.11	18.13	41.82	0.63	0.21	0.00	0.000	100.05
mid34	05/12/14	5	1190	Ol	5	39.06	0.038	0.09	0.10	17.90	41.67	0.63	0.25	0.04	0.000	99.77
mid34	05/12/14	5	1190	Ol	6	39.25	0.038	0.22	0.06	17.25	41.87	0.67	0.26	0.00	0.005	99.61
mid34	05/12/14	5	1190	Ol	7	38.76	0.055	0.10	0.11	18.70	41.09	0.74	0.23	0.00	0.008	99.79
mid34	05/12/14	5	1190	Ol	8	39.41	0.037	0.08	0.05	17.33	42.40	0.60	0.23	0.01	0.001	100.14
mid34	05/12/14	5	1190	Ol	9	39.10	0.015	0.08	0.08	17.61	42.06	0.62	0.30	0.01	0.011	99.90
mid34	05/12/14	5	1190	OI	10	38.30	0.054	0.08	0.10	21.29	39.17	0.62	0.30	0.00	0.000	99.91
mid34	05/12/14	5	1190		11	39.05	0.024	0.38	0.05	17.98	41.40	0.69	0.23	0.01	0.016	99.84
mid34	05/12/14	5	1190		12	39.20	0.035	0.09	0.09	10.95	40.81	0.64	0.26	0.01	0.025	100.18
mid34	05/12/14	5	1190		14	38.09	0.031	0.07	0.07	20.85	39.86	0.65	0.20	0.03	0.008	99.91
mid34	05/12/14	5	1190	01	15	38.40	0.066	0.08	0.06	22.91	38.18	0.63	0.34	0.01	0.024	100.70
mid34	05/12/14	5	1190	OI	16	38.41	0.052	0.07	0.08	21.41	38.66	0.62	0.34	0.01	0.006	99.67
mid34	05/12/14	5	1190	Ol	17	38.79	0.046	0.11	0.12	20.10	40.16	0.63	0.31	0.02	0.005	100.30
mid34	05/12/14	5	1190	Ol	18	38.79	0.027	0.09	0.03	20.13	40.41	0.66	0.34	0.03	0.010	100.52
mid34	05/12/14	5	1190	Ol	20	38.55	0.028	0.08	0.03	20.98	39.18	0.68	0.30	0.02	0.005	99.87
mid34	05/12/14	5	1190	Ol	21	39.62	0.028	0.07	0.12	16.21	43.03	0.55	0.25	0.02	0.000	99.89
mid34	05/12/14	5	1190	Ol	22	39.27	0.029	0.07	0.08	18.65	41.31	0.54	0.28	0.01	0.000	100.24
mid34	05/12/14	5	1190	Ol	23	39.68	0.024	0.11	0.07	15.89	43.47	0.56	0.28	0.01	0.007	100.10
mid34	05/12/14	5	1190		24	39.07	0.041	0.08	0.06	19.12	40.95	0.52	0.26	0.00	0.002	100.11
mid34	05/12/14	5	1190		25	39.05	0.004	0.10	0.15	19.71	40.55	0.54	0.50	0.00	0.002	100.40
mid34	05/12/14	5	1190	01	20	39.09	0.012	0.07	0.09	19.07	40.68	0.55	0.31	0.00	0.009	99.89
mid34	05/12/14	5	1190	OI	28	39.38	0.033	0.07	0.09	19.12	41.37	0.52	0.29	0.02	0.005	100.89
mid34	05/12/14	5	1190	Ol	29	38.96	0.049	0.10	0.07	20.22	40.03	0.56	0.32	0.00	0.004	100.30
mid34	05/12/14	5	1190	Ol	30	38.95	0.030	0.07	0.05	19.64	40.18	0.56	0.33	0.01	0.013	99.84
	00/11/14	10	1210	01	16	40.20	0.015	0.00	0.12	12.50	45.01	0.42	0.12	0.11	0.011	100.59
mid25	08/11/14	10	1210		10	40.20	0.013	0.09	0.12	13.30	45.91	0.45	0.15	0.11	0.001	100.38
mid25	08/11/14	10	1310	01	18	40.29	0.000	0.09	0.08	13.04	46.35	0.45	0.18	0.15	0.008	101 48
mid25	08/11/14	10	1310	OI	19	40.30	0.007	0.09	0.09	13.53	45.68	0.42	0.19	0.07	0.014	100.39
mid10	21/10/14	10	1300	Ol	c1	39.94	0.014	0.08	0.14	13.22	46.59	0.45	0.21	0.01	0.000	100.65
mid10	21/10/14	10	1300	Ol	c10a	39.55	0.018	0.07	0.16	13.14	46.61	0.45	0.16	0.02	0.000	100.17
mid10	21/10/14	10	1300	Ol	c11a	39.27	0.010	0.07	0.06	12.81	46.43	0.44	0.24	0.02	0.008	99.36
mid10	21/10/14	10	1300	Ol	c11b	39.27	0.013	0.06	0.14	13.20	46.44	0.42	0.20	0.00	0.001	99.75
mid10	21/10/14	10	1300	Ol	c12a	39.72	0.006	0.07	0.14	12.94	46.60	0.46	0.22	0.00	0.008	100.17
mid10	21/10/14	10	1300	Ol	c14a	40.16	0.029	0.06	0.06	13.47	46.51	0.43	0.18	0.01	0.000	100.90
mid10	21/10/14	10	1300		c15a	40.02	0.010	0.07	0.04	12.78	46.36	0.49	0.20	0.00	0.008	99.98
mid10	21/10/14	10	1300		c2b	39.92	0.011	0.07	0.14	12.75	40.22	0.46	0.21	0.01	0.000	99.78
mid10	21/10/14	10	1300		c5h	39.90	0.007	0.07	0.10	13.28	46.79	0.46	0.24	0.01	0.002	100.83
mid10	21/10/14	10	1300	Ol	c8a	39.41	0.005	0.07	0.12	13.00	46.50	0.45	0.17	0.02	0.004	99.74
mid10	21/10/14	10	1300	OI	c9a	39.50	0.016	0.08	0.09	13.08	46.58	0.43	0.23	0.02	0.000	100.02
mid10	21/10/14	10	1300	Ol	r10a	39.59	0.019	0.11	0.14	13.01	46.34	0.50	0.17	0.01	0.002	99.89
mid10	21/10/14	10	1300	Ol	r11a	39.44	0.016	0.07	0.07	13.09	46.50	0.45	0.20	0.01	0.005	99.84
mid10	21/10/14	10	1300	Ol	r11b	39.28	0.020	0.07	0.13	13.03	46.12	0.46	0.19	0.00	0.007	99.30
mid10	21/10/14	10	1300	Ol	r11c	39.37	0.015	0.06	0.13	12.90	46.20	0.44	0.21	0.00	0.000	99.32
mid10	21/10/14	10	1300	Ol	r13a	40.07	0.026	0.07	0.09	12.93	46.53	0.49	0.19	0.02	0.002	100.42
mid10	21/10/14	10	1300	OI	r14a	39.88	0.028	0.07	0.12	13.15	46.00	0.43	0.18	0.01	0.000	99.86
mid10	21/10/14	10	1300		r14b	40.06	0.018	0.07	0.13	13.04	46.81	0.46	0.21	0.01	0.007	100.81
mid10	21/10/14	10	1300		115a r1o	39.90 30.46	0.023	0.07	0.11	12.02	40.00	0.49	0.20	0.00	0.000	100.52 90.02
mid10	21/10/14	10	1300	01	r1h	39.88	0.028	0.07	0.11	12.97	46 65	0.48	0.22	0.02	0.002	100.26
mid10	21/10/14	10	1300	OI	r2a	39.95	0.020	0.07	0.07	13.28	46.47	0.47	0.18	0.00	0.013	100.52
mid10	21/10/14	10	1300	Ol	r2c	39.69	0.021	0.07	0.11	13.15	46.25	0.46	0.28	0.00	0.000	100.03
mid10	21/10/14	10	1300	Ol	r3a	39.83	0.032	0.08	0.08	12.97	46.27	0.49	0.13	0.02	0.000	99.91
mid10	21/10/14	10	1300	Ol	r3b	39.69	0.012	0.07	0.09	13.04	46.10	0.45	0.19	0.00	0.003	99.65
mid10	21/10/14	10	1300	Ol	r4a	39.39	0.030	0.08	0.12	13.04	46.33	0.49	0.22	0.00	0.007	99.71
mid10	21/10/14	10	1300	Ol	r5a	39.73	0.019	0.07	0.13	12.91	46.76	0.46	0.22	0.02	0.000	100.30

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MgO	CaO	MnO	NiO	P_2O_5	Total
mid10	21/10/14	10	1300	Ol	r5b	39.61	0.007	0.07	0.11	13.14	46.77	0.47	0.15	0.00	0.009	100.34
mid10	21/10/14	10	1300	Ol	r5c	39.26	0.016	0.09	0.15	13.24	46.42	0.43	0.17	0.02	0.007	99.80
mid10	21/10/14	10	1300	Ol	r5d	39.70	0.024	0.07	0.12	12.86	46.77	0.45	0.23	0.00	0.005	100.22
mid10	21/10/14	10	1300	Ol	r6a	39.45	0.009	0.11	0.09	12.61	46.49	0.50	0.22	0.01	0.008	99.49
mid10	21/10/14	10	1300	Ol	r7a	39.96	0.021	0.27	0.15	12.91	46.26	0.57	0.17	0.01	0.004	100.32
mid10	21/10/14	10	1300		r8a	39.71	0.013	0.07	0.12	12.92	46.56	0.49	0.20	0.01	0.000	100.09
mid10	21/10/14	10	1300		180 rQa	39.39	0.024	0.07	0.08	12.75	46.17	0.47	0.18	0.00	0.000	99.14
mid22	03/11/14	10	1290	01	1	40.93	0.013	0.11	0.09	13.80	46.98	0.44	0.20	0.01	0.003	102.59
mid22	03/11/14	10	1290	Ol	2	39.79	0.024	0.10	0.12	13.45	45.49	0.44	0.18	0.01	0.000	99.60
mid22	03/11/14	10	1290	Ol	3	39.78	0.023	0.11	0.10	13.45	45.77	0.42	0.21	0.01	0.022	99.89
mid14	24/10/14	10	1280	Ol	1	39.18	0.032	0.10	0.04	17.72	41.93	0.42	0.21	0.15	0.005	99.78
mid14	24/10/14	10	1280	Ol	2	39.93	0.032	0.61	0.07	17.32	40.02	1.37	0.24	0.11	0.001	99.71
mid14	24/10/14	10	1280	Ol	4	39.02	0.024	0.12	0.02	18.57	41.21	0.43	0.29	0.19	0.010	99.89
mid29	20/11/14	10	1280	Ol	4	39.02	0.028	0.09	0.04	18.30	42.42	0.43	0.22	0.13	0.005	100.69
mid14	24/10/14	10	1280	Ol	5	38.72	0.027	0.10	0.07	18.43	41.25	0.48	0.27	0.17	0.000	99.52
mid29	20/11/14	10	1280	OI	6	39.33	0.041	0.16	0.01	17.94	41.59	0.58	0.30	0.14	0.004	100.09
mid14	24/10/14	10	1280		7	39.72	0.024	0.52	0.06	16.01	40.91	1.08	0.20	0.14	0.012	100.14
mid14 mid20	24/10/14	10	1280		8	40.09	0.001	0.62	0.07	17.45	40.00	0.44	0.19	0.07	0.000	100.06
mid14	20/11/14	10	1280		0	39.52	0.034	0.09	0.05	17.45	42.00	0.44	0.20	0.09	0.000	100.55
mid29	29/11/14	10	1280	01	9	39.45	0.032	0.10	0.00	17.59	42.03	0.45	0.23	0.10	0.005	100.20
mid14	24/10/14	10	1280	Ol	10	39.51	0.020	0.12	0.06	16.89	43.00	0.48	0.24	0.11	0.000	100.42
mid29	20/11/14	10	1280	Ol	10	39.22	0.015	0.09	0.05	17.68	43.01	0.41	0.22	0.06	0.005	100.76
mid14	24/10/14	10	1280	Ol	11	39.18	0.016	0.09	0.07	17.37	42.96	0.42	0.24	0.09	0.008	100.43
mid14	24/10/14	10	1280	Ol	12	39.25	0.024	0.10	0.08	17.45	42.61	0.44	0.17	0.12	0.008	100.24
mid14	24/10/14	10	1280	Ol	3b	38.91	0.040	0.18	0.06	18.48	40.44	0.70	0.27	0.21	0.003	99.29
mid27	13/11/14	10	1270	Ol	1	37.73	0.037	0.09	0.05	24.46	37.00	0.44	0.30	0.08	0.004	100.20
mid27	13/11/14	10	1270	Ol	2	38.26	0.046	0.48	0.01	24.29	37.04	0.57	0.24	0.10	0.010	101.05
mid27	13/11/14	10	1270	Ol	3	37.47	0.050	0.11	0.02	24.51	36.80	0.49	0.36	0.06	0.007	99.88
mid27	13/11/14	10	1270	Ol	4	37.59	0.037	0.38	0.07	24.50	36.26	0.67	0.29	0.09	0.003	99.88
mid27	13/11/14	10	1270		5	38.64	0.077	0.57	0.04	23.61	35.72	1.24	0.32	0.06	0.000	100.26
mid27	13/11/14	10	1270		07	37.08	0.052	0.58	0.02	24.50	35.97	0.74	0.37	0.09	0.005	100.07
mid27	13/11/14	10	1270		8	37.90	0.040	0.11	0.03	24.24	36.88	0.42	0.32	0.07	0.001	100.20
mid27	13/11/14	10	1270	01	9	37.71	0.043	0.48	0.02	24.72	35.73	0.62	0.36	0.11	0.011	99.81
mid27	13/11/14	10	1270	Ol	10	37.59	0.047	0.09	0.02	25.30	36.26	0.44	0.30	0.13	0.006	100.16
mid27	13/11/14	10	1270	Ol	11	37.45	0.048	0.08	0.01	25.03	35.97	0.48	0.32	0.13	0.003	99.53
mid23	05/11/14	10	1240	Ol	1	39.17	0.022	0.09	0.05	21.33	39.68	0.44	0.29	0.10	0.001	101.16
mid23	05/11/14	10	1240	Ol	2	38.73	0.014	0.09	0.02	20.69	40.09	0.38	0.27	0.12	0.000	100.41
mid23	05/11/14	10	1240	Ol	3	39.19	0.030	0.10	0.07	21.11	39.82	0.44	0.29	0.09	0.000	101.14
mid23	05/11/14	10	1240	Ol	4	38.65	0.024	0.08	0.07	21.76	39.24	0.44	0.29	0.07	0.008	100.63
mid23	05/11/14	10	1240	Ol	5	38.71	0.015	0.17	0.01	22.19	38.57	0.64	0.27	0.06	0.002	100.62
mid23	05/11/14	10	1240	OI	6	39.10	0.028	0.11	0.02	20.47	39.91	0.41	0.24	0.09	0.001	100.38
mid23	05/11/14	10	1240		/	38.41	0.021	0.07	0.03	21.88	39.27	0.39	0.30	0.11	0.007	100.49
mid23	05/11/14	10	1240		10	39.07	0.034	0.08	0.08	21.00	39.40 40.20	0.39	0.24	0.09	0.003	100.98
mid23	05/11/14	10	1240		11	38.91	0.024	0.09	0.00	21.12	39.48	0.45	0.20	0.08	0.003	100.30
mid23	05/11/14	10	1240	01	13	38.66	0.022	0.08	0.03	21.12	39.28	0.42	0.29	0.04	0.000	100.49
mid23	05/11/14	10	1240	Ol	14	39.24	0.050	0.30	0.03	21.72	38.03	0.81	0.25	0.07	0.006	100.53
mid23	05/11/14	10	1240	Ol	15	39.40	0.026	0.10	0.04	19.67	40.74	0.46	0.26	0.13	0.006	100.84
mid23	05/11/14	10	1240	Ol	16	39.33	0.011	0.08	0.04	19.39	41.33	0.47	0.22	0.09	0.014	100.97
mid23	05/11/14	10	1240	Ol	18	38.84	0.024	0.09	0.05	20.54	39.92	0.41	0.28	0.13	0.009	100.27
mid23	05/11/14	10	1240	Ol	19	38.64	0.031	0.08	0.02	21.76	39.18	0.43	0.30	0.12	0.005	100.57
mid23	05/11/14	10	1240	Ol	20	39.07	0.025	0.11	0.01	21.36	39.49	0.64	0.24	0.11	0.007	101.05
mid23	05/11/14	10	1240	Ol	21	38.46	0.018	0.07	0.03	21.93	38.80	0.42	0.27	0.08	0.004	100.10
mid23	05/11/14	10	1240	Ol	22	39.10	0.058	0.36	0.03	21.16	38.11	1.14	0.31	0.13	0.006	100.39
mid23	05/11/14	10	1240	Ol	23	38.41	0.037	0.09	0.06	21.74	38.90	0.45	0.25	0.12	0.005	100.05
mid23	05/11/14	10	1240		24	38.90	0.031	0.09	0.06	20.79	39.32	0.49	0.29	0.11	0.000	100.07
mid23	05/11/14	10	1240	OI	25	38.57	0.046	0.09	0.04	22.22	38.65	0.47	0.30	0.09	0.021	100.49

 Table C.13 Major element data (wt%) from experimental plagioclase (Pl). No. - analysis number.

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al_2O_3	FeO	MgO	CaO	MnO	Na ₂ O	К 2 О	Total
mid13	23/10/14	0.001	1220	Pl	1	45.98	0.03	33.43	0.46	0.27	17.82	0.000	1.34	0.007	99.34
mid13	23/10/14	0.001	1220	Pl	2	45.23	0.03	33.78	0.61	0.26	17.57	0.000	1.08	0.000	98.56
mid13	23/10/14	0.001	1220	Pl	4	45.76	0.03	33.67	0.51	0.26	17.59	0.000	1.38	0.000	99.20

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al_2O_3	FeO	MgO	CaO	MnO	Na ₂ O	K ₂ O	Total
mid13	23/10/14	0.001	1220	Pl	5	45.60	0.03	33.49	0.52	0.27	17.50	0.023	1.38	0.000	98.83
mid13	23/10/14	0.001	1220	Pl	6	45.76	0.03	33.62	0.58	0.32	18.03	0.000	1.14	0.019	99.49
mid13	23/10/14	0.001	1220	PI PI	/	45.95	0.05	32.62 33.73	0.68	0.49	17.51	0.017	1.39	0.009	98.72
mid13	23/10/14	0.001	1220	Pl	10	45.50	0.02	33.86	0.48	0.27	17.99	0.000	1.16	0.021	99.31
mid08	14/10/14	0.001	1210	Pl	1	46.94	0.07	31.70	1.09	0.89	17.39	0.000	1.39	0.008	99.48
mid08	14/10/14	0.001	1210	Pl	2	46.19	0.15	30.85	1.57	1.45	17.41	0.000	1.26	0.014	98.89
mid08	14/10/14	0.001	1210	Pl	3	45.78	0.04	33.31	0.61	0.37	17.78	0.020	1.38	0.000	99.28
mid08	14/10/14	0.001	1210	PI DI	4	46.43	0.04	32.94	0.58	0.40	17.60	0.000	1.41	0.009	99.40
mid08	14/10/14	0.001	1210	Pl	6	46.04	0.10	31.94	1.18	0.99	17.33	0.000	1.42	0.003	99.03
mid08	14/10/14	0.001	1210	Pl	7	45.58	0.04	33.10	0.68	0.39	17.36	0.033	1.30	0.000	98.49
mid08	14/10/14	0.001	1210	Pl	9	46.15	0.09	31.89	0.82	0.78	17.27	0.003	1.38	0.000	98.39
mid08	14/10/14	0.001	1210	Pl	10	46.23	0.06	32.32	0.75	0.53	17.15	0.000	1.49	0.000	98.54
mid24	05/11/14	0.001	1205	Pl	1	47.08	0.06	32.10	0.71	0.55	17.18	0.042	1.62	0.000	99.35
mid24 mid24	05/11/14	0.001	1205	PI PI	2	46.12	0.03	33.17	0.61	0.35	17.70	0.047	1.50	0.005	99.39
mid24	05/11/14	0.001	1205	Pl	4	48.16	0.09	30.61	1.36	1.03	16.26	0.038	1.86	0.018	99.42
mid24	05/11/14	0.001	1205	Pl	6	49.32	0.05	31.33	0.87	0.73	16.41	0.041	2.03	0.000	100.77
mid24	05/11/14	0.001	1205	Pl	7	46.11	0.04	33.16	0.58	0.32	17.72	0.000	1.24	0.003	99.17
mid24	05/11/14	0.001	1205	Pl	8	47.88	0.06	31.16	0.88	0.96	16.34	0.041	1.89	0.009	99.23
mid24	05/11/14	0.001	1205	Pl	9	45.79	0.04	33.50	0.57	0.26	17.85	0.072	1.27	0.005	99.35
mid24 mid24	05/11/14	0.001	1205	PI DI	10	47.05	0.04	32.09	0.71	0.48	16.98	0.000	1.64	0.014	98.99
mid24	05/11/14	0.001	1205	Pl	12	46.83	0.08	31.76	0.91	0.69	17.45	0.043	1.69	0.017	99.19
mid24	05/11/14	0.001	1205	Pl	13	46.22	0.05	33.41	0.69	0.52	17.74	0.007	1.30	0.000	99.94
mid24	05/11/14	0.001	1205	Pl	14	47.93	0.05	31.24	0.84	0.67	16.58	0.000	1.81	0.001	99.12
mid24	05/11/14	0.001	1205	Pl	15	48.06	0.08	31.23	1.01	0.84	16.67	0.015	1.89	0.022	99.83
mid24	05/11/14	0.001	1205	Pl	16	47.41	0.06	31.34	0.77	0.60	16.78	0.006	1.74	0.000	98.70
mid24 mid24	05/11/14	0.001	1205	PI DI	17	46.28	0.03	32.56	0.75	0.49	17.18	0.026	1.41	0.001	98.74
mid24	05/11/14	0.001	1205	Pl	10	46.04	0.00	33.03	0.59	0.32	17.69	0.033	1.20	0.000	98.92
mid24	05/11/14	0.001	1205	Pl	20	46.33	0.04	32.96	0.61	0.35	17.43	0.000	1.33	0.000	99.05
mid20	01/11/14	0.001	1200	Pl	1	47.57	0.06	31.90	0.87	0.78	16.88	0.035	1.56	0.026	99.68
mid20	01/11/14	0.001	1200	Pl	2	47.20	0.04	32.14	0.71	0.54	17.17	0.005	1.49	0.005	99.31
mid20	01/11/14	0.001	1200	Pl	4	48.01	0.05	30.57	0.92	0.90	16.76	0.037	1.67	0.007	98.92
mid20 mid20	01/11/14	0.001	1200	PI DI	5	46.91	0.04	31.65	0.65	0.52	16.68	0.055	1.68	0.031	98.23
mid20	01/11/14	0.001	1200	Pl	7	48.31	0.05	30.75	0.96	0.88	16.93	0.051	1.79	0.004	99.74
mid20	01/11/14	0.001	1200	Pl	8	47.10	0.03	32.33	0.55	0.38	16.92	0.015	1.62	0.000	98.95
mid20	01/11/14	0.001	1200	Pl	9	45.40	0.03	33.08	0.56	0.25	17.45	0.035	1.39	0.006	98.20
mid20	01/11/14	0.001	1200	Pl	10	47.26	0.08	28.78	1.01	2.45	17.70	0.042	1.51	0.007	98.84
mid20	01/11/14	0.001	1200	Pl	11	47.31	0.06	31.53	0.80	0.70	17.06	0.000	1.58	0.019	99.07
mid20 mid20	01/11/14	0.001	1200	PI PI	12	40.01	0.04	32.31	0.50	0.45	17.05	0.035	1.57	0.024	98.04 98.14
mid20	01/11/14	0.001	1200	Pl	15	45.66	0.03	33.65	0.58	0.26	17.58	0.010	1.21	0.003	98.99
mid20	01/11/14	0.001	1200	Pl	16	46.60	0.06	32.46	0.70	0.54	17.25	0.021	1.34	0.000	98.97
mid20	01/11/14	0.001	1200	Pl	17	47.62	0.08	31.55	1.02	0.81	17.01	0.027	1.51	0.005	99.63
mid20	01/11/14	0.001	1200	Pl	18	46.36	0.03	33.04	0.56	0.35	17.43	0.013	1.40	0.012	99.21
mid09 mid00	15/10/14	0.001	1190	PI	1	48.45	0.08	30.37	1.10	0.80	16.18	0.017	2.13	0.062	99.18
mid09	15/10/14	0.001	1190	PI PI	2	40.81	0.05	28.52	1.66	2.51	16.09	0.033	2.23	0.035	100.33
mid09	15/10/14	0.001	1190	Pl	5	48.60	0.09	30.01	1.26	1.04	16.63	0.006	1.92	0.013	99.56
mid09	15/10/14	0.001	1190	Pl	6	47.08	0.05	31.69	0.67	0.51	17.06	0.014	1.64	0.000	98.71
mid09	15/10/14	0.001	1190	Pl	7	47.68	0.06	31.88	0.80	0.61	17.28	0.029	1.64	0.000	99.99
mid09	15/10/14	0.001	1190	Pl	9	47.98	0.07	30.49	1.08	1.05	17.16	0.025	1.76	0.006	99.62
mid09	15/10/14	0.001	1190	Pl	11	47.69	0.06	30.78	0.85	0.88	16.53	0.033	1.97	0.003	98.80
mid09	15/10/14	0.001	1190	PI DI	13	46.93	0.07	32.01	0.80	0.56	16.96	0.000	1.70	0.021	99.12
mid09	15/10/14	0.001	1190	Pl	14	46.93	0.08	31.61	1.07	0.72	17.03	0.064	1.60	0.000	99.12
mid12	22/10/14	0.001	1170	Pl	3	48.33	0.05	31.33	0.86	0.69	16.55	0.024	2.04	0.009	99.88
mid12	22/10/14	0.001	1170	Pl	5	47.65	0.06	31.88	0.71	0.61	16.74	0.000	1.90	0.000	99.55
mid12	22/10/14	0.001	1170	Pl	6	48.23	0.08	31.34	0.99	0.65	16.27	0.000	2.16	0.006	99.71
mid12	22/10/14	0.001	1170	Pl	7	47.40	0.14	30.76	1.52	1.00	16.81	0.000	1.66	0.000	99.28
mid12	22/10/14	0.001	1170	Pl	8	47.73	0.05	31.97	0.64	0.34	16.87	0.028	1.99	0.025	99.64
mid12	22/10/14	0.001	1170	ri Pl	9 10	47.91 47.50	0.06	31./1	0.81	0.45	10.91 16 70	0.008	1.97	0.012	99.83
mid12	22/10/14	0.001	1170	Pl	12	49.34	0.04	29.84	1.07	0.89	15.67	0.078	2.25	0.004	99.23
mid12	22/10/14	0.001	1170	Pl	13	46.94	0.19	30.41	1.72	1.41	16.94	0.015	1.49	0.000	99.11
mid12	22/10/14	0.001	1170	Pl	14	47.86	0.11	30.94	1.18	0.88	16.69	0.004	1.81	0.006	99.48
mid12	22/10/14	0.001	1170	Pl	15	48.07	0.06	31.62	0.89	0.47	16.26	0.000	2.19	0.019	99.58

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	$\mathrm{Al}_2\mathrm{O}_3$	FeO	MgO	CaO	MnO	Na ₂ O	K ₂ O	Total
mid34	05/12/14	5	1190	Pl	1	50.35	0.09	30.28	0.65	0.86	14.57	0.000	3.31	0.013	100.12
mid34	05/12/14	5	1190	Pl	2	49.99	0.18	28.72	1.25	0.66	14.02	0.009	3.59	0.040	98.47
mid34	05/12/14	5	1190	Pl	3	50.28	0.04	30.87	0.38	0.12	14.32	0.002	3.23	0.029	99.28
mid34	05/12/14	5	1190	Pl	4	49.37	0.07	30.51	1.08	0.42	14.22	0.000	3.28	0.023	98.98
mid34	05/12/14	5	1190	Pl	6	48.15	0.07	32.89	0.51	0.17	16.62	0.032	2.38	0.000	100.82
mid34	05/12/14	5	1190	Pl	9	49.54	0.22	27.76	2.54	0.97	14.50	0.041	3.18	0.012	98.75
mid34	05/12/14	5	1190	Pl	10	50.54	0.12	29.86	0.72	0.38	14.54	0.039	3.47	0.033	99.70
mid34	05/12/14	5	1190	Pl	11	49.73	0.14	29.65	1.02	1.02	14.54	0.036	3.33	0.045	99.51
mid34	05/12/14	5	1190	Pl	13	50.01	0.14	29.94	0.82	0.60	14.44	0.033	3.34	0.021	99.35
mid34	05/12/14	5	1190	Pl	14	50.21	0.12	30.13	0.68	0.58	14.53	0.018	3.37	0.056	99.69
mid34	05/12/14	5	1190	Pl	15	48.67	0.24	27.59	3.88	0.70	13.48	0.061	3.60	0.044	98.26
mid34	05/12/14	5	1190	Pl	16	50.49	0.09	30.38	0.68	0.55	14.47	0.000	3.32	0.026	100.00
mid34	05/12/14	5	1190	Pl	17	48.70	0.06	31.94	0.52	0.25	16.04	0.023	2.34	0.004	99.88
mid34	05/12/14	5	1190	Pl	19	49.69	0.14	29.47	1.92	0.31	14.06	0.058	3.36	0.076	99.11
mid34	05/12/14	5	1190	Pl	20	50.38	0.13	29.91	0.53	1.13	14.68	0.012	3.26	0.034	100.06
mid14	24/10/14	10	1280	Pl	1	48.05	0.02	32.01	0.39	0.21	16.27	0.000	2.26	0.024	99.23
mid14	24/10/14	10	1280	Pl	2	48.36	0.03	31.85	0.29	0.21	16.06	0.058	2.38	0.006	99.24
mid29	20/11/14	10	1280	Pl	2	49.05	0.04	32.05	0.40	0.33	15.80	0.000	2.27	0.030	99.96
mid14	24/10/14	10	1280	Pl	3	48.23	0.02	32.33	0.39	0.19	16.41	0.000	2.10	0.007	99.69
mid29	20/11/14	10	1280	Pl	3	48.58	0.06	32.48	0.76	0.28	15.84	0.047	2.47	0.013	100.54
mid14	24/10/14	10	1280	Pl	4	47.71	0.04	32.00	0.39	0.19	16.48	0.003	2.24	0.019	99.08
mid29	20/11/14	10	1280	Pl	4	49.17	0.03	32.47	0.44	0.18	15.94	0.046	2.32	0.018	100.61
mid14	24/10/14	10	1280	Pl	5	48.10	0.03	31.35	0.39	0.48	15.80	0.001	2.38	0.033	98.57
mid14	24/10/14	10	1280	Pl	6	47.98	0.04	32.24	0.40	0.29	16.17	0.047	2.40	0.021	99.59
mid29	20/11/14	10	1280	Pl	6	48.80	0.04	32.52	0.39	0.30	16.26	0.000	2.21	0.027	100.55
mid14	24/10/14	10	1280	Pl	7	47.74	0.07	29.60	0.71	1.59	16.21	0.000	2.24	0.020	98.17
mid14	24/10/14	10	1280	Pl	8	48.47	0.02	31.83	0.44	0.22	16.21	0.020	2.34	0.024	99.58
mid14	24/10/14	10	1280	Pl	9	48.45	0.04	32.57	0.53	0.36	16.82	0.000	1.93	0.009	100.71
mid27	13/11/14	10	1270	Pl	1	48.82	0.04	31.42	0.39	0.17	15.35	0.000	2.72	0.038	98.96
mid27	13/11/14	10	1270	Pl	2	49.08	0.06	30.86	0.59	0.24	15.01	0.002	2.82	0.018	98.68
mid27	13/11/14	10	1270	Pl	4	48.76	0.05	31.34	0.46	0.14	15.31	0.022	2.62	0.020	98.72
mid27	13/11/14	10	1270	Pl	5	48.92	0.09	30.51	1.02	0.79	15.03	0.016	2.84	0.035	99.25
mid27	13/11/14	10	1270	Pl	6	48.33	0.06	31.28	0.58	0.26	15.59	0.031	2.55	0.024	98.68
mid27	13/11/14	10	1270	Pl	7	48.89	0.05	31.72	0.55	0.17	15.53	0.000	2.65	0.020	99.57
mid27	13/11/14	10	1270	Pl	10	48.25	0.08	31.05	1.05	0.43	15.25	0.020	2.59	0.030	98.75
mid23	05/11/14	10	1240	Pl	1	49.04	0.03	31.91	0.37	0.28	15.61	0.000	2.46	0.017	99.71
mid23	05/11/14	10	1240	Pl	2	49.03	0.07	29.57	0.63	1.35	15.47	0.061	2.13	0.017	98.32
mid23	05/11/14	10	1240	Pl	3	49.39	0.03	31.81	0.31	0.19	15.48	0.000	2.46	0.010	99.67
mid23	05/11/14	10	1240	Pl	4	49.65	0.06	30.84	0.65	1.18	15.94	0.010	2.29	0.007	100.62
mid23	05/11/14	10	1240	Pl	5	49.85	0.02	32.46	0.33	0.19	15.62	0.000	2.52	0.020	101.01
mid23	05/11/14	10	1240	Pl	6	49.39	0.04	32.24	0.38	0.19	15.59	0.007	2.55	0.000	100.40
mid23	05/11/14	10	1240	Pl	7	49.14	0.03	31.89	0.38	0.19	15.52	0.000	2.45	0.029	99.62
mid23	05/11/14	10	1240	Pl	8	50.22	0.04	32.11	0.48	0.25	15.26	0.003	2.55	0.027	100.92
mid23	05/11/14	10	1240	Pl	9	50.07	0.03	32.07	0.33	0.18	15.68	0.011	2.53	0.008	100.90
mid23	05/11/14	10	1240	Pl	10	48.61	0.02	32.58	0.35	0.20	16.30	0.000	2.16	0.028	100.25

Table C.14 Major element data (wt%) from experimental clinopyroxene (Cpx). No. - analysis number.

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	$\mathrm{Al}_2\mathrm{O}_3$	Cr_2O_3	FeO	MgO	CaO	MnO	Na ₂ O	Total
mid20	01/11/14	0.001	1200	Срх	1	52.79	0.32	2.74	0.96	4.66	17.31	21.50	0.16	0.14	100.58
mid20	01/11/14	0.001	1200	Срх	2	52.46	0.32	2.87	1.04	4.74	17.07	21.37	0.15	0.20	100.22
mid20	01/11/14	0.001	1200	Срх	3	51.42	0.55	5.65	1.02	5.44	14.62	20.41	0.12	0.37	99.60
mid20	01/11/14	0.001	1200	Срх	4	52.85	0.35	3.10	0.95	4.64	16.82	20.88	0.07	0.11	99.76
mid20	01/11/14	0.001	1200	Срх	5	51.74	0.33	2.95	0.97	4.54	16.49	21.72	0.12	0.16	99.01
mid20	01/11/14	0.001	1200	Срх	6	52.07	0.36	3.12	1.08	4.61	16.22	21.40	0.12	0.12	99.11
mid20	01/11/14	0.001	1200	Срх	7	51.65	0.40	3.70	1.02	4.55	16.22	21.27	0.10	0.19	99.11
mid20	01/11/14	0.001	1200	Срх	8	51.88	0.39	3.26	1.21	4.57	16.55	21.75	0.08	0.16	99.84
mid20	01/11/14	0.001	1200	Срх	9	52.60	0.32	2.76	0.92	4.60	16.88	21.37	0.10	0.14	99.69
mid20	01/11/14	0.001	1200	Срх	10	51.74	0.39	3.42	1.10	4.19	16.21	21.65	0.12	0.16	98.98
mid20	01/11/14	0.001	1200	Срх	11	51.81	0.42	4.41	1.10	5.58	17.07	20.14	0.19	0.12	100.83
mid20	01/11/14	0.001	1200	Срх	12	51.99	0.33	3.20	1.05	4.80	17.13	21.03	0.15	0.14	99.82
mid20	01/11/14	0.001	1200	Срх	13	51.91	0.32	2.76	0.92	4.64	17.20	21.10	0.16	0.15	99.17
mid20	01/11/14	0.001	1200	Срх	14	53.22	0.29	2.33	0.74	4.62	17.38	21.32	0.13	0.14	100.16
mid20	01/11/14	0.001	1200	Срх	15	52.20	0.36	3.06	1.01	4.65	16.49	21.53	0.08	0.20	99.59

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al_2O_3	Cr ₂ O ₃	FeO	MgO	CaO	MnO	Na ₂ O	Total
mid20	01/11/14	0.001	1200	Срх	16	52.24	0.38	3.24	1.16	4.33	16.48	21.93	0.05	0.17	99.97
mid20	01/11/14	0.001	1200	Срх	17	52.28	0.35	3.14	1.07	4.64	16.96	21.32	0.12	0.16	100.04
mid20	01/11/14	0.001	1200	Срх	18	53.37	0.31	2.45	0.81	4.49	16.88	21.75	0.10	0.18	100.34
mid20 mid20	01/11/14	0.001	1200	Cpx	20	52.81	0.32	2.56	0.90	4.53	10.70	21.52	0.07	0.16	99.63
mid12	22/10/14	0.001	1170	Срх	1	52.05	0.58	4.65	0.53	5.79	15.12	20.74	0.10	0.41	99.96
mid12	22/10/14	0.001	1170	Срх	2	52.17	0.59	4.84	0.69	6.08	15.59	20.63	0.07	0.32	100.99
mid12	22/10/14	0.001	1170	Cpx	3	48.46	1.00	18.76	0.04	10.36	5.36	14.47	0.15	1.69	100.29
mid12	22/10/14	0.001	1170	Срх	4	48.98	1.25	12.65	0.20	10.76	8.90	14.40	0.25	1.38	98.76
mid12	22/10/14	0.001	1170	Срх	5	50.94	0.77	8.08	0.65	7.00	13.35	18.33	0.05	0.68	99.84
mid12	22/10/14	0.001	1170	Срх	6	48.76	1.14	13.36	0.21	10.74	8.35	15.49	0.19	1.48	99.72
mid12	22/10/14	0.001	1170	Срх	8	51.55	0.52	2.01	0.75	5.49	15.72	21.32	0.15	0.28	99.78
mid12	22/10/14	0.001	1170	Срх	9	52.10	0.43	3.06	0.63	5.17	15.62	21.05	0.11	0.23	98.40
mid12	22/10/14	0.001	1170	Срх	10	52.00	0.61	5.40	0.73	5.69	14.87	20.55	0.16	0.31	100.31
mid12	22/10/14	0.001	1170	Срх	11	52.36	0.46	2.90	0.82	5.38	16.54	21.27	0.12	0.14	99.99
mid12	22/10/14	0.001	1170	Срх	12	51.61	0.61	4.09	0.60	6.12	15.96	21.20	0.15	0.13	100.47
mid12	22/10/14	0.001	1170	Срх	13	52.00	0.42	3.72	0.92	5.06	15.67	21.43	0.12	0.20	99.54
mid12	22/10/14	0.001	1170	Срх	14	49.93	1.15	11.01	0.19	10.39	10.15	15.88	0.17	1.32	100.19
mid12	22/10/14	0.001	1170	Срх	15	49.17	1.34	23.31	0.13	5.70	4.74	13.81	0.15	1.41	08.43
mid12	22/10/14	0.001	1170	Срх	17	51.85	0.72	5.49	0.60	6.03	14.75	20.54	0.19	0.42	100.59
mid12	22/10/14	0.001	1170	Срх	18	49.23	1.32	13.28	0.09	12.01	8.08	14.32	0.26	1.55	100.14
mid12	22/10/14	0.001	1170	Срх	19	49.02	1.32	14.22	0.11	11.75	7.55	13.99	0.26	1.65	99.87
mid12	22/10/14	0.001	1170	Срх	20	51.69	0.78	7.21	0.47	6.92	13.08	19.87	0.13	0.45	100.61
mid12	22/10/14	0.001	1170	Срх	21	49.30	1.09	18.36	0.09	9.54	6.08	14.73	0.13	1.67	100.98
mid35	05/12/14	5	1210	Срх	2	51.31	0.91	4.96	0.32	7.20	16.11	19.15	0.19	0.36	100.51
mid35	05/12/14	5	1210	Срх	3	49.94	0.94	7.33	0.73	3.16	14.61	21.99	0.00	0.60	99.31
mid35	05/12/14	5	1210	Срх	4	51.11	0.97	5.73	0.34	6.46	14.95	21.09	0.08	0.52	101.25
mid35	05/12/14	5	1210	Срх	5	50.32	1.11	5.71	0.39	6.79	14.51	19.92	0.16	0.40	99.32
mid35	05/12/14	5	1210	Срх	0	51.02	0.96	0.00	0.50	4.18	15.01	10.64	0.08	0.50	100.85
mid35	05/12/14	5	1210	Срх	8	51.49	0.76	5.87	0.60	4.05	15.47	21.90	0.05	0.58	100.77
mid35	05/12/14	5	1210	Срх	9	49.50	1.03	7.87	0.27	6.36	16.09	18.59	0.32	0.37	100.40
mid35	05/12/14	5	1210	Cpx	10	50.03	1.05	6.66	0.37	5.66	15.40	20.70	0.25	0.52	100.63
mid35	05/12/14	5	1210	Срх	11	50.16	0.94	6.98	0.64	4.81	14.71	20.78	0.08	0.52	99.62
mid35	05/12/14	5	1210	Срх	12	50.00	0.91	6.43	0.47	4.45	14.56	21.86	0.09	0.59	99.36
mid35	05/12/14	5	1210	Срх	13	51.83	0.41	3.82	0.57	5.45	16.98	21.03	0.11	0.07	100.29
mid34	05/12/14	5	1210	Срх	14	50.25 40.70	1.11	5.67	0.55	4.15	14.54	18 74	0.08	0.60	00.20
mid34	05/12/14	5	1190	Срх	2	50.17	1.44	4.99	0.29	8.63	14.59	19.07	0.18	0.33	99.93
mid34	05/12/14	5	1190	Срх	4	50.81	1.19	5.61	0.40	7.47	14.34	20.02	0.11	0.44	100.39
mid34	05/12/14	5	1190	Срх	5	50.45	0.67	9.80	0.36	4.66	12.76	20.27	0.12	1.20	100.27
mid34	05/12/14	5	1190	Срх	6	49.35	0.98	7.84	0.46	6.52	13.61	19.46	0.16	0.74	99.13
mid34	05/12/14	5	1190	Срх	7	50.57	1.36	5.01	0.25	8.82	14.79	19.41	0.21	0.43	100.85
mid34	05/12/14	5	1190	Срх	8	49.80	1.15	9.81	0.28	6.93	11.60	18.92	0.19	0.94	99.62
mid34 mid34	05/12/14	5	1190	Cpx	10	50.86 48.76	1.43	4.50	0.26	8.88	14.73	19.18	0.22	2.00	00.44
mid34	05/12/14	5	1190	Срх	14	50.47	1.39	5.82	0.21	9.60	12.99	18.67	0.20	0.41	99.76
mid34	05/12/14	5	1190	Срх	15	50.53	1.50	4.88	0.28	9.64	14.52	18.32	0.33	0.31	100.30
mid34	05/12/14	5	1190	Срх	16	46.54	0.93	4.65	0.21	10.62	19.30	15.86	0.28	0.61	98.99
mid34	05/12/14	5	1190	Срх	17	50.34	1.52	4.71	0.35	9.41	14.60	18.81	0.29	0.33	100.35
mid34	05/12/14	5	1190	Срх	18	50.01	1.21	10.80	0.28	5.94	11.22	18.73	0.17	1.14	99.49
mid34	05/12/14	5	1190	Срх	20	50.30	1.60	4.96	0.33	9.36	14.34	18.92	0.29	0.35	100.45
mid25	08/11/14	10	1310	Срх	1	52.09	0.32	6.28	0.53	4.94	17.94	18.72	0.12	0.27	101.22
mid25	08/11/14	10	1310	Срх	2	51.53	0.33	6.34	0.44	4.87	17.74	18.73	0.13	0.30	100.43
mid25	08/11/14	10	1310	Срх	4	51.70	0.28	5.23	0.40	5.15	18.21	17.89	0.09	0.31	99.27
mid25	08/11/14	10	1310	Cpx	5	51.91	0.29	5.14	0.39	5.26	18.36	17.80	0.18	0.26	99.60
mid25	08/11/14	10	1310	Срх	0 7	51.95 50.24	0.29	4.77	0.43	3.04 4.92	16.23	17.91	0.19	0.25	99.00 100 34
mid25	08/11/14	10	1310	Cpx	8	51.36	0.37	6.86	0.49	5.11	17.42	18.51	0.10	0.32	100.54
mid25	08/11/14	10	1310	Срх	9	52.55	0.31	5.28	0.48	5.07	18.39	18.12	0.17	0.26	100.63
mid25	08/11/14	10	1310	Срх	10	52.83	0.29	4.87	0.31	5.47	18.28	18.15	0.14	0.23	100.58
mid25	08/11/14	10	1310	Срх	12	49.80	0.50	8.77	0.45	4.75	16.02	18.82	0.14	0.30	99.55
mid25	08/11/14	10	1310	Срх	13	49.84	0.46	8.50	0.43	5.21	16.16	18.47	0.10	0.27	99.44
mid25	08/11/14	10	1310	Срх	14	49.32	0.50	8.78	0.41	5.06	15.96	18.41	0.22	0.31	98.96
mid25	08/11/14	10	1310	Срх	15	49.73	0.54	9.52	0.40	5.07	15.89	19.28	0.07	0.50	100.67
mid25	08/11/14	10	1310	CDX	17	49.08	0.52	9.52	0.49	4.49	15.89	18.85	0.17	0.27	99.22

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MgO	CaO	MnO	Na ₂ O	Total
mid25	08/11/14	10	1310	Срх	18	48.42	0.56	10.17	0.58	4.26	15.44	19.04	0.13	0.31	98.91
mid25	08/11/14	10	1310	Cpx	19	48.69	0.56	10.03	0.63	4.42	15.50	19.05	0.14	0.32	99.34
mid25	08/11/14	10	1310	Cpx	20	48.68	0.54	10.09	0.62	4.45	15.33	19.46	0.09	0.29	99.53
mid22 mid22	03/11/14	10	1290	Cpx	2	48.20	0.60	11.40	0.42	4.75	15.45	18.61	0.05	0.30	99.79
mid22	03/11/14	10	1290	Срх	3	48.57	0.60	11.66	0.46	4.60	15.05	18.83	0.10	0.30	100.18
mid22	03/11/14	10	1290	Срх	4	48.17	0.62	11.63	0.40	4.47	15.03	19.31	0.15	0.27	100.05
mid22	03/11/14	10	1290	Cpx	5	48.61	0.59	11.00	0.52	4.80	15.65	18.13	0.23	0.28	99.81
mid22	03/11/14	10	1290	Cpx	6	47.68	0.82	12.30	0.35	4.54	14.57	19.38	0.13	0.27	100.05
mid22 mid22	03/11/14	10	1290	Cpx	7	48.98	0.64	10.15	0.34	4.75	15.98	18.55	0.15	0.30	99.83
mid22	03/11/14	10	1290	Срх	9	48.19	0.86	11.54	0.32	4.50	14.99	19.47	0.13	0.29	100.49
mid22	03/11/14	10	1290	Срх	10	48.23	0.81	11.35	0.53	4.44	15.19	18.57	0.17	0.29	99.59
mid22	03/11/14	10	1290	Cpx	11	48.00	0.85	11.96	0.42	4.71	14.96	19.00	0.09	0.30	100.29
mid22	03/11/14	10	1290	Cpx	12	48.82	0.58	10.49	0.48	4.47	15.44	19.54	0.19	0.32	100.34
mid22 mid22	03/11/14	10	1290	Cpx	13	47.84	0.62	11.31	0.47	4.37	15.08	19.54	0.13	0.30	99.67
mid22	03/11/14	10	1290	Cpx	14	48.48	0.39	12.26	0.43	4.44	13.51	19.37	0.12	0.32	100.23
mid14	24/10/14	10	1280	Срх	2	50.12	0.36	6.52	0.86	5.59	16.39	20.17	0.16	0.13	100.25
mid14	24/10/14	10	1280	Срх	3	49.19	0.70	8.40	0.32	6.66	16.21	18.27	0.16	0.19	100.11
mid14	24/10/14	10	1280	Cpx	4	51.94	0.28	4.09	0.66	4.61	16.85	21.27	0.10	0.18	99.97
mid14	24/10/14	10	1280	Cpx	5	49.67	0.43	8.78	0.50	5.61	16.65	18.20	0.17	0.33	100.33
mid14	24/10/14	10	1280	Срх	0	49.65	0.58	8.78	0.40	5.48	15.58	19.28	0.15	0.57	00.28
mid14	24/10/14	10	1280	Срх	10	49.63	0.53	8.70	0.28	5.19	15.80	18.92	0.07	0.36	99.47
mid14	24/10/14	10	1280	Cpx	13	49.66	0.52	8.47	0.40	5.06	15.59	19.18	0.08	0.33	99.29
mid14	24/10/14	10	1280	Срх	14	49.47	0.53	8.71	0.39	5.26	15.83	19.36	0.10	0.36	100.01
mid14	24/10/14	10	1280	Cpx	16	49.79	0.53	8.81	0.35	5.47	15.67	18.82	0.13	0.40	99.97
mid14	24/10/14	10	1280	Срх	1/	49.31	0.53	8.63	0.37	5.50	15.60	19.25	0.14	0.33	99.67
mid14	24/10/14	10	1280	Срх	20	49.87	0.59	8.65	0.30	5.35	15.62	19.17	0.15	0.32	100.23
mid27	13/11/14	10	1270	Срх	1	49.39	0.80	8.24	0.27	7.23	15.16	18.48	0.19	0.39	100.14
mid27	13/11/14	10	1270	Срх	2	48.67	0.74	8.15	0.30	6.73	15.33	17.98	0.15	0.34	98.39
mid27	13/11/14	10	1270	Cpx	3	49.11	0.80	7.57	0.21	7.88	15.75	16.73	0.22	0.39	98.66
mid27	13/11/14	10	1270	Срх	4	48.30	0.98	8.04	0.17	7.99	14.77	17.79	0.19	0.44	98.68
mid27	13/11/14	10	1270	Срх	7	48.87	1.00	8.44	0.14	7.93	14.83	18.13	0.26	0.41	100.06
mid27	13/11/14	10	1270	Срх	8	48.67	0.83	8.30	0.23	7.41	15.21	18.02	0.22	0.41	99.29
mid27	13/11/14	10	1270	Срх	9	49.33	0.88	7.18	0.19	8.44	14.98	17.52	0.16	0.51	99.18
mid27	13/11/14	10	1270	Срх	10	48.12	0.88	8.42	0.22	7.67	15.24	17.51	0.18	0.41	98.66
mid27	13/11/14	10	1270	Срх	11	48.89	0.63	8.30	0.33	6.04	15.35	18.11	0.22	0.42	98.88
mid27	13/11/14	10	1270	Срх	12	51.05	0.30	4.58	0.85	4.77	16.50	21.22	0.10	0.28	99.66
mid27	13/11/14	10	1270	Срх	14	49.14	0.52	8.70	0.31	6.23	15.47	18.29	0.17	0.42	99.25
mid27	13/11/14	10	1270	Срх	15	49.52	0.48	7.93	0.40	6.28	15.72	18.64	0.11	0.31	99.40
mid27	13/11/14	10	1270	Срх	16	49.25	0.46	8.33	0.50	6.42	15.81	18.71	0.06	0.27	99.82
mid27 mid23	13/11/14	10	1270	Срх	17	49.16	0.44	5.88	0.56	5.84	15.50	19.02	0.08	0.20	99.20
mid23	05/11/14	10	1240	Срх	2	50.27	0.52	7.33	0.30	7.14	16.05	17.99	0.12	0.32	100.38
mid23	05/11/14	10	1240	Срх	3	51.07	0.44	6.91	0.54	6.13	16.15	18.87	0.25	0.31	100.68
mid23	05/11/14	10	1240	Cpx	4	50.84	0.37	7.08	0.92	6.07	16.49	18.64	0.06	0.23	100.70
mid23	05/11/14	10	1240	Cpx	5	50.41	0.56	7.51	0.26	6.85	16.01	18.30	0.11	0.34	100.35
mid23 mid23	05/11/14	10	1240	Срх	6	50.78	0.83	8.26 5.18	0.21	8.77	15.16	16.23	0.17	0.60	101.01
mid23	05/11/14	10	1240	Срх	8	50.11	0.46	6.77	0.31	5.92	16.20	18.69	0.13	0.21	98.96
mid23	05/11/14	10	1240	Срх	9	51.04	0.45	5.81	0.27	7.02	16.34	17.92	0.29	0.28	99.42
mid23	05/11/14	10	1240	Cpx	10	50.29	0.49	7.88	0.26	5.72	15.70	18.85	0.16	0.30	99.65
mid23	05/11/14	10	1240	Cpx	11	50.68	0.32	5.39	0.42	4.76	17.50	19.26	0.13	0.25	98.69
mid23 mid23	05/11/14	10	1240	Cpx	12	49.68 50.16	0.79	9.00	0.23	9.21 5.55	13.92	16.64	0.11	0.59	100.17
mid23	05/11/14	10	1240	Срх	14	50.45	0.49	7.56	0.30	5.88	16.09	19.07	0.10	0.28	100.22
mid23	05/11/14	10	1240	Ċpx	15	51.06	0.44	6.22	0.37	6.42	16.51	18.22	0.17	0.28	99.68
mid23	05/11/14	10	1240	Срх	16	52.44	0.30	5.41	0.26	5.32	17.79	18.84	0.13	0.29	100.76
mid23	05/11/14	10	1240	Срх	17	47.67	1.50	13.35	0.12	10.15	12.50	15.10	0.20	0.69	101.29
mid23 mid23	05/11/14	10	1240	Cpx	18	50.28	0.34	5.90 8.62	0.86	5.20 6.84	1/.22	18.76	0.13	0.23	99.85
mid23	05/11/14	10	1240	Срх	20	47.59	1.17	14.44	0.11	11.00	10.22	14.42	0.22	1.19	100.36
mid23	05/11/14	10	1240	Срх	21	51.09	0.42	7.35	0.39	4.97	16.74	19.41	0.16	0.27	100.80
mid23	05/11/14	10	1240	Cpx	22	51.02	0.40	7.22	0.43	5.26	16.64	19.39	0.17	0.27	100.82
mid23	05/11/14	10	1240	Срх	23	52.02	0.30	5.43	0.75	5.11	17.48	19.02	0.10	0.24	100.45
mid23	05/11/14	10	1240	Срх	24	50.60	0.55	8.11	0.40	6.09	15.68	19.14	0.14	0.33	101.04

Experiment	Date	Р	Т	Phase	No.	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MgO	CaO	MnO	Na ₂ O	Tota
mid23	05/11/14	10	1240	Срх	25	50.72	0.46	7.58	0.38	5.63	16.04	19.17	0.11	0.30	100.40

Table C.15 Major element standards of experimental glass (Gl), olivine (Ol), plagioclase (Pl), and clinopyroxene (Cpx) for EPMA sessions. All in wt%.

Standard	Session	Phase	SiO ₂	TiO ₂	Al_2O_3	Cr ₂ O ₃	FeO	MgO	CaO	MnO	NiO	Na ₂ O	к20	P_2O_5	Total
B113498	March	Gl	50.22	4.10	12.25	0.00	12.87	5.09	8.75	0.28	0.08	2.76	0.90	0.46	97.79
B113498	March	Gl	50.16	4.16	12.24	0.04	13.35	5.03	9.01	0.26	0.01	2.78	0.96	0.49	98.51
B113498	March	Gl	50.75	4.16	12.32	0.04	13.06	4.84	9.01	0.20	0.00	2.70	0.91	0.45	98.47
B113498	March	Gl	50.48	4.10	12.13	0.00	13.21	4.98	9.25	0.17	0.00	2.58	0.88	0.44	98.25
B113498	March	Gl	50.26	4.15	12.10	0.00	13.27	4.92	8.88	0.24	0.07	2.67	0.90	0.45	97.93
B113498	March	GI	50.41	4.11	12.10	0.00	13.23	4.97	9.02	0.20	0.00	2.72	0.85	0.43	98.06
B113498	March	GI	50.09 40.82	4.11	12.15	0.00	12.00	4.98	8.95	0.22	0.00	2.69	0.90	0.44	97.52
B113498	Inly	GI	49.85 50.94	4.17	12.04	0.00	13.64	5.08	8.90	0.23	0.00	2.02	0.87	0.40	97.02
B113498	July	GI	51.60	4.15	12.32	0.00	13.19	5.04	9.19	0.22	0.00	2.78	0.84	0.47	99.83
B113498	July	Gl	50.40	4.17	12.10	0.01	13.18	4.94	8.77	0.23	0.00	2.55	0.81	0.42	97.62
B113498	July	Gl	50.97	4.11	12.19	0.01	13.46	5.02	9.19	0.22	0.01	2.72	0.91	0.40	99.24
B113498	July	Gl	51.08	4.11	12.11	0.03	13.27	4.86	8.79	0.21	0.03	2.61	0.90	0.40	98.43
B113498	August	Gl	50.67	4.21	12.48	0.05	13.49	4.88	9.20	0.22	0.00	2.70	0.85	0.43	99.19
B113498	August	Gl	50.65	4.19	12.51	0.03	13.59	4.91	8.98	0.23	0.00	2.75	0.89	0.47	99.22
B113498	August	Gl	50.57	4.22	12.48	0.00	13.67	4.99	9.13	0.17	0.03	2.66	0.89	0.48	99.33
B113498	August	Gl	50.39	4.17	12.48	0.02	13.60	4.85	9.19	0.14	0.00	2.72	0.91	0.45	98.95
B113498	August	GI	50.45	4.19	12.43	0.00	13.73	5.08	9.06	0.24	0.00	2.73	0.93	0.46	99.34
B113498	August	GI	50.65	4.18	12.30	0.03	13.74	4.92	8.89	0.20	0.00	2.58	0.89	0.47	98.89
SanC-Ol	March	Ol	40.72	0.02	0.04	0.00	9.79	49.37	0.10	0.16	0.37	na	na	0.01	100.58
SanC-Ol	March	Ol	40.56	0.00	0.04	0.02	10.13	49.63	0.11	0.21	0.39	na	na	0.01	101.09
SanC-OI	March		40.82	0.01	0.03	0.05	9.//	49.52	0.12	0.13	0.37	na	na	0.01	100.63
SanC-Ol	March		40.85	0.00	0.04	0.02	9.85	49.58	0.11	0.13	0.38	na	na	0.01	100.96
SanC-Ol	March	01	40.88	0.01	0.04	0.01	10.05	49.05	0.11	0.15	0.38	na	na	0.01	100.55
SanC-Ol	March	01	40.70	0.00	0.04	0.02	9.82	49.61	0.10	0.17	0.38	na	na	0.01	100.85
SanC-Ol	March	OI	40.67	0.00	0.04	0.00	9.96	49.31	0.10	0.15	0.39	na	na	0.00	100.61
SanC-Ol	July	Ol	41.09	0.00	0.04	0.00	9.90	48.88	0.15	0.16	0.36	na	na	0.01	100.58
SanC-Ol	July	Ol	41.16	0.00	0.04	0.03	9.98	49.19	0.14	0.11	0.37	na	na	0.01	101.02
SanC-Ol	July	Ol	40.55	0.01	0.04	0.00	9.81	47.94	0.52	0.17	0.38	na	na	0.01	99.43
SanC-Ol	July	Ol	41.16	0.00	0.04	0.00	9.70	49.28	0.12	0.13	0.38	na	na	0.02	100.83
SanC-Ol	July	Ol	40.86	0.00	0.03	0.02	9.74	48.63	0.12	0.16	0.35	na	na	0.02	99.93
SanC-Ol	August	Ol	40.34	0.00	0.04	0.00	10.00	48.38	0.12	0.18	0.39	na	na	0.02	99.47
SanC-Ol	August		40.22	0.00	0.04	0.00	9.83	47.96	0.13	0.14	0.36	na	na	0.01	98.70
SanC-OI	August		40.55	0.00	0.04	0.05	9.42	48.38	0.15	0.15	0.38	na	na	0.02	99.08
SanC-Ol	August		40.83	0.00	0.04	0.01	9.05	48.71	0.20	0.17	0.37	na	na	0.00	100.31
-	. inguist		10.05	0.00	0.01	0.00	0.50			0.21	0.07			0.01	
An13/041	March	PI	43.31	0.00	34.61	na	0.50	0.01	18.94	0.00	na	0.54	0.04	na	97.95
An137041	March	PI DI	43.07	0.00	34.83	na	0.40	0.01	19.25	0.00	na	0.56	0.05	na	99.00
An137041	March	Pl	44 12	0.02	34.87	na	0.40	0.01	18.70	0.01	na	0.50	0.04	na	98.84
An137041	March	Pl	43.78	0.02	34.96	na	0.48	0.02	18.98	0.00	na	0.55	0.04	na	98.83
An137041	March	Pl	43.74	0.01	34.97	na	0.43	0.02	19.14	0.00	na	0.60	0.01	na	98.93
An137041	March	Pl	43.57	0.01	34.88	na	0.52	0.01	18.98	0.00	na	0.60	0.00	na	98.57
An137041	July	Pl	44.56	0.00	34.85	na	0.54	0.01	19.19	0.00	na	0.53	0.02	na	99.71
An137041	July	Pl	44.43	0.02	34.76	na	0.53	0.01	19.20	0.00	na	0.54	0.02	na	99.52
An137041	July	Pl	44.33	0.01	35.04	na	0.45	0.01	19.10	0.00	na	0.52	0.02	na	99.48
An137041	July	Pl	44.56	0.01	35.49	na	0.54	0.01	19.15	0.00	na	0.60	0.01	na	100.35
An137041	July	Pl	44.27	0.00	35.11	na	0.54	0.01	19.30	0.00	na	0.57	0.03	na	99.82
An13/041	July	PI	44.96	0.01	35.13	na	0.49	0.02	18.96	0.04	na	0.52	0.04	na	100.16
An13/041	August	PI DI	45.60	0.01	34.31 24.44	na	0.47	0.02	19.39	0.00	na	0.51	0.02	na	98.34
An137041	August	F1 Pl	42 52	0.00	33.76	ne ne	0.50	0.02	19.14	0.01	na	0.47	0.02	ne	96.04 96.24
An137041	August	PI	43 71	0.01	34 58	ne	0.50	0.02	19.42	0.00	na	0.50	0.04	ne	98 77
An137041	August	Pl	43.53	0.01	34.54	na	0.46	0.02	19.32	0.03	na	0.55	0.03	na	98.48
An137041	August	Pl	43.78	0.01	34.52	na	0.47	0.01	19.26	0.00	na	0.45	0.02	na	98.53
Di117733	August	Cpx	55.86	0.01	0.11	0.06	0.17	18.26	25.56	0.04	na	0.11	na	na	100.18
Di117733	August	Cpx	55.67	0.00	0.10	0.03	0.23	18.21	25.62	0.00	na	0.10	na	na	99.95
Di117733	August	Ċpx	55.46	0.01	0.15	0.02	0.33	18.00	25.84	0.03	na	0.12	na	na	99.95

Standard	Session	Phase	SiO_2	TiO_2	Al_2O_3	Cr_2O_3	FeO	MgO	CaO	MnO	NiO	Na ₂ O	K_2O	P_2O_5	Total
Di117733	August	Срх	55.72	0.00	0.09	0.02	0.22	18.19	25.77	0.04	na	0.13	na	na	100.19
Di117733	August	Cpx	55.33	0.02	0.14	0.06	0.23	18.08	25.34	0.00	na	0.14	na	na	99.34
Aug164905	August	Cpx	50.79	0.52	7.60	0.86	4.80	17.21	17.44	0.07	na	0.85	na	na	100.13
Aug164905	August	Cpx	51.23	0.50	7.62	0.79	4.68	17.15	17.44	0.07	na	0.90	na	na	100.38
Aug164905	August	Cpx	51.38	0.50	7.57	0.94	4.67	17.09	17.50	0.13	na	0.83	na	na	100.62
Aug164905	August	Cpx	50.39	0.51	7.55	1.02	4.93	17.00	17.41	0.18	na	0.83	na	na	99.83
Aug164905	August	Срх	50.67	0.49	7.66	0.85	4.73	17.06	17.40	0.12	na	0.87	na	na	99.85