Geochemical, Isotopic, and Mineralogical
Characterization of the Frobisher Suite Mafic-
Ultramafic Sills, Southern Baffin Island, Nunavut, with
Implications for Metallogenic Potential

by

Dustin Andrew Liikane, B.Sc.H.

A thesis submitted to the Faculty of Graduate Studies and Research

in partial fulfillment of the requirements for the degree of

Master of Science

Ottawa-Carleton Geoscience Centre and

Department of Earth Sciences

Carleton University

Ottawa, Ontario

© 2017

Dustin Andrew Liikane

Dedication

In loving memory of my brother, Matthew Rein Liikane, who supported me throughout the duration of this thesis project, both in this world, and from the next.

April 2, 1993 - March 22, 2015

Frontispiece

Layered Mafic-Ultramafic sill, Markham Bay

An aerial photograph of one of the more majestic-looking layered mafic-ultramafic sills, located in Markham Bay, Meta Incognita Peninsula, southern Baffin Island, Nunavut. The sill is overturned, such that the mafic unit forms the structural base of the intrusion, and the thin ultramafic component is situated on top. The sill is hosted in metasedimentary quartzite (white weathering - bottom) and psammite (rusty-brown weathering – top) units of the Lake Harbour Group.

Abstract

The Frobisher Suite is a newly recognized magmatic province comprised of mafic-ultramafic sills in southern Baffin Island, Nunavut. The sills are emplaced into the metasedimentary strata of the middle Paleoproterozoic Lake Harbour Group, and vary in thickness from several meters to hundreds of meters. Based on major and trace element chemistry, the Frobisher Suite can be divided into eight geochemical groups. The major geochemical groups indicate melting occurred to differing degrees within a mantle plume head, at various depths below the Meta Incognita microcontinent at ca. 1900 Ma. A number of sills formed *via* injections of multiple, distinct magma types, some of which were influenced by subcontinent lithospheric mantle. Disseminated sulphides are present in the sills, and in some cases, the adjacent host psammite. At this stage, the Frobisher Suite may be considered to be part of its own Large Igneous Province (LIP), termed the “Frobisher LIP.”

Acknowledgements

First off, I would like to graciously thank my thesis supervisors Dr. Richard Ernst (Carleton University), Dr. Marc St-Onge (Geological Survey of Canada), and Dr. Bruce Kjarsgaard (Geological Survey of Canada) for all of their guidance, support, patience and encouragement throughout the duration of this M.Sc. project.

Thank you to all the enthusiastic field assistants and fellow mappers, most of whom assisted with sample collection: Timothy Chadwick, Dr. Owen Weller, Dr. Brenden Dyck, Dr. Diane Skipton, Terry Milton, Sean Noble-Nowdluk, Tyler Rowe, Angela Ford, Harry Ireyak, Alia Bigio, and Randy Hinanik. Thank you to Jerry Nuttall and Gerard Hartery from Universal Helicopters for their safe and professional air-support in the field. Thanks to the camp cook, Debra Guilfoyle, for providing high quality and diverse meals, daily, which kept me fuelled for the two summers of fieldwork. Thanks to the Polar Continental Shelf Program for providing logistical support.

A special thanks to Holly Steenkamp (CNGO) and David Mate (CNGO) for overseeing the process for obtaining thin sections and bulk-rock geochemical data. Thanks to Celine Gilbert for data-management in the field, as well as later correspondence with regard to Inuit place names, used to name the Killapait and Chidliak sills. Thanks to Micheal Burns (GSC) for carefully preparing samples (e.g. trimming, slabbing) and thin section blocks prior to shipment. Thanks to Vancouver Petrographic for providing quality thin sections and to ALS Chemex for obtaining precise and accurate geochemical data from the samples. Thank

you to microprobe technician Peter Jones for his time and expertise in obtaining mineral geochemistry and back-scatter electron images.

A big thanks to Nicole Rayner (GSC) for assistance in attempting to date the Frobisher suite. While not every attempt panned out, due to lack of igneous zircons, the minimum crystallization age of 1900 Ma obtained from magmatic zircons in the Killapait sill was key in the regional comparison and modelling sections of the thesis. Also, thank you to the other geochronology staff for their contributions to the U-Pb dating process of the Frobisher suite, specifically, Julie Peressini, Dr. Bill Davis, Tom Pestaj, Pat Hunt, and Carole Lafontaine.

I greatly appreciate the thorough and insightful reviews of the Summary of Activities paper [i.e. [Liikane et al. \(2015\)](#)] by Drs. Thomas Skulski (GSC) and Tony Peterson (GSC), and thanks to Tammy Sifer for assistance formatting the text and figures for publication.

A special thanks to Angela Ford for her GIS expertise and assistance with geological map preparation. Thanks to Douglas Oneschuk for plotting maps and conference posters.

Thank you to Dr. Natasha Wodicka for taking the time to sit down with me and discuss the evolution of the syn-magmatic sedimentary basin adjacent to the Meta Incognita microcontinent, and thanks to Dr. Fred Gaidies for discussing the metamorphism of mafic to ultramafic rocks. Thank you to PhD candidates Nico Kastek and Chris Rogers for discussions on geochemistry.

A special thanks to the Geological Survey of Canada (GSC), Canada Nunavut Geoscience Office (CNGO), Carleton University, Ernst Geosciences, and the LIPs-Supercontinent Reconstruction Project for funding this project.

Thank you to defence committee members Dr. Richard Ernst, Dr. James Mungall, Dr. Sarah Dare, and Dr. Ian Honsberger for their participation in the defence process, and for insightful comments and suggestions.

Lastly, I would like to thank friends and family for their support throughout the duration of my master's degree program. And thank you to my girlfriend, Pamela Iraheta-Muñiz, for her unwavering support and encouragement during the writing process of the thesis.

Method of Presentation

The body of the thesis is presented in four chapters (Chapters 3 & 5-7), with each focusing on separate aspects of the Frobisher suite mafic, ultramafic, and mafic-ultramafic sills of southern Baffin Island, Nunavut. They are prefaced by an introduction (Chapter 1) and background (Chapter 2), and followed by a discussion, conclusions and future recommendations (Chapter 8). Between Chapters 3 and 5 there is also an overview of the methods used in the completion of this thesis (Chapter 4).

Chapter 1 – A brief introduction to the study area, and the purpose and objectives of the study.

Chapter 2 – A detailed background of: (1) Large Igneous Provinces (LIPs), describing all of their components and models of formation; (2) Layered intrusions, including their geometry, distribution and mechanisms of formation; (3) Mafic-ultramafic intrusion-hosted mineral deposits, mainly focused on Ni-Cu-PGE mineralization; and (4) Metamorphism of rocks mafic to ultramafic in composition.

Chapter 3 – Introduces the Meta Incognita microcontinent and the Paleoproterozoic Trans-Hudson orogen. The regional geology of the Québec-Baffin Island segment of the orogen is first presented, followed by its tectonic architecture, deformation and metamorphic events, and finally, the Frobisher suite sills and their Lake Harbour Group host rocks.

Chapter 4 – Overview of the methods used in completing the thesis, including sample collection and processing, and analytical methods.

Chapter 5 – Detailed petrography of the Frobisher suite sills, where mineral assemblages and textures are described.

Chapter 6 – Mineral chemistry of the Frobisher suite sills.

Chapter 7 – Detailed geochemical study of the Frobisher suite sills, where eight geochemical groups are recognized and described individually, with some comparison between groups.

Chapter 8 – Discussion of the general characteristics and petrogenesis of each geochemical group, the petrogenesis and models for the formation of the two mafic-ultramafic sills sampled in detail, the candidacy for the Frobisher suite being considered a LIP, evidence for mantle plume-involvement versus passive rift magmatism, post-emplacement modification, regional implications, a model of formation for the Frobisher suite, and its metallogenic potential. This is followed by the summary of the main conclusions of the study, and future considerations.

Appendices: Appendix A contains sample coordinates, Appendix B contains field and petrographic data (Chapter 5), Appendix C contains bulk-rock geochemical data (Chapter 7), Appendix D contains mineral geochemical data (Chapter 6), and Appendix E contains Sm-Nd isotopic data (Chapter 7).

Table of Contents

Frontispiece	iii
Abstract.....	iv
Acknowledgements.....	v
Method of Presentation.....	viii
Table of Contents.....	x
List of Appendices.....	xviii
List of Tables	xix
List of Figures	xx
Chapter 1 - Introduction	1
1.1 Aim of Study.....	2
Chapter 2 – Background	4
2.1 Large Igneous Provinces and their Plumbing Systems.....	5
2.1.1 Sub-lithospheric mantle source distribution and depth.....	6
2.1.1.1 Edge-driven convection model	6
2.1.1.2 Lithospheric delamination model	7
2.1.1.3 Mantle plume model	7
2.1.1.4 Passive rifting model.....	9
2.1.2 Lithospheric entry points	10
2.1.3 Intra-lithospheric distribution	10
2.1.3.1 Dykes versus sills	10
2.1.3.2 General characteristics of sills	12
2.1.3.3 Sill provinces.....	14
2.1.4 Surface magmatism	15
2.2 Layered Intrusions.....	16
2.2.1 Geometry of layered intrusions	17
2.2.1.1 Stratiform (sill-like) intrusions	18
2.2.1.2 Dyke-like intrusions.....	18

2.2.1.3 Funnel-shaped intrusions	19
2.2.1.4 Chonoliths	20
2.2.2 Distribution of layered intrusions in LIPs	21
2.2.3 Mechanisms of layer formation in layered intrusions	21
2.2.3.1 Fractional crystallization	21
2.2.3.2 Composite sills versus vertically-stacked sills	23
2.3 Mafic-Ultramafic Intrusion-Hosted Mineral Deposits	27
2.3.1 Ni-Cu-PGE deposit petrogenesis	29
2.3.1.1 Geodynamic history and setting	29
2.3.1.2 Sources and compositions of ore-forming magmas	29
2.3.1.3 Sources of sulphur and transport of sulphide liquid.....	30
2.3.1.4 Mineralized intrusion formation	31
2.4 Metamorphism of Mafic and Ultramafic Rocks.....	32
Chapter 3 – Regional Geology	60
3.1 Geological Framework	61
3.2 Deformation Events and Metamorphism.....	66
3.3 Lake Harbour Group.....	70
3.4 Frobisher Suite Mafic, Ultramafic and Mafic-Ultramafic Sills.....	74
3.4.1 Mafic units.....	77
3.4.2 Ultramafic units	78
Chapter 4 – Methods	122
4.1 Fieldwork and Sample Collection	123
4.1.1 Regional Sampling	124
4.1.2 Detailed Sampling	124
4.2 Sample Selection and Preparation.....	125
4.3 Petrography.....	126
4.4 Photography and Microphotography	127
4.5 Analytical Methods	128

4.5.1 Whole Rock Geochemistry	128
4.5.2 Mineral Geochemistry	133
4.5.3 Isotope Geochemistry	135
4.6 Potential Error in Geochemical Analyses	139
Chapter 5 – Petrography	154
5.1 General Observations	155
5.1.1 Mineral assemblages	155
5.1.2 Textures	156
5.1.3 Metamorphism	157
5.2 Mineralogical Characteristics of Geochemical Groups	158
5.2.1 Group A1	158
5.2.2 Group A2	159
5.2.3 Group B	159
5.2.4 Group C1	160
5.2.5 Group C2	161
5.2.6 Group D	161
5.2.7 Group E	162
5.2.8 Group F	162
5.3 Silicate Mineralogy	163
5.3.1 Olivine	163
5.3.2 Clinopyroxene	164
5.3.3 Orthopyroxene	164
5.3.4 Amphibole	165
5.3.5 Plagioclase	165
5.3.6 Phlogopite	166
5.4 Oxide Mineralogy	166

5.4.1 Spinel.....	166
5.4.2 Magnetite and ilmenite.....	166
5.5 Ore Mineralogy.....	167
5.5.1 Pyrrhotite	167
5.5.2 Chalcopyrite.....	168
5.5.3 Pentlandite	168
5.5.4 Pyrite	168
5.6 Mineralogy of Detailed Sample Transects Across Two Mafic-Ultramafic Sills ...	168
5.6.1 Chidliak mafic-ultramafic sill.....	168
5.6.2 Killapait mafic-ultramafic sill.....	169
Chapter 6 – Mineral Chemistry.....	190
6.3 Mineral Chemistry.....	191
6.3.1 Silicate Mineral Chemistry.....	191
6.3.1.1 Olivine	191
6.3.1.2 Clinopyroxene.....	192
6.3.1.3 Orthopyroxene	193
6.3.1.4 Amphibole	193
6.3.1.5 Plagioclase	194
6.3.1.6 Phlogopite	194
6.3.2 Oxide Mineralogy	195
6.3.2.1 Spinel.....	195
6.3.3 Ore Mineralogy	195
6.3.3.1 Pyrrhotite	196
6.3.3.2 Chalcopyrite.....	196
6.3.3.3 Pentlandite	196

6.3.3.4 Pyrite.....	196
6.3.4 Mineralogy of Detailed Sample Transects Across Two Mafic-Ultramafic Sills.....	197
6.3.4.1 Chidliak mafic-ultramafic sill	197
6.3.4.2 Killapait mafic-ultramafic sill.....	197
Chapter 7 – Geochemistry	220
7.1 Whole Rock Geochemistry	221
7.1.1 Element mobility.....	222
7.1.2 Classification through chondrite-normalized rare-earth element plots	223
7.1.3 Major elements	228
7.1.3.1 Major and minor oxides versus Mg#	228
7.1.3.2 High- and low-Ti groups.....	230
7.1.3.3 FeO versus MgO	231
7.1.4 Trace elements	232
7.1.4.1 Trace elements versus Mg#.....	232
7.1.4.2 Primitive mantle-normalized spider diagrams	234
7.1.4.3 MORB-normalized plots.....	236
7.1.4.4 Immobile trace element ratio discrimination	237
7.1.4.5 Alkaline versus tholeiitic.....	238
7.2 Sm-Nd Isotopes Geochemistry	240
7.2.1 Epsilon-Nd (ϵ Nd)	241
7.2.2 ϵ Nd versus $^{147}\text{Sm}/^{144}\text{Nd}$ diagram	242
7.3 Geochemistry of Detailed Sample Transects Across Two Mafic-Ultramafic Sills	243
7.4 Interpretation.....	244
7.4.1 Single injection versus multiple injection origin of layered sills.....	244

7.4.2 Mineral fractionation	247
7.4.2.1 Olivine and pyroxene fractionation	248
7.4.2.2 Plagioclase fractionation	249
7.4.2.3 Apatite fractionation	251
7.4.2.4 Fe-Ti-oxide fractionation	253
7.4.3 Contamination	254
7.4.4 Depth and degree of partial melting	258
7.4.5 Magma mixing	261
7.4.5.1 Group A1-like Group A2 samples	261
Chapter 8 – Discussion & Conclusion	311
8.1 Petrogenesis	312
8.1.1 Petrogenesis of major geochemical groups	312
8.1.1.1 High-Ti magma type (Group D)	312
8.1.1.2 Low-Ti magma types (Groups A1 & A2)	315
8.1.2 Petrogenesis of minor geochemical groups	317
8.1.2.1 Groups B, C1, and C2 magma types	317
8.1.2.1.1 Group C2 - an in-depth look	319
8.1.2.2. Group E magma type	321
8.1.2.3 Group F magma type	321
8.1.3 Potential parental magma compositions	323
8.1.4 Petrogenesis of the Killapait sill	323
8.1.4.1 Models of Killapait sill formation	327
8.1.5 Petrogenesis of the Chidliak sill	330
8.1.5.1 Models of Chidliak sill formation	332
8.1.6 Candidacy for a Large Igneous Province (LIP)	335

8.1.6.1 Areal Extent	335
8.1.6.2 Volume of magma	336
8.1.6.3 Duration and pulsed nature of magmatism	336
8.1.6.4 Intraplate character	336
8.1.6.5 Provinciality within Frobisher suite	337
8.1.6.6 Broad geochemical similarities to LIPs	338
8.1.6.7 LIP versus non-LIP?	338
8.1.7 Evidence for mantle plume-involvement	339
8.1.7.1 Field evidence	339
8.1.7.2 Geochemical evidence	339
8.1.8 Sill emplacement	342
8.1.9 Post-emplacement modification	345
8.1.10 Metamorphic Character of sills	346
8.2 Regional Implications.....	347
8.2.1 Regional comparison with other LIP and non-LIP events	347
8.3 Model for Frobisher Suite Formation.....	350
8.3.1 Pre-Stage 1	350
8.3.2 Stage 1	351
8.3.3 Stage 2.....	351
8.3.4 Stage 3.....	351
8.3.5 Stage 4.....	351
8.4 Mineral Deposit Potential of Frobisher Suite	352
8.4.1 Field and petrographic observations	353
8.4.1.1 14SAB-L009 gossan	354
8.4.2 Geochemical perspective	355

8.4.2.1 Chalcophile element geochemistry	356
8.5 Conclusions	359

List of Appendices

Appendix A	Sample Coordinates	407
Appendix B	Field and Petrographic Data	409
Appendix C	Bulk-Rock Geochemical Data.....	412
Appendix D	Mineral Geochemical Data	417
Appendix E	Sm-Nd Isotopic Data.....	436

List of Tables

Table 4.1	List of Analytes, their lower detection limits and units.....	151
Table 4.2	Elements, X-ray lines and standards for EMP.....	153
Table 5.1	Comparison of mineral contents.....	189
Table 6.1	Ranges in Fo-and Ni-content in olivine.....	216
Table 6.2	Ranges in Di- and Cr-content in clinopyroxene.....	217
Table 6.3	Ranges in En- and Cr-content in orthopyroxene.....	218
Table 6.4	Ranges in An-content in plagioclase.....	219
Table 7.1	Number of occurrences of geochemical groups	309
Table 7.2	Relationship in occurrences between geochemical groups.....	310
Table 8.1	Details of LIPs which occurred between 1845 - 1945 Ma.....	404
Table 8.2	Comparison of Frobisher Suite chemistry with other major LIPs.....	405

List of Figures

Figure 2.1	Karoo-LIP sill distribution in sedimentary basin	55
Figure 2.2	Saucer shape of sill	57
Figure 2.3	Plume emplacement beneath Archean craton.....	58
Figure 3.1	Summary geology map of North America and Greenland.....	94
Figure 3.2	Simplified geological map of Canadian Shield and west Greenland.....	96
Figure 3.3	Geological terrane map of Quebec-Baffin segment of THO.....	98
Figure 3.4	Tectonic summary diagram for Quebec-Baffin segment of THO	100
Figure 3.5	Dome-and-basin patterns in bedrock geology of Meta Incognita Peninsula.....	102
Figure 3.6	Lake Harbour Group units and 14SAB-L009 gossan	104
Figure 3.7	Photographs of layered mafic-ultramafic sill and Chidliak sill	106
Figure 3.8	Bedrock map of Meta Incognita Peninsula	108
Figure 3.9	Bedrock map of northern Hall Peninsula	111
Figure 3.10	Concordant intrusive contact.....	114
Figure 3.11	Sill morphology.....	116
Figure 3.12	Zircon photographs and concordia diagrams	118
Figure 3.13	Gabbro, peridotite and pyroxenite sill components	120
Figure 4.1	Major oxide data from bulk-geochemistry standards	145
Figure 4.2	Trace element data from bulk-geochemistry standards	147
Figure 4.3	Base metal data from bulk-geochemistry standards	149
Figure 5.1	IUGS-based classification diagrams	171
Figure 5.2	Textures in hand sample and thin section	173
Figure 5.3	Textures in thin section.....	175
Figure 5.4	Reaction textures and exsolution laminae.....	177
Figure 5.5	Ore mineral photomicrographs	179
Figure 5.6	Mineralogical changes through the Chidliak sill	181
Figure 5.7	Mineralogical and textural changes through the Chidliak sill.....	183
Figure 5.8	Mineralogical changes through the Killapait sill	185
Figure 5.9	Mineralogical and textural changes through the Killapait sill.....	187
Figure 6.1	Ni content in olivine compared to Fo content	201
Figure 6.2	MnO content in olivine compared to Fo content	202
Figure 6.3	Clinopyroxene classification	203

Figure 6.4	Orthopyroxene classification	205
Figure 6.5	Amphibole classification	207
Figure 6.6	Plagioclase classification	209
Figure 6.7	Phlogopite classification	211
Figure 6.8	Spinel classification	213
Figure 6.9	Spinel composition	215
Figure 7.1	Geochemical group legend	274
Figure 7.2	Assessment of element mobility	275
Figure 7.3	Normalized trace-element spiderdiagrams	277
Figure 7.4	LREE and HREE slope comparison	281
Figure 7.5	Major oxide variation with Mg#	282
Figure 7.6	High- and low-Ti groups.....	283
Figure 7.7	FeO versus MgO	284
Figure 7.8	Trace-element variation with Mg#	285
Figure 7.9	Discriminating ternary diagrams	288
Figure 7.10	Alkaline versus tholeiitic.....	290
Figure 7.11	ϵNd versus $^{147}\text{Sm}/^{144}\text{Nd}$; L057B02 contamination	291
Figure 7.12	Distinguishing contamination from source enrichment.....	293
Figure 7.13	Geochemical variation through the Killapait sill	294
Figure 7.14	Geochemical variation through the Chidliak sill	296
Figure 7.15	Chondrite-normalized REE pattern comparison within sills.....	298
Figure 7.16	Europium anomalies	300
Figure 7.17	Recognizing contamination and changes in melting depth.....	301
Figure 7.18	Further distinguishing contamination from source enrichment.....	302
Figure 7.19	Recognizing crustal contamination.....	303
Figure 7.20	Group C2 compared to upper crustal units	304
Figure 7.21	Depth of melting	305
Figure 7.22	Depth of melting	306
Figure 7.23	La/Sm versus Yb.....	307
Figure 7.24	Magma mixing.....	308
Figure 8.1	Comparison between Groups A2 & B, and C1 & D.....	376
Figure 8.2	Xenoliths and plagioclase megacrysts in Killapait sill	378
Figure 8.3	Pods of Fe oxide within Killapait sill	380

Figure 8.4	Model #1 of formation for Killapait sill	381
Figure 8.5	Model #2 of formation for Killapait sill	383
Figure 8.6	Metapyroxenitic xenolith in metaperidotite within Chidliak sill	385
Figure 8.7	Model #1 of formation for Chidliak sill	386
Figure 8.8	Model #2 of formation for Chidliak sill	388
Figure 8.9	Distribution of LIPs of similar age	390
Figure 8.10	Comparison of Frobisher Suite with Circum-Superior LIP.....	392
Figure 8.11	Relative position of lithotectonic elements at ca. 1900 Ma	394
Figure 8.12	Cross-section from Superior craton through to Meta Incognita microcontinent at ca. 1900 Ma	396
Figure 8.13	Model of formation of Frobisher Suite	398
Figure 8.14	Mineralized hand samples from 14SAB-L009	400
Figure 8.15	Chalcophile element variation.....	402
Figure 8.16	Log(Cu/Pd) versus Mg# diagram	403

Chapter 1 - Introduction

1.1 Aim of Study

Prior to the initiation of this study, very few mafic to ultramafic sills within southern Baffin Island had been documented and sampled, and no systematic attempt had been made to understand this potentially economically-interesting group of rocks. The aim of this study was to characterize the widespread, yet understudied, suite of mafic, ultramafic, and layered mafic-ultramafic sills distributed throughout southern Baffin Island, which I have informally named the “Frobisher suite,” based on their geographic proximity to Frobisher Bay. This research involved the integration of two summers of fieldwork on Baffin Island, during which regional mapping was conducted, with quantitative petrographic, geochemical, and metallogenic characterization of the mafic to ultramafic units. In addition, unpublished U-Pb geochronology was integrated to constrain the timing of this magmatism and propose a time-accurate model.

The results of this thesis suggest that the Frobisher suite represents a geochemically-diverse, plume-generated large igneous province (LIP), which may correlate with already established LIPs elsewhere around the globe. The Frobisher suite is compared to the Circum-Superior LIP, which is proximal to Baffin Island and host to a number of mafic-ultramafic bodies containing significant nickel-copper-PGE mineralization (Ni-Cu-PGE; e.g., Cape Smith Belt). With the recognition that this suite may represent a plume-generated LIP, the conclusions of this thesis may be taken a step further by others, to improve the understanding of the distribution and interaction of tectonic plates during the

middle Paleoproterozoic (ca. 1.9 billion years ago), and thus, improve existing tectonic and tectono-magmatic models for the eastern Canadian Arctic.

Given the context of the study, a comprehensive background is warranted. In the background (Chapter 2), four important topics of relevance to the thesis are covered: (1) Large igneous provinces (LIPs); (2) Layered intrusions; (3) Mafic-ultramafic intrusion-related mineralization; and (4) Metamorphism of mafic and ultramafic rocks. This is followed by an overview of the regional geology of southern Baffin Island, including its tectonic setting, in Chapter 3.

Chapter 2 – Background

A substantial amount of this background was compiled from [Cawthorn \(2012\)](#), [Zientek, \(2012\)](#), [Ernst \(2014\)](#), [Barnes et al. \(2015\)](#), [Maier et al. \(2015\)](#), and [Bucher and Grapes \(2011\)](#).

2.1 Large Igneous Provinces and their Plumbing Systems

It has been recognized that throughout the Earth's history, magmatic events characterized by the rapid emplacement of mafic, ultramafic, and less often, silicic magma within intraplate or incipient plate margin settings have occurred. The magmatic events are termed "Large Igneous Provinces (LIPs)" ([Coffin and Eldholm, 1992](#)), and have most recently been defined as "mainly mafic (+ultramafic) magmatic province[s] with areal extent[s] > 0.1 Mkm² and igneous volume[s] > 0.1 Mkm³, that [have] intraplate characteristics, and [are] emplaced in a short duration pulse, or in some cases multiple pulses (less than 1-5 Ma), with a maximum duration of < c. 50 Ma" ([Ernst, 2014](#)).

LIPs may comprise continental (and ocean basin) flood basalts, volcanic rifted margins, oceanic plateaus, and feeder systems (in deeply eroded Archean to Proterozoic terrains), with the latter including sill provinces, layered intrusions and regional dyke swarms ([Ernst et al., 2008](#)). The plumbing system for LIPs can be described in terms of the different levels of magma transport from mantle source to paleo-surface, which includes the sub-lithospheric mantle source distribution and depth, lithospheric entry points, intra-lithospheric distribution, and surface magmatism ([Ernst et al., 2008](#)).

2.1.1 Sub-lithospheric mantle source distribution and depth

The anomalously high magma fluxes required to form LIPs can be produced through all, or a combination of, three mantle processes: 1) increasing the temperature of the source (e.g. mantle plume); 2) decreasing the pressure acting on the source material (decompression melting); and 3) increased fertility and volatile content of the source to generate high degrees of melt (e.g. subduction zones; [Saunders, 2005](#)). The first two processes, in combination, are key for voluminous melt generation, which is implicit in the plume model ([Saunders, 2005](#)). Impacts from extraterrestrial bodies can also have the ability to produce the vast amounts of melt required to form a LIP ([Saunders, 2005](#)), but this is exceptionally rare (e.g. Sudbury impact structure, Canada) and requires considering the impact melt sheets of crust as part of the LIP (J.Mungall, pers. comm., 2017). Other models for LIP generation include edge-driven convection, lithospheric delamination, plume + lithospheric delamination, passive rifting, and asthenospheric convection.

2.1.1.1 Edge-driven convection model

[King and Anderson \(1995\)](#) recognized that many continental flood basalts lie close to Archean cratonic margins, and proposed an edge driven convection (EDC) model for LIP formation. It involves the craton thermally insulating the underlying asthenosphere, raising its temperature. The hot asthenosphere then slides out from under the thick craton, towards areas of thinner lithosphere. The

mantle then decompresses and melts as it continues to ascend, producing large volumes of magma. [King and Anderson \(1998\)](#) argue that this ‘edge-driven convection’ model also involves a secondary convection cell active along the cratonic margin, which leads to greater magma production.

2.1.1.2 Lithospheric delamination model

[Elkins-Tanton \(2005\)](#) proposed an alternative model (‘delamination model’) whereby the lithospheric mantle becomes more dense than adjacent lithospheric mantle through processes including conversion to eclogitic phase assemblages, melt injection, or by thickening and cooling of a lithospheric root (e.g. [Elkins-Tanton \(2005\)](#); [Lustrino, 2005](#)). As the lithosphere detaches and sinks into the mantle, asthenospheric material flows into the ensuing space and melts extensively (adiabatically). Substantial uplift also occurs as the lithosphere rebounds. Additionally, hydrous magmas may be produced by fluid-induced melting as the delaminated lithospheric material dehydrates, releasing fluids as it progressively heats and descends into the mantle ([Elkins-Tanton, 2005](#)). However, it is debatable whether normal-temperature asthenosphere is capable of producing the required volumes of magma and high Mg-melts ([Saunders, 2005](#)).

2.1.1.3 Mantle plume model

In the plume model ([Morgan, 1971](#); [Campbell and Griffiths, 1989](#); [Richards et al., 1989](#); [Campbell, 2001](#); [Campbell, 2005](#); [Campbell, 2007](#)), hot,

deep mantle material buoyantly rises, convectively, in a plume and melts at a shallow depth ([White, 2010](#)). Many plumes are believed to be rooted at the core-mantle boundary (a thermal boundary), however, they need not be rooted there, as they are also believed to spawn at different levels within the lower mantle. The large voluminous heads (or small heads) ascend to shallow depths, extensively melt, and trigger voluminous basaltic eruptions. The initiation of plume upwelling may be caused by cold mantle downwellings, or perhaps, disturbances at the boundary due to plunging oceanic slabs ([White, 2010](#)). In other cases, plumes may have very small heads, or are completely headless, which is believed to be responsible for small oceanic plateau formation (e.g. the Azores). The idea of differing shapes and sizes of mantle plumes is supported by the modelling of thermochemical plumes ([Farnetani & Samuel, 2005](#)). In addition to individual plumes, superplumes originating from the deep mantle, and plumes originating from the upper mantle-lower mantle transition zone are also recognized. An alternative origin of plumes was proposed by [Davies & Bunge \(2006\)](#), who showed that plumes may arise from cold mantle downwellings plunging into hotter mantle regions, initiating upwelling plumes on the fringes of hot regions.

The melting products of plume material are significantly higher in temperature (200-300°C, typically) compared to mid-ocean ridge basalt (MORB), which are derived from melting of upper mantle material. This is supported by observed negative seismic velocity anomalies beneath 30 hotspots ([Montelli et al., 2004](#)), which implies significantly hotter mantle exists in these regions ([White,](#)

2010). The anomalies extend into the deep mantle, and in some cases, down to the core-mantle boundary (Burke & Torsvik, 2004).

Plumes are generally heterogeneous, containing both incompatible-element-enriched and depleted components. However, the depleted plume material is chemically distinct from typical mid-ocean ridge basalt (MORB) source, in most cases (Chadwick et al., 2005). Some hypotheses proposed for the origin of the geochemical signature of plumes include the recycling of ocean crust (Hofmann & White, 1982) and the recycling of sub-continental lithospheric mantle (McKenzie & O’Nions, 1983).

Plumes are not completely stationary, moving relative to one another, but far less than lithospheric plates do (White, 2010). Mantle plumes may have the ability to affect plate motion and initiate continental breakup, but do not drive plate motion (White, 2010).

2.1.1.4 Passive rifting model

Numerous flood basalt provinces were linked to decompression melting at the onset of crustal extension by White and McKenzie (1989). However, these models required a thermal anomaly ~50-200°C greater than that of ambient mantle. A later paper by van Wijk et al., (2001) considered the high beta-factor (i.e. high crustal stretching) rifting influence on a narrow zone (175 km), and concluded that sufficient volumes of magma could be produced even at normal upper mantle temperatures (1333°C), consistent with magma production produced along rifted margins of the North Atlantic Igneous Province (NAIP). This

model suggests volcanic rifted margins do not always require a plume, however, the evidence for an initial pulse of pre-rifting LIP magmatism is not addressed (Ernst, 2014).

2.1.2 Lithospheric entry points

There are a variety of ways in which melts from the underlying asthenospheric mantle can initially enter the lithosphere. The magmas can reach shallow levels via active or failed rifts (e.g. triple-junctions rifts) or other trans-lithospheric fracture zones (e.g. Begg et al., 2009, 2010). In some cases, sub-lithospheric channeling of magma to “thin spots” within the lithosphere or at the edge of cratons may occur (e.g. Oyarzun et al., 1997; Ebinger and Sleep, 1998; Duggen et al., 2009; Ernst, 2014).

2.1.3 Intra-lithospheric distribution

2.1.3.1 Dykes versus sills

The fundamental architecture of the magma transport network arises from dyke and sill propagation and linkage as local crustal structure directs magma transport (Kavanagh, et al., 2006; Menand, 2008; Bungler and Cruden, 2011; Magee et al. 2016). Magma ascends via the upward propagation of dykes, where buoyancy of the magma and overpressure at the tips of the dykes are large enough to drive upward crack propagation (Lister and Kerr, 1991). Dyke swarms also have the ability to transport magma laterally for distances >2000 km and may feed distal volcanic or sill provinces (e.g. 2215 Ma Nipissing-Ungava event,

Canada). Dykes may reorient to form sills, and in a similar manner dykes may emerge from sills, controlled by the changing regional stress field. Within sills, magma may also be transported great distances, laterally (e.g. potentially up to 4100 km within mafic sill complexes; [Cartwright and Hansen, 2006](#); [Leat, 2008](#)). The orientation of these sheets of magma is also influenced and controlled by propagating sills and dykes interacting with pre-existing mechanical anisotropies, such as foliations, lithological contacts and fractures ([Barnes et al., 2015](#)). The majority of sills have been emplaced in sedimentary basins ([Figure 2.1](#)), but seismic images have identified sills hosted by crystalline basement ([Mandler and Clowes, 1997](#); [Welford and Clowes, 2004](#)). Both dykes and sills can vary in composition from mafic to ultramafic (e.g. Scourie dykes, Scotland: [Hughes et al., 2014](#); Montagnais sills, Labrador Trough, Canada: [Rohon, et al., 1993](#)); however, most regional dyke swarms are found to be mafic.

Traditionally, a sill can be defined as “a tabular igneous intrusion that parallels the planar structure of the surrounding rock” ([Bates and Jackson, 1980](#)). This type of tabular intrusion is in contrast to a dyke, which is traditionally defined as “a tabular igneous intrusion that cuts across the bedding or foliation of the country rock” ([Bates and Jackson, 1980](#)). However, more recent definitions by [Ernst and Buchan \(2004\)](#) do not define sills and dikes based on their relationship to foliation or bedding, but based on their orientation at the time of emplacement (*cf.* [Hall, 1996](#)). As stated by [Ernst and Buchan \(2004\)](#), a dyke may be defined as “a tabular igneous body that was sub-vertical at the time of emplacement,” and in a similar manner, a sill could be defined as an originally sub-horizontal tabular

body. A major determinant in whether the magma is injected as a sill or a dyke is the regional stress pattern (Ernst, 2014). This is because the sheet intrusions will align normal to the direction of minimum compression (Gretener, 1969; Ernst, 2014). An alternative explanation attributes sill formation to variations in a “compensation surface,” where magma pressure equaled lithostatic pressure (Bradley, 1965; Meyboom and Wallace, 1978).

Prior to crustal infiltration, mafic-ultramafic magmas often pond at the base of the crust, where isobaric fractionation of primary picritic magmas can occur. Eventually the ponded magma will completely solidify, forming a dense crustal “underplate” (Cox 1980, 1993; Ernst, 2014). This underplate may be recognized as high-velocity lower crust (HVLC) at the base of the crust, and is often associated with LIPs and volcanic rifted margins. Examples of LIPs with a seismically-recognized underplate component include the Keweenawan (Canada) and Emeishan (China) LIPs. Furthermore, significant volumes of silicic melts may be generated by the melting of fusible crustal rocks at the base of the crust (Ernst, 2014).

Given that the intrusions of the Frobisher suite are all sill-like intrusions, the remainder of the background is focused primarily on the characteristics of sills, and consequently, dyke are only discussed where appropriate or necessary.

2.1.3.2 General characteristics of sills

Individual thicknesses of sills may vary from only a few millimeters to hundreds of meters, and sills may extend up to hundreds of kilometers along

strike (Cawthorn, 2012; Ernst, 2014). Thicker sills may form differentiated intrusions, however, many appear to be the result of repeated injections of magma, where subsequent intrusions are injected before the previous intrusions crystallized significantly (Cawthorn, 2012). The thickest sill is the Windamurra Complex (Australia), which is estimated to be ~13 km thick (Mathison and Ahmat, 1996).

A study of the Karoo sills (South Africa; Figure 2.1) thickness was conducted in order to typify sill thickness. Of the 160 sills intersected in cores drilled through the Karoo Supergroup, 52 were < 10 m thick and as the thickness of the sills increased, the frequency significantly decreased, with only 3 sills exceeding 140 m thick (Cawthorn, 2012). However, it is suggested that the thicker sills may be composite (Galerie et al., 2008; 2011). For example, one 189 m-thick sill recently drilled in the middle of the Karoo appears to have been inflated an additional 102 m by two thinner sills, both with quenched margins (Slement, 2010). Without proper field exposure, it might have been wrongly identified as a single 290 m thick sill (Cawthorn, 2012).

In recent decades, advances in the understanding of sill-province geometry has been drawn from modern seismic imaging, experimental modelling, magnetic fabric studies, and detailed mapping (Chevallier and Woodford, 1999; Davies et al., 2002; Smallwood and Maresh, 2002; Thomson and Hutton, 2004; Planke et al., 2005; Hansen and Cartwright, 2006; Thomson and Schofield et al., 2008; Polteau et al., 2008a; Galland et al., 2009; Miles and Cartwright, 2010; Schofield et al., 2010). Through integrated detailed mapping

and three-dimensional seismic images, the morphology of sills in sedimentary basins has been shown as generally saucer-shaped, characterized by a circular or elliptical shape in plan view with an inner flat dish surrounded by an inward arcuate rim and outer sill ([Figure 2.2](#)). Where pre-existing anisotropies exist (e.g., foliations and fractures), complex stair-stepping geometries can develop. Additionally, individual sills appear to be comprised of a series of lobes (analogous to those observed in lava flows) which branch out from a central source or feeder. Their formation has been linked to a shallow depth of emplacement (low confining pressure) into poorly consolidated sediments ([Schofield, 2012](#)).

A positive correlation (correlation factor of 4-5) has been observed between the diameter of the inner sill, and the emplacement depth (e.g. [Malthe-Sørenssen et al., 2004](#); [Polteau et al., 2008b](#); [Galland et al., 2009](#)). This relationship supports the observation of greater lateral extents of sills emplaced deep in the crust, particularly in the basement. A correlation also exists between the emplacement of saucer-shaped sills and local domal uplift of the overlying crust ([Polteau et al., 2008b](#); [Galland et al., 2009](#)). With regard to channelized flow, thin sill bodies emplaced in cold country rocks will tend to “freeze” in situ due to rapid cooling and crystallization rates ([Barnes et al., 2015](#)). As a result, magma will tend to be focused in thicker sections of the intrusion where complete solidification has not been attained ([Barnes et al., 2015](#)).

2.1.3.3 Sill provinces

Sills that cluster in a restricted geographic region are often related and may comprise part of a sill province. This typically occurs when LIP-scale volumes of mafic ($\pm$ ultramafic) magma are injected into thick sedimentary basins, where it is common for extensive arrays of stacked saucer-shaped and/or tabular sills to be linked by dykes (Hansen and Cartwright, 2006; Polteau et al., 2008; Galland et al., 2009). This geometry of sills has been widely developed within the Siberian LIP (e.g., Yakubchuk and Nikishin, 2004) and Karoo Basin (Figure 2.1; e.g., Galerne, 2011; Svensen et al., 2012). The mechanics of shallow intrusions control these geometries, as well as the interaction between the transport network and Earth's surface and space creation through roof lifting (Galland et al., 2009; Bungler and Cruden, 2011). Sill provinces range in size from tens of square kilometers, where they are associated with local volcanic centres, to thousands of square kilometers (e.g. Karoo sills of southern Africa). Despite the large amounts of magma involved, saucer-shaped sills are empirically highly unfavorable for the formation of major mineral deposits.

2.1.4 Surface magmatism

In many cases, LIP plumbing systems are expressed on surface in the form of continental flood basalts or as point sources (e.g., shield volcanoes), which appear to be coeval with intrusion emplacement in specific provinces (e.g., Karoo; Cawthorn, 2012). Flood basalts are typically fed from fissures, and lavas can flow over surface distances exceeding 300 km. The longest distance estimated for a single lava flow is nearly 1000 km, in the Deccan LIP, where flows

typically vary in thickness between 255 and 550 m (India; [Self et al. 2008](#)). Furthermore, a considerable portion of the LIP's plumbing system may not connect to the surface (i.e., may not supply magma for extrusion on surface).

2.2 Layered Intrusions

There is an important type of igneous intrusion that is distinguished from all other mafic to ultramafic intrusions by the presence of layering, and these are termed *layered intrusions*. These intrusions mimic sedimentary layering, containing layers ranging from millimeters to hundreds of meters in thickness, and are commonly associated with LIP events ([Wilson, 1989](#)). They likely form from a variety of different processes and can take on a variety of shapes, including lopoliths, sills, and dikes. They provide direct samples of crystallization products from the magmas in high-level magma chambers ([Wilson, 1989](#)). Layered intrusions were emplaced during both the Precambrian and Phanerozoic ([Bogatikov et al., 2000](#)), with many significant ones preserved in the latter. This type of intrusion also commonly hosts magmatic ore deposits (e.g. Ni-Cu-PGE, Ti-V, P).

For the purpose of this paper, a *layer* may be defined as a sheet-like cumulate unit, distinctive in its textural and compositional features ([Irvine, 1981](#)). Layers may be described in terms of their texture, composition, thickness, form or internal structure ([Irvine, 1981](#)). *Laminae* refer to thin layers less than 3 cm in thickness, that are considered individual distinctive units ([Irvine, 1981](#)). Igneous laminations, in contrast, are grain-size scale features that are much more subtle

and exist pervasively throughout the cumulate fabric (Irvine, 1981).

Layered intrusions may contain a number of different types of layering, including rhythmic, phase, and modally graded layering, which describe the mode of repetition (Irvine, 1981). Rhythmic layering (also termed cyclic layering) is characterized by “the conspicuous, systematic recurrence of distinctive layers or sequences of layers of the same kind” (Irvine, 1981); for example, the rhythmically layered sequence: dunite, olivine websterite, dunite, olivine websterite, etc. Phase layering may be defined by the appearance and disappearance of a certain mineral (Irvine, 1981), for example, the phase layering sequence: harzburgite, chromite harzburgite, chromite wehrlite. Lastly, modally graded layering is used for layers that are systematically graded stratigraphically, with different mineralogies or modal contents (Irvine, 1981), for example, gabbro, leucogabbro with higher plagioclase modal content, and anorthosite. Another common type of layering is cryptic layering, which describes the variation in cumulus-mineral composition (Irvine, 1981); for example, a dunite with olivine Mg#~70 is stratigraphically higher than a dunite with olivine Mg#~80 and stratigraphically lower than a dunite with olivine Mg#~60.

2.2.1 Geometry of layered intrusions

The geometry of mafic to ultramafic intrusions is controlled by a number of factors, including: the structure of the crust (homogeneous or stratified, and the distribution of fracture and zones of weakness), far-field stresses governed by tectonic setting; near-field stresses in conjunction with buoyancy forces occurring

at propagating tips of magma-filled cracks; existence of pre-existing structures; thermo-mechanical erosion influence on conduit geometry; and the interplay between the combined rates of flow, crystallization, and country rock melting and assimilation (Barnes et al., 2015). Roof uplift and floor depression can account, in part, for space creation in some mineralized mafic-ultramafic intrusions, however, the margins of chonolith intrusions in Norils’k-Talnakh (Russia) truncate layering in host sediments, indicating the intrusions created their own space (Barnes et al., 2015). This appears to be a crucial feature of intrusions hosting ore (Barnes et al., 2015). In addition to the typical tabular morphology of sills, layered intrusions can take on a variety of forms, including chonoliths and dykes.

2.2.1.1 Stratiform (sill-like) intrusions

Stratiform, sill-like, intrusions are laterally continuous over great distances. Examples of these include the Bushveld Complex of South Africa (e.g. Mungall et al., 2016), and the Giles Complex of central Australia (e.g. Maier et al., 2015). The latter has been interpreted to originally represent a single, laterally extensive stratiform intrusion that was tectonically dismembered into numerous distinct bodies (Maier et al., 2015), including approximately twenty major layered intrusions and several generations of sills and dykes, which are compositionally diverse (Ballhaus & Glikson, 1995).

2.2.1.2 Dyke-like intrusions

Dyke-like layered intrusions (also termed “funnel dykes”; e.g. Hall, 1996)

are unusually-thick, vertical sheet-like bodies that exceed the thickness of typical regional diabase dykes. In cross-section, these intrusions exhibit a Y-shaped profile, narrowing with increasing depth to a feeder dyke or “keel” (e.g. [Podmore and Wilson, 1987](#)). In some cases, these dyke-like intrusions are associated with regional dyke swarms (e.g. the Muskox intrusion, which is part of the Mackenzie dyke swarm/LIP; [Baragar et al., 1996](#)). Dyke-like layered intrusions include (as per summary in [Ernst and Buchan 1997a](#)) the 2575 Ma Great Dyke of Zimbabwe of South Africa (550-km-long and up to 11 km wide; [Podmore and Wilson, 1987](#); [Wilson, 1996](#)), the Jimberlana dyke of western Australia (up to 2.5 km wide; part of 2410 Ma Widgiemooltha LIP; [McClay and Campbell, 1976](#)), the Muskox intrusion (up to 0.5 km wide; part of the 1270 Ma Mackenzie LIP; [Irvine, 1980](#); [Francis, 1994](#)), and the ca. 1160 Ma giant Tugtutoq dykes of southern Greenland (0.5 to 0.8 km wide; [Upton and Thomas, 1980](#); [Upton et al., 1996](#)). The Great Dyke of Zimbabwe narrows from ~11 km to under 1 km over a distance of less than 3 km below the surface. In a similar manner, the Muskox layered intrusion narrows from > 11 km to a feeder dyke of ~500 m in thickness.

2.2.1.3 Funnel-shaped intrusions

Funnel-shaped differentiated intrusions are a class of layered intrusion that are approximately equant in shape in plan view, and funnel-, cone- or bowl-shaped in cross-section ([Petrasko et al., 1978](#); [Loney and Himmelberg, 1983](#); [Hall, 1996](#)). Depending on the specific geometry, the terms plug, stock, lopolith, or laccolith may also apply (e.g. [Corry, 1988](#); [Bunger and Cruden, 2011](#)).

Examples of this style of intrusion include the Rum and Skye intrusions of the British Tertiary Province of the North Atlantic Igneous Province (NAIP; e.g. [Emeleus and Gyopari, 1992](#)), the Kiglapait intrusion of the Nain Plutonic suite of eastern Canada (e.g. [Morse, 1979](#)), and the Skaergaard intrusion of the east Greenland portion of the NAIP (e.g. [Wager and Brown, 1968](#); [McBirney, 1996](#)). They have been inferred to be fed by pipe-like feeders located beneath the deepest part of the funnels ([Loney and Himmelberg, 1983](#); [Hall, 1996](#)), or simply represent localized areas of upflow along a dyke ([Ernst and Buchan, 1997b](#)). Funnel-shaped intrusions are sometimes associated with LIP events (e.g. Skaergaard intrusion), while others are associated with subduction settings (Alaskan-type intrusions; e.g. [Loney and Himmelberg, 1983](#); [Eyuboglu et al., 2011](#)). Funnel-shaped intrusions may also represent the roots of volcanic systems ([Ernst, 2014](#)).

2.2.1.4 Chonoliths

Chonoliths are elongate, tube-like intrusions. Examples include the Nebo-Babel intrusion in the Musgrave Province of western Australia ([Seat et al., 2007](#)), Nkomati/Uitkomst of South Africa ([Maier et al., 2004](#)) and the Limoeiro Complex in Brazil ([Mota-e-Silva et al., 2013](#)). Some mineralized intrusions in the Noril'sk-Talnakh camp are also chonoliths ([Czamanske et al., 1995](#); [Sluzhenikin et al., 2014](#)), however, they are considerably wider than they are thick, and are therefore ribbon-shaped. The tube-like host chambers of Eagle (Midcontinent rift, USA), Kalatongke (Xinjiang, China), and Haungshandong (Central Asian

Orogenic Belt; [Lightfoot and Evans-Lamswood, 2015](#)) have dyke-like keels, which have led to the suggestion that these intrusions may have resulted from the lateral flow in flared dykes, or they may represent forms of blade-shaped dykes ([Barnes et al., 2015](#)).

2.2.2 Distribution of layered intrusions in LIPs

It is common for layered intrusions, as well as associated continental flood basalts, to be divided based on their incompatible-element contents, into high-Ti and low-Ti magma types ([Cox et al., 1967](#); [Marsh et al., 2001](#); [Jourdan et al., 2007a](#); [Bryan and Ernst, 2008](#)). Within certain LIPs, there is a marked provinciality in the distribution of the two magma types (e.g. Paraná-Entendeka, Karoo, and Emeishan).

2.2.3 Mechanisms of layer formation in layered intrusions

Given that layering is prevalent in sills across the Frobisher suite sills, descriptions of mechanisms of layer formation is justified.

2.2.3.1 Fractional crystallization

The traditional explanation for the formation of layered igneous intrusions is through the process of fractional crystallization. Gravitational settling is believed to be the physical process involved, in which minerals crystallize and then sink or float to either the base or roof of a cooling magma body ([Wager and Brown, 1967](#)). This process results in successive batches of crystals accumulating and being preserved as layers ([Wager and Brown, 1967](#)). Cumulate is a term that

was created to describe layers formed exclusively via the crystal settling process (Wager and Brown, 1967). The crystals that are interpreted to have settled out a melt are termed ‘cumulus minerals,’ and the crystals derived from cooling of the melt surrounding these crystals were termed ‘intercumulus crystals’ (Wager and Brown, 1967).

However, in the presence of convection, the gravitational settling process is deemed inadequate, as crystals would stay in suspension (Sparks et al., 1984). This reasoning has brought forward an alternative process to layer formation in igneous intrusions termed “convective fractionation.” This process takes place at margins of an intrusion and involves magma adjacent to growing crystals convecting away as it progressively develops a different density compared to the parental magma, during crystallization (Sparks et al., 1984).

There is also a possibility that crystallizing magmas in layered intrusions behave as double diffusive convection systems. In such case, the magma would be stacked as a series of thick convecting layers, separated by thin, static “diffusive” boundary layers (Kruger and Smart, 1987). The intrusions would then crystallize from the base upwards, where chemical changes due to fractional crystallization of the lowermost convecting layer would diffuse across the thin “diffusive” boundary layers that separate the thick, stacked convection layers of magma (Kruger and Smart, 1987). This process arises because of the interaction between two fluid components that have different rates of diffusion (Radko, 2013).

An injection of a slurry of crystals may also sort itself by size, density or

shape, leading to the development of well-formed layers (Marsh, 2013). In this case, the layering present in sills could reflect crystal slurry injection over rhythmic crystallization (Marsh, 2013). Candidates for intrusions derived through this form of emplacement would be those where the sandwich horizon is clearly located above the centre of the intrusion (Marsh, 2013).

Certain rock-types, particularly those of mafic composition, often contain layering (Wager and Brown, 1967). However, numerous massive mafic intrusions have been documented as well (Wager and Brown, 1967). The massive mafic intrusions often have limited thickness (i.e. less than 300 m thick sills), and as a result, likely cooled relatively quickly; reflected by the presence of ophitic and sub-ophitic textures (Wager and Brown, 1967). Coarser-grained mafic units are more likely to experience significant crystal settling and develop layering (Wager and Brown, 1967). Also note that a narrow intrusion may achieve a gabbroic texture if it is emplaced in a hot environment (Wager and Brown, 1967), allowing for a slower than normal rate of cooling.

Amongst these sheets, there must be an absence of chilled margins and cross-cutting relationships, and in most cases, signs of cryptic variation are present (Wager and Brown, 1967). Cryptic variation is simply a systematic change in the nature and composition of the mineral phases within an intrusion and relates to the stratigraphic height in the layering (Wager and Brown, 1967).

2.2.3.2 Composite sills versus vertically-stacked sills

A more recently recognized mechanism for the formation of “cumulate layers” in layered intrusions is the injection of multiple batches of magma either into a previously emplaced, partially crystallized sill (e.g. [Mungall et al., 2016](#)), or underplating previously emplaced, mostly or fully crystallized, sills (e.g. [Menand, 2008](#)).

Composite sills may develop when one or more sills are injected inside one another (e.g., Bushveld Complex; [Cawthorn, 2012](#); [Mungall et al., 2016](#)). Within these composite sills, internal chilled selvages may or may not be present, depending on the degree to which the initial sill(s) cooled and crystallization prior to subsequent sill emplacement ([Cawthorn, 2012](#)). If multiple injections of magma occur over a short time interval, relative to the time it takes a sill of a certain thickness to cool, this magma will inflate the magma chamber and a proper layered intrusion will form ([Cawthorn, 2012](#)).

Within a sill 50 m in thickness, it would only take ~10 years for its temperature to decrease from 1200°C (liquidus temperature) to 1100°C, at which point the sill would be ~50% crystallized (based on modeling from [Wright et al., 1976](#); similar results were obtained by [Zavala et al., 2011](#)). At these temperatures, chilled margins are expected to occur between a second magma injection and the resident magma, and mixing between the two magmas is unlikely to occur (resulting in composite sill formation, rather than a layered intrusion). Furthermore, modelling by [Tegner et al. \(1993\)](#) suggests a temperature difference as little as 50°C is enough to cause sufficient chilling to isolate the first and second magma injections. Additionally, 50% crystallization is

considered to be the “rheological lock-up point” (e.g. [Johnson and Glazner, 2011](#)), and beyond this point, subsequent injections into the cooling intrusion will retain their physical identity ([Patwardham and Marsh, 2011](#)). Therefore, for a newly injected magma to inflate the the center of an intrusion, the temperature of the resident magma had to have been greater 1100°C at the time of the second injection.

The time period required for a sill to completely cool is proportional to the intrusion’s thickness to the power two (squared), so if the intrusion increases in thickness twofold, it will take four times as long to cool ([Cawthorn, 2012](#)). Therefore, if multiple injections occur, continually, before the intrusion becomes ~50% crystallized (<1100°C), the chamber inflation process becomes self-sustaining and a single magma chamber will continue to increase in size and thickness indefinitely (e.g., Mount Ayliff intrusion of the Karoo; [Marsh et al., 2003](#); [Cawthorn and Kruger, 2004](#); [Cawthorn, 2012](#)).

Alternatively, numerous discrete, *vertically-stacked sills* may be emplaced within a sedimentary package (e.g. Karoo dolerites; [Cawthorn, 2012](#)). If sills are emplaced adjacent to one another, internal chilled margins are expected to form between each sill, if the earlier sill was already solid ([Menand, 2008](#)). In the Karoo, the majority of sills occur as discrete entities, however, composite sills with internal chilled salvages are occasionally documented ([Galerie et al., 2008](#)).

A key determinant as to whether a vertically-stacked sill complex or a layered intrusion forms is the time interval during which intrusion occurs, and their respective thicknesses ([Cawthorn, 2012](#)). Assuming a liquidus of 1200°C

(typical for basaltic magma), sills are virtually solid at 950°C, given relatively cool host rocks, and heat-loss through conduction: a 1 m sill will be solid in less than a day, a 10 m thick sill will cool in ~1 year, and a 200 m sill will take ~260 years. However, if a significant volume of crystal-free magma persisted in the core of a thicker sill, it is possible that convection could occur, resulting in more rapid cooling of the sill. However, relatively thin sills are not likely to convect ([Marsh, 1989; 2007](#)). With regard to the physical aspects of multiple sill injection, [Menand \(2008\)](#) demonstrated that subsequent injections are more likely to underplate, rather than overplate the previous intrusion. Additionally, if the initial intrusion has not completely cooled and a subsequent injection of magma truncates its lower chilled margin, it would likely be trapped in the liquid core of the initial sill ([Cawthorn, 2012](#)).

Definitively distinguishing between a single evolving magma chamber and a sheeted sill complex has generally not been possible due to a lack of extensive, high-precision geochronological data ([Mungall et al., 2016](#)). However, modern high-precision U-Pb geochronology, on chemically abraded single magmatic zircon grains, can radically improve overall age accuracy and provide sub-million-year age resolution. For example, [Mungall et al. \(2016\)](#) has demonstrated, through high-precision geochronology, that individual chromite layers of the Rustenburg Layered Suite (of the Bushveld Igneous Complex), presumed to have been formed via crystal settling within a vast magma chamber, formed instead as discrete sill-like intrusions that were emplaced beneath older layers, forming a stack of individual sill-like intrusions.

2.3 Mafic-Ultramafic Intrusion-Hosted Mineral Deposits

Within the geologic record, layered mafic, ultramafic, and mafic-ultramafic intrusions are relatively uncommon ([Zientek, 2012](#)). However, these intrusions are host to magmatic sulphide ore deposits, which occasionally contain economic concentrations of copper (Cu), nickel (Ni), chromium (Cr) and platinum-group-elements (PGE; [Schulz, 2010](#)), and as a result, are exploited for their base and precious metals. Layered intrusion-hosted mineral deposits contain the world's only economic concentrations of PGE ([Zientek, 2012](#)). These metals are of major importance to our economy and everyday living, where copper is used for electrical wiring and plumbing, nickel to make nickel steels and other highly-utilized alloys, and PGEs for their catalytic properties.

There are two major types of magmatic sulphide deposits that occur within layered mafic, ultramafic, and mafic-ultramafic intrusions:

1. Cu- and Ni-rich deposits: These deposits are primarily valued for their Cu and Ni contents, and are characteristically sulphide-rich (10-90% sulphide; [Schulz, 2010](#)).
2. PGE-rich deposits: These deposits are primarily valued for their PGE contents (namely platinum (Pt), Palladium (Pd), and Rhodium (Rh), and are characteristically sulphide-poor (0.5-5% sulphide; [Schulz, 2010](#)). The deposits can be subdivided into two types on the basis of their style of mineralization:
 - a. PGE-reef mineralization (e.g. Bushveld): “PGE reefs” comprise stratabound PGE-enriched lode mineralization ([Zientek, 2012](#)), and are

sulphide-poor (<5% sulphide; [Schulz, 2010](#)). The reef-type mineralization is laterally continuous along strike for the length of the intrusion, which is often tens to hundreds of kilometers ([Zientek, 2012](#)). However, the mineralized interval is thin, varying from centimeter to meter-scale in thickness, compared to the hundred to thousand-meter stratigraphic thickness in a given intrusion ([Zientek, 2012](#)). The term “reef” is used to refer to the rock layer that is mineralized and has a distinctive texture ([Naldrett, 2004](#)), or to the PGE-enriched sulphide mineralization that occurs within a rock layer ([Zientek, 2012](#)). The primary commodities extracted from these PGE-reef deposits are platinum, palladium, and rhodium, with ruthenium, iridium, osmium, copper, nickel and gold as by-products ([Zientek, 2012](#)).

- b. PGE-enriched (contact-type) sulphide mineralization: This type of mineralization occurs near the contacts or margins of layered mafic-ultramafic intrusions ([Iljina and Lee, 2005](#)), and is characterized by zones of disseminated to massive concentrations of iron-copper-nickel-PGE enriched sulphide, tens to hundreds of meters in thickness ([Zientek, 2012](#)). The igneous host rocks often contain modes and textures that irregularly vary on the centimeter to meter scale. Mineralization is concentrated within both the igneous intrusion and its surrounding country rocks, and tends to preferentially localize along the contacts with sulfur-, iron-, or CO₂-bearing units within the host country rock ([Zientek, 2012](#)). The deposits are polymetallic, containing variable proportions of primary

commodities Ni, Cu, and PGE, with gold recovered as a by-product (Zientek, 2012).

2.3.1 Ni-Cu-PGE deposit petrogenesis

2.3.1.1 Geodynamic history and setting

There is a consensus that several major camps (e.g. Duluth Complex deposits of the Midcontinent Rift; Ripley, 2014) are associated with rifted environments where large volumes of mostly mantle-plume-derived mafic or ultramafic magmas were injected into sedimentary basins. Furthermore, a spatial association of Ni-Cu-PGE sulphide deposits, large and small, with ancient Archean cratonic margins, whether the margins are rifted or not, is recognized (Begg et al., 2010). This is believed to be a consequence of the deflection of ascending plume heads towards zones of thinner lithosphere (i.e. cratonic margins; Sleep et al., 2002; Sobolev et al., 2011). Plume deflection (Figure 2.3) may account for linear belts of plume magmatism and ore deposits which must have been derived from a “bullseye” point-source thermal anomaly ascending from the deep mantle (e.g. Circum-Superior LIP; Barnes et al., 2015).

2.3.1.2 Sources and compositions of ore-forming magmas

The metals contained within magmatic mineral deposits ultimately originate from the mantle source from which the transporting magma is derived, which is usually sub-lithospheric and mantle-plume-derived (i.e. deep mantle; Barnes et al., 2015). Mineralized provinces tend to be associated with hotter, high-MgO

magmas, rather than typical ambient mantle melting products (i.e. MORB; [Barnes et al., 2015](#)). Essential to the formation of these deposits is voluminous, high-temperature melts from sub-lithospheric source, making mantle plumes the most likely source for these deposits ([Barnes et al., 2015](#)). Consequently, deposits within any province are often associated with the highest degree partial melts, the most adcumulate rocks (i.e. those with highest ratio/proportion of cumulus mineral to intercumulus trapped liquid), and least fractionated magmas within conduit systems ([Barnes et al., 2015](#)).

2.3.1.3 Sources of sulphur and transport of sulphide liquid

In nearly all Ni-Cu-PGE deposits, the S isotopic evidence strongly supports the role of assimilated crustal sulphur in the formation of the deposits ([Ripley and Li, 2003, 2013](#); [Keays and Lightfoot, 2010](#)). This is particularly important in komatiite-hosted deposits and others associated with high-T magmas, because the S content of their deep mantle sources is far below the sulphide content at sulphide liquid saturation (SCSS; [Barnes et al., 2015](#)). Furthermore, due to the negative correlation of SCSS with pressure, the magma becomes more undersaturated as it ascends ([Mavrogenes and O'Neil, 1999](#)). The best indicator of prospectivity is the presence of proximal S-rich sediments ([Barnes et al., 2015](#)). However, current models for many significant deposits involve initial segregation of immiscible liquid occurring deep within the lower crust, far below the level where the actual deposit forms, followed by subsequent entrainment, upward transport, and mechanical deposition ([Barnes et al., 2015](#)).

Examples of deposits for which this model has been applied include Voisey's Bay, Canada (Lightfoot et al., 2012; Saumur et al., 2013), Noril'sk-Talnakh, Russia (Arndt et al., 2003; Li et al., 2009), Eagle, U.S.A. (Ding et al., 2012b), and Jinchuan, China (Tang, 1991; Song et al., 2012). This model is largely driven by the fact that these deposits host a vast excess of sulphide and chalcophile elements that could not have possibly been dissolved in the volume of magma that is their host (Naldrett, 1992; Lightfoot and Keays, 2005; Song et al., 2012).

2.3.1.4 Mineralized intrusion formation

There are two distinctly different mechanisms proposed for the deposition of cumulus sulphide liquid: 1) in-situ nucleation and growth of sulphide droplets from a silicate melt; and 2) a form of mechanical deposition of transported droplets (Barnes et al., 2015). The former process occurs in-situ and involves the segregation of an immiscible sulphide liquid from a silicate magma (Barnes et al., 2015). The segregation occurs in response to changes in temperature and composition of a host magma that causes the S content to exceed sulphide liquid saturation (see review in Barnes et al., 2015). The precipitated sulphide droplets will either co-precipitate with sub-liquidus silicates resulting in disseminated ores, or the droplets will be transported further along in the plumbing system (Barnes et al., 2015). In the latter case, the mechanical segregation of sulphide liquid is driven by density contrast, and may be enhanced by coalescence of the droplets, mechanical riffling, rapid deposition - favouring changes in the flow regime, or

kinetic sieving in density currents (charges with silicate crystals and sulphide liquid; [Barnes et al., 2015](#)).

An essential characteristic of many mineralized intrusions is the discordant relationship of the igneous contact with host country rocks ([Barnes et al., 2015](#)). This indicates that space was created via assimilation (physical erosion), rather than roof-lifting, as is the case for saucer-shaped sills ([Barnes et al., 2015](#)). If an intrusion is small, a chilled margin will form, preventing heat transfer to wall-rocks. If the intrusion is much larger, and of high-flux, the initial chilled margin will re-melt and hot magma will come into contact with wall rocks, leading to assimilation ([Huppert and Sparks, 1988](#)).

2.4 Metamorphism of Mafic and Ultramafic Rocks

Given that the geology of southern Baffin Island comprises rock units that have been subjected to amphibolite- to granulite-facies metamorphism, an overview of the metamorphism of medium to high grade mafic to ultramafic rocks is warranted. This section is based largely on [Bucher & Grapes \(2011\)](#).

Mafic units metamorphosed to amphibolite and granulite facies conditions will display the characteristic assemblages of plagioclase + hornblende +/- clinopyroxene (or garnet +/- biotite), and plagioclase + clinopyroxene + garnet (or orthopyroxene), respectively. Plagioclase composition ranges from oligoclase (~500°C; lower amphibolite), to andesine (~600°C; middle amphibolite), to labradorite (~650°C; upper amphibolite), however, the anorthite content of the metamorphic plagioclase is influenced also by the bulk composition of the rock.

The appearance of clinopyroxene or garnet marks the beginning of upper amphibolite facies (~650°C).

The transition from hydrous amphibolite to anhydrous granulite occurs over a 200°C interval (650-850°C). The appearance of orthopyroxene in clinopyroxene-bearing, quartz-free rocks is clear evidence that granulite facies conditions were attained, even if hornblende is still present. Mafic granulites typically contain either the assemblage plagioclase + clinopyroxene + garnet (at high-pressure), or plagioclase + clinopyroxene + orthopyroxene (at low-pressure). As calcic hornblende breaks down, the dehydration reactions add Di- and An- components to clinopyroxene and plagioclase, respectively. This results in abundant labradorite within granulite mafic rocks. High-grade brown hornblende is very stable and can persist to temperatures >950°C in metabasic rocks. As a result many mafic granulites still contain hornblende. Hercynitic spinel is a common accessory in mafic granulites, diagnostic of low-P granulite conditions (<400 MPa).

Ultramafic units metamorphosed to amphibolite and granulite facies, will contain the diagnostic assemblages enstatite + forsterite + hornblende + spinel, and enstatite + forsterite + diopside + spinel, respectively. At lower temperatures (<700°C), ultramafic rocks contain chlorite as the aluminous phase. At high temperatures (>700°C), ultramafic rocks also contain one of three different aluminous mineral phases: anorthite at low pressure, spinel at medium to high pressure, and garnet at ultra high pressure (>~1 GPa; eclogite facies). If subjected to retrograde conditions, the aluminous phase may back react to form

chlorite, and clinopyroxene will form amphibole.

References

- Arndt, N.T., Czamanske, G.K., Walker, R.J., Chauvel, C., Fedorenko, V.A. (2003). Geochemistry and origin of the intrusive hosts of the Noril'sk–Talnakh Cu–Ni–PGE sulfide deposits. *Economic Geology Bulletin, Society of Economic Geologists* 98, 495–515.
- Ballhaus, C. and Glikson, A.Y. (1995). The petrology of layered mafic– ultramafic intrusions in the Giles Complex, western Musgrave block, central Australia: *AGSO Journal*, v. 16, p. 69–89.
- Baragar, W.R.A., Ernst, R.E., Hulbert, L. & Peterson, T. (1996). Longitudinal petrochemical variation in the Mackenzie dyke swarm, northwestern Canadian Shield. *Journal of Petrology*, 37: 317–359.
- Barnes, S.J., Cruden, S., Arndt, N., Saumur, B. (2015). The mineral system approach applied to magmatic Ni-Cu-PGE sulphide deposits, *Ore Geology Review* 71, 673-702.
- Barnes, S-J., Lightfoot, P.C. (2005). Formation of magmatic nickel sulfide deposits and processes affecting their copper and platinum group element contents. *Economic Geology 100th Anniversary Volume*, pp. 179–214.
- Bates, R.L. and Jackson, J.A. (ed.). (1980). *Glossary of Geology*. American Geological Institute, Falls Church, Virginia.
- Begg G.C., Griffin W.L., Natapov L.M., O'Reilly S.Y., Grand S.P., O'Neill C.J., Hronsky J.M.A., Poudjom Djomani Y., Swain, C.J., Deen T. & Bowden P. (2009). The lithospheric architecture of Africa: Seismic tomography, mantle petrology and tectonic evolution, *Geosphere*, 5, 23-50. (2009). The

-
- lithospheric architecture of Africa: seismic tomography, mantle petrology, and tectonic evolution. *Geosphere*, 5: 23-50.
- Begg, G.C., Hronsky, J.A.M., Arndt, N.T., Griffin, W.L., O'Reilly, S.Y., & Hayward, N. (2010). Lithospheric, Cratonic, and Geodynamic Setting of Ni-Cu-PGE Sulfide Deposits, *Economic Geology*, 105, 1057–1070.
- Bogatikov, O.A., Kovalenko V.I., Sharkov, E.V., & Yarmolyuk, V.V. (2000). *Magmatism and Geodynamics. Terrestrial Magmatism Throughout the Earth's History*. Gordon and Breach Science Publication, 511.
- Bradley, J. (1965). Intrusion of major dolerite sills. *Transactions of the Royal Society of New Zealand*, 3 (4): 27-55.
- Bryan, S., & Ernst, R.E. (2008). Revised definition of Large Igneous Provinces (LIPs). *Earth-Science Reviews*, 86: 175–202.
- Bunger, A.P., Cruden, A.R. (2011). Modeling the growth of laccoliths and large mafic sills; role of magma body forces. *Journal of Geophysical Research*, 116.
- Burke K, & Torsvik TH. (2004). Derivation of Large Igneous Provinces of the past 200 million years from long-term heterogeneities in the deep mantle. *Earth Planetary Science Letter*, 227:531–38.
- Campbell, I.H. (2001). Identification of ancient mantle plumes. In R.E. Ernst , K.L. Buchan (eds.), *Mantle Plumes: Their Identification through Time*. Boulder, CO: Geological Society of America, Special Paper 352, pp. 5–21.
- Campbell, I.H. (2005). Large igneous provinces and the mantle plume hypothesis. *Elements*, 1: 265–269.

-
- Campbell, I.H. (2007). Testing the plume theory. *Chemical Geology*, 241: 153–176.
- Campbell, I.H., Griffiths, R.W., Hill, R.I. (1989). Melting in an Archaean mantle plume: heads it's basalts, tails it's komatiites. *Nature* 339, 697–699.
- Cartwright, J. and Hansen, D.M. (2006). Magma transport through the crust via interconnected sill complexes: *Geology*, v. 34, p. 929–932, doi:10.1130/G22758A.1.
- Cawthorn, R.G. (2012). Multiple sills or a layered intrusion? Time to decide. *South African Journal of Geology*, 115: 283–290.
- Cawthorn, R.G. & Kruger F.J. (2004). Petrology and Ni-Cu-PGE potential of the Insizwa lobe, Mount Ayliff Intrusion, South Africa. *Canadian Mineralogist*, 42, 303-324.
- Chadwick, J., Perfit, M., Ridley, I., Jonasson, I., Kamenov, G., Chadwick, W., Embley, R., le Roux, P., and Smith, M. (2005). Magmatic effects of the Cobb hot spot on the Juan de Fuca Ridge. *Journal of Geophysical Research*, 110:B03101, doi:10.1029/2003JB002767
- Chevallier, L. and Woodford, A. (1999). Morpho-tectonics and mechanism of emplacement of the dolerite rings and sills of the western Karoo, South Africa. *Transactions of the Geological Society of South Africa*, 102, 43-54
- Coffin, M.F., Eldholm, O. (1992). Volcanism and continental break-up: a global compilation of large igneous provinces. Geological Society, London, Special Publications 68, 17-30.
- Corry, C.E., (1988). Laccolith: Mechanics of emplacement and growth, *Spec.*

-
- Pap., Geological Society of America, 220.
- Cox, K.G., Macdonal, R., Hornung, G. (1967). Geochemical and petrographic provinces in Karoo basalts of southern Africa. *American Mineralogist*, 52, 1451–1474.
- Cox, K.G. (1980). A model for flood basalt vulcanism. *Journal of Petrology*, 21: 629–650.
- Cox, K.G. (1993). Continental Magmatic Underplating. *Philosophical Transactions of the Royal Society, A* 1993 342 155-166; DOI: 10.1098/rsta.1993.0011.
- Czamanske, G.K., Zen'ko, K.E., Fedorenko, V., Calk, L.C., Budahn, J.R., Bullock, J.H.J., Fries, T.L., King, B.S., Siems, D.F. (1995). Petrographic and geochemical characterization of ore-bearing intrusions of the Noril'sk Type, Siberia; with discussion of their origin. *Resource Geology Special Issue* 18, 1–48.
- Davies, J.H., Bunge, H-P. (2006). Are splash plumes the origin of minor hotspots? *Geology* 34:349–52.
- Davies, R., Bell, B.R., Cartwright, J.A., and Shoulders, S. (2002), Threedimensional seismic imaging of Paleogene dike-fed submarine volcanoes from the Northeast Atlantic margin, *Geology*, 30, 223–226.
- Ding, X., Ripley, E.M., Shirey, S.B., Li, C. (2012). Os, Nd, O and S isotope constraints on country rock contamination in the conduit-related Eagle Cu–Ni–(PGE) deposit, Midcontinent Rift System, upper Michigan. *Geochimica et Cosmochimica Acta*, 89, 10–30.

-
- Duggen, S., Hoernle, K.A., Hauff, F., *et al.* (2009). Flow of Canary mantle plume material through a subcontinental lithospheric corridor beneath Africa to the Mediterranean. *Geology*, 37: 283–286.
- Ebinger, C.J., and Sleep, N.H. (1998). Cenozoic magmatism throughout East Africa resulting from impact of a single plume. *Nature*, 395: 788–791.
- Elkins-Tanton, L.T. (2005). Continental magmatism caused by lithospheric delamination. In: Foulger GR, Natland JH, Presnall DC, Anderson DL (eds) *Plates, Plumes and Paradigms*. Geological Society of America Special Paper 388, pp 449-462.
- Emeleus, C.H. and Gyopari, M.C. (1992), *British Tertiary Volcanic Province*, Geological Conservation Review Series, No. 4, Chapman and Hall, London, 259 pages, illustrations, A4 hardback, ISBN 0 41247 980 7.
- Ernst, R.E. (2014). *Large Igneous Provinces*: Cambridge, UK, Cambridge University Press, 653 p.
- Ernst, R.E. and Buchan, K.L. (1997a). Giant Radiating Dyke Swarms: Their use in identifying pre-Mesozoic large igneous provinces and mantle plumes, *In* Mahoney, J. & Coffin, M. (eds.) Large Igneous Provinces: Continental, Oceanic, and Planetary Volcanism, AGU Geophysical Monograph Series Volume 100, pp. 297-333.
- Ernst, R.E. and Buchan, K.L. (1997b). Layered mafic intrusions: a model for their feeder systems and relationship with giant dyke swarms and mantle plume centres: *South African Journal of Geology*, v. 100, p. 319-334

- Ernst, R.E. and Buchan, K.L. (2004). Igneous rock associations in Canada 3. Large Igneous Provinces (LIPs) in Canada and adjacent regions: 3 Ga to present: *Geoscience Canada*, v. 31, p. 103-126.
- Ernst, R.E., Buchan, K.L., Bleeker, W. (2008). The Plumbing System of Large Igneous Provinces (LIPs), Paper presented at the 33rd International Geological Congress (33IGC), Oslo, Norway, 6-14 August, 2008.
- Eyuboglu, Y., Santosh, M., Bektas, O., Chung, S.-L. (2011). Late Triassic subduction-related ultramafic–mafic magmatism in the Amasya region (eastern Pontides, N. Turkey): implications for the ophiolite conundrum in Eastern Mediterranean. *Journal of Asian Earth Sciences*, 42: 234–257.
- Farnetani, C.G., and Samuel, H. (2005). Beyond the thermal plume paradigm. *Geophysical Research Letters*, 32: L07311.
- Francis, D. (1994). Chemical interaction between picritic magmas and upper crust along the margins of the Muskox intrusion, Northwest Territories. *Geological Survey of Canada Paper*, 92-12, 94pp.
- Galerne, C.Y., Galland, O., Neumann, E.-R. and Planke, S., (2011). 3-D relationships between sills and their feeders: evidence from the Golden Valley Sill Complex (Karoo basin) and experimental modelling. *Journal of Volcanology and Geothermal Research*, 202, 189-199.
- Galerne, C., Neumann, E.-R. and Planke, S. (2008). Emplacement mechanisms of sill complexes: information from the geochemical architecture of the Golden Valley sill complex. *Journal of Volcanology and Geothermal Research*, 177, 425-440.

- Galland, O., Planke, S., Neumann, E.-R., and Maltte-Sørenssen, A. (2009). Experimental modelling of shallow magma emplacement: application to saucer-shaped intrusions. *Earth and Planetary Science Letters*, 277: 373–383.
- Gretener, P.E. (1969). On the mechanics of the intrusion of sills. *Journal of Earth*, 6(6):1415-1419. doi: 10.1139/e69-143.
- Griffiths, R.W., Campbell, I.H. (1990). Stirling and structure in mantle starting plumes. *Earth Planetary Science Letters*, 99, 66–78.
- Hall, A. (1996). *Igneous Petrology*, 2nd edition. Harlow: Longman Science & Technology.
- Hansen, D.M. and Cartwright, J. (2006). Saucer-shaped sill with lobate morphology revealed by 3D seismic data: Implications for resolving a shallow level sill emplacement mechanism. *Journal of the Geological Society, London*. 163, 509–523.
- Hofmann, A.W. and White, W.M. (1982). Mantle plumes from ancient oceanic crust. *Earth Planetary Science Letters*, 57:421–36
- Hughes, H.S.R., MacDonald, I., Goodenough, K.M., Ciborowski, T.J., Kerr, A.C., Davies, J.H.F.L. and Shelby D. (2014). Enriched lithospheric mantle keel below the Scottish margin of the North Atlantic Craton: evidence from the Palaeoproterozoic Scourie Dyke Swarm and mantle xenoliths. *Precambrian Research*, vol 250, p.97-126.
- Huppert, H.E., and Sparks, R.S.J. (1988). The generation of granitic magmas by intrusion of basalt into continental crust. *Journal of Petrology*, 29, 599–

624.

- Iljina, M.J., and Lee, C.A. (2005). Chapter 4: PGE deposits in the marginal series of layered intrusions, in Mungall, J.E., ed., 2005, Exploration for platinum-group element deposits: Ottawa, Mineralogical Association of Canada Short Course Series Volume 35, p. 75–96.
- Irvine T.N. (1980). Magmatic infiltration metasomatism, double-diffusive fractional crystallization, and adcumulus growth in the Muskox intrusion and other layered intrusions. In: R.B. Hargreaves (Ed.) Physics of magmatic processes. Princeton University Press. Pp. 325-383.
- Irvine, T.N. (1981). Terminology for Layered Intrusion. Geophysical Laboratory, Carnegie Institution of Washington, Washington, D.C. 20008.
- Johnson, B.R. and Glazner, A.F. (2011). Formation of K-feldspar megacrysts in granodioritic plutons by thermal recycling and late-stage textural coarsening. *Contributions to Mineralogy and Petrology*, 159, 599-619.
- Jourdan, F., Bertrand, H., Sharer, U., Blitchert-Toft, J., Féraud, G., and Kampunzu, A.B. (2007). Major-trace element and Sr–Nd–Hf–Pb isotope compositions of the Karoo large igneous province in Botswana–Zimbabwe. *Journal of Petrology*, 48: 1043–1077.
- Kavanagh, J.L., Menand, T., Sparks, R.S.J. (2006). An experimental investigation of sill formation and propagation in layered elastic media. *Earth Planetary Science Letters*, 245, 799–813. <http://dx.doi.org/10.1016/j.epsl.2006.03.025>.
- Keays, R.R., Lightfoot, P.C. (2010). Crustal sulfur is required to form magmatic

-
- Ni–Cu sulfide deposits; evidence from chalcophile element signatures of Siberian and Deccan Trap basalts. *Mineralium Deposita*, 45, 241.
- King S.D. and Anderson D.L. (1995). An alternative mechanism of flood basalt formation. *Earth and Planetary Science Letters* 136: 269-279
- King S.D. and Anderson D.L. (1998). Edge-driven convection. *Earth and Planetary Science Letters* 160: 289-296.
- Kruger, F. J. and Smart, R. (1987). Diffusion of trace elements during bottom crystallisation of double-diffusive convection systems: The magnetite layers of the Bushveld Complex. *Journal of Volcanology, Geothermal Research*, 34, 133-42.
- Leat, P.T. (2008). On the long-distance transport of Ferrar magmas, in Thomson, K., and Petford, N., eds., *Structure and emplacement of high-level magmatic systems: Geological Society of London Special Publication 302*, p. 45–61, doi:10.1144/SP302.4.
- Li, C., Ripley, E.M., Naldrett, A.J. (2009). A new genetic model for the giant Ni–Cu–PGE sulfide deposits associated with the Siberian flood basalts. *Economic Geology Bulletin, Society of Economic Geologists*, 104, 291–301.
- Lightfoot, P.C., and Keays, R.R. (2005). Siderophile and Chalcophile Metal Variations in Flood Basalts from the Siberian Trap, Noril'sk Region: implications for the Origin of the Ni–Cu–PGE Sulfide Ores. *Economic Geology*, 100, 439–462.
- Lightfoot, P.C., Keays, R.R., Evans-Lamswood, D., Wheeler, R. (2012). S-

- saturation history of Nain plutonic suite mafic intrusions; origin of the Voisey's Bay Ni–Cu–Co sulfide deposit, Labrador, Canada. *Mineralium Deposita*, 47, 23–50.
- Liikane, D.A., St-Onge, M.R., Kjarsgaard, B.A., Rayner, N.M., Ernst, R.E. and Kastek, N. (2015). Frobisher suite mafic, ultramafic and layered mafic-ultramafic sills, southern Baffin Island, Nunavut; in *Summary of Activities 2015*, Canada-Nunavut Geoscience Office, p. 21–32.
- Lister, J.R., and Kerr, R.C. (1991). Fluid-mechanical models of crack propagation and their application to magma transport in dykes. *Journal of Geophysical Research*, 96-B6, 10049–10077. <http://dx.doi.org/10.1029/91JB00600>.
- Loney, R.A., and Himmelberg, G.R. (1983). Structure and petrology of the La Perouse gabbro intrusion, Fairweather Range, southeastern Alaska. *Journal of Petrology*, 24: 377–423.
- Lustrino, M. (2005). How the delamination and detachment of lower crust can influence basaltic magmatism. *Earth-Science Reviews*, 72: 21–38.
- Magee, C., Muirhead, J.D., Karvelas, A., Holford, S.P., Jackson, C.A.L., Bastow, I.D., Schofield, N., Stevenson, C.T.E., McLean, C., McCarthy, W., and Shtukert, O. (2016). Lateral magma flow in mafic sill complexes: *Geosphere*, v. 12, no. 3, p. 1–33, doi:10.1130/GES01256.1.
- Maier, W., Rasmussen, B., Fletcher, I., Godel, B., Barnes, S.J., Fisher, L., and Yang, S. (2015). Petrogenesis of the ~2.77 Ga Monts de Cristal complex, Gabon: Evidence for direct precipitation of Pt-arsenides from basaltic magma. *Journal of Petrology*, 56, 1285–1308.

-
- Malthe-Sørensen, A., Planke, S., Svensen, H., and Jamtveit, B. (2004), Formation of saucer shaped sills, in *Physical Geology of High Level Magmatic Systems*, Geol. Soc. London, Spec. Publ., vol. 234, edited by C. Breitkreuz and N. Petford, pp. 215–227, Geological Society, London.
- Mandler, H.A.F. and Clowes, R. M. (1997). Evidence for extensive tabular intrusions in the Precambrian shield of western Canada: A 160-km-long sequence of bright reflections. *Geology*, 25: 271-274.
- Marsh, B.D., (1989). On convection style and vigour in sheet-like magma chambers. *Journal of Petrology*, 30, 479-530.
- Marsh, B.D., (2007). Magmatism, magma and magma chambers. In: A.B. Watts (Editor), *Treatise on Geophysics*, Elsevier, Amsterdam, The Netherlands, 276-333.
- Marsh, B.D., (2013). On some fundamentals of igneous petrology, *Contributions to Mineralogy and Petrology*, 166: 665–690
- Marsh, J.S., Allen, P. and Fenner, N. (2003). The geochemical structure of the Insizwa lobe of the Mount Ayliff Complex with implications for the emplacement and evolution of the complex and its Ni-sulphide potential. *South African Journal of Geology*, 106, 409-428.
- Marsh, J.S., Ewart, A., Milner, S.C., Duncan, A.R., & Miller, R. McG. (2001). The Etendeka igneous province: magma types and their stratigraphic distribution with implications for the evolution of the Paraná–Etendeka flood basalt province. *Bulletin of Volcanology*, 62: 464–486.
- Mathison, C.J. and Ahmet, A.L. (1996). The Windamurra Complex, Western

-
- Australia. In: R.G. Cawthorn (Editor), *Layered Intrusions*, Elsevier, The Netherlands, 485-510.
- Mavrogenes, J.A., O'Neill, H.S.C. (1999). The relative effects of pressure, temperature and oxygen fugacity on the solubility of sulfide in mafic magmas. *Geochimica et Cosmochimica Acta*, 63, 1173–1180.
- McBirney, A.R. (1996). The Skaergaard Intrusion. In R.G. Cawthorn (ed.), *Layered Intrusions*. Amsterdam: Elsevier, pp. 147–180
- McClay, K.R. & Campbell, I.H. (1976). The structure and shape of the Jimberlana Intrusion, Western Australia, as indicated by an investigation of the Bronzite Complex. *Geological Magazine*, 113: 129–139
- McKenzie, D.P. & O'Nions, R.K. (1983). Mantle reservoirs and ocean island basalts. *Nature* 301:229–31.
- Menand, T., 2008. The mechanics and dynamics of sills in layered elastic rocks and their implications for the growth of laccoliths and other igneous complexes. *Earth Planet.ary Science Letters*, 267, 93–99. <http://dx.doi.org/10.1016/j.epsl.2007.11.043>
- Meyer, R., Nicoll, G.R., Hertogen, J., Troll, V.R., Ellam, R.M. and Emeleus, C.H. (2009). Trace element and isotope constraints on crustal anatexis by upwelling mantle melts in the North Atlantic igneous province: an example from the Isle of Rum, NW Scotland. *Geological Magazine*, 146: 382–399.
- Meyboom, A.F. and Wallace, R.C. (1978). Occurrence and origin of ring-shaped dolerite outcrops in the eastern Cape Province and western Transkei. *Transactions of the Geological Society of South Africa*, 81, 95-99.

- Miles, A., and Cartwright, J. (2010). Hybrid flow sills: a new mode of igneous sheet intrusion. *Geology*, 38: 343–346.
- Montelli, R., Nolet, G., Dahlen, F.A., Masters, G., Engdahl, R., Hung, S-H. (2004). Finite-frequency tomography reveals a variety of plumes in the mantle. *Science* 303:338–43.
- Morgan, W.J. (1971). Convection plumes in the lower mantle. *Nature*, 230: 42–43.
- Morse, S.A. (1979) Kiglapait geochemistry II: Petrography. *Journal of Petrology* 20:591-624.
- Mungall, J.E. (2016). U-Pb geochronology documents out-of-sequence emplacement of ultramafic layers in the Bushveld Igneous Complex of South Africa. *Nature Communication* 7:13385 doi:10.1038/ncomms13385.
- Naldrett, A.J. (1992). A model for the Ni–Cu–PGE ores of the Noril'sk region and its application to other areas of flood basalt. *Economic Geology*, 87, 1945–1962.
- Naldrett, A.J. (1999). World Class Ni-Cu-PGE Deposits: Key Factors in their Genesis: in *Mineralium Deposita*, 34, 227-240.
- Naldrett, A.J. (2004). *Magmatic sulfide deposits—Geology, geochemistry, and exploration*: Berlin, Springer-Verlag, 727 p.
- Natural Resources Canada (2016). GEM: Geo-mapping for Energy and Minerals, Government of Canada, accessed January 2017, < <https://www.nrcan.gc.ca/earth-sciences/resources/federal-programs/geomapping-energy-minerals/18215>>

- Neuendorf, K.K.E., Mehl, J.P. Jr., and Jackson, J.A. (2005). *Glossary of Geology*. American Geological Institute, Alexandria.
- Oyarzun, R., Doblas, M., López-Ruiz, J., and Cebra, J.M. (1997). Opening of the central Atlantic and asymmetric mantle upwelling phenomena: implications for long-lived magmatism in western North Africa and Europe. *Geology*, 25: 727–730.
- Patwardharn, K. and Marsh, B.D. (2011). Dynamics of the development of the Isle au Haut gabbro-diorite layered complex: quantitative implications for mafic-silicic magma interactions. *Journal of Petrology*, 52, 2365-2395.
- Petraskes, A.K., Hodge, D.S., and Shaw, R. (1978). Mechanics of emplacement of basic intrusions. *Tectonophysics*, 46: 41–63.
- Planke, S., Rasmussen, T., Rey, S. and Myklebust, R. (2005), Seismic characteristics and distribution of volcanic intrusions and hydrothermal vent complexes in the Vøring and Møre basins, in *Petroleum Geology: North-West Europe and Global Perspectives—Proceedings of the 6th Petroleum Geology Conference*, edited by A. G. Doré and B. A. Vining, pp. 833–844, Geological Society, London.
- Podmore, F. and Wilson, A.H. (1987). A reappraisal of the structure and emplacement of the Great Dyke, Zimbabwe. *Geological Association Canada Special Paper 33*, pp. 317–330.
- Polteau, S., Ferré, E.C., Planke, S., Neumann, E.-R. and Chevallier, L. (2008a). How are saucer-shaped sills emplaced? Constraints from the Golden Valley Sill, South Africa. *Journal of Geophysical Research*, 113: B12104.

- Polteau, S., Mazzini, A., Galland, O., Planke, S. and Malthe-Sorensen, A. (2008b). Saucer-shaped intrusions: occurrences, emplacement and implications. *Earth and Planetary Science Letters*, 266: 195–204
- Radko, T. (2013). *Double Diffusive Convection*. Cambridge University Press, Cambridge, UK.
- Richards, M.A., Duncan, R.A., & Courtillot, V.E. (1989). Flood basalts and hot-spot tracks – plume heads and tails. *Science*, 246: 103–107.
- Ripley, E.M., 2014. Ni–Cu–PGE mineralization in the Partridge River, South Kawishiwi, and Eagle Intrusions; a review of contrasting styles of sulfide-rich occurrences in the Midcontinent Rift System. *Economic Geology Bulletin, Society of Economic Geology*, 109, 309–324. <http://dx.doi.org/10.2113/econgeo.109.2.309>
- Ripley, E.M., Li, C.S. (2003). Sulfur isotope exchange and metal enrichment in the formation of magmatic Cu–Ni–(PGE) deposits. *Economic Geology Bulletin, Society of Economic Geology*, 98, 635–641.
- Ripley, E.M., Li, C.S. (2013). Sulfide saturation in mafic magmas; is external sulfur required for magmatic Ni–Cu–(PGE) ore genesis? *Economic Geology Bulletin, Society of Economic Geology*, 108, 45–58. <http://dx.doi.org/10.2113/econgeo.108.1.45>.
- Rohon, M.-L., Vialette, Y., Clark, T., Roger, G., Ohnenstetter, and Vidal, P.H. (1993). Aphebian mafic-ultramafic magmatism in the Labrador Trough (New Quebec): its age and the nature of its mantle source; *Canadian Journal of Earth Sciences*, v. 30, p. 1582-1593.

- Saumur, B.M., Cruden, A.R., Boutelier, D., Evans-Lamswood, D.M. (2013). Structure and dynamics of magmatic Ni–Cu sulphide deposits: theory, analogue modelling and insights from Voisey's Bay (Labrador, Canada). 13th biennial SGA meeting. SGA, Uppsala, Sweden, pp. 1054–1057.
- Saunders, A.D. (2005). Large igneous provinces: origin and environmental consequences. *Elements*, 1: 259–263.
- Schofield, N., Stevenson, C., & Reston, T. (2010). Magma fingers and host rock fluidization in the emplacement of sills. *Geology*, 38: 63–66.
- Schulz, K.J., Chandler, V.W., Nicholson, S.W., Piatak, Nadine, Seall, II, R.R., Woodruff, L.G., and Zientek, M.L. (2010). Magmatic sulfide-rich nickel-copper deposits related to picrite and (or) tholeiitic basalt dike-sill complexes—A preliminary deposit model: U.S. Geological Survey Open-File Report 2010–1179, 25 p. (available at <http://pubs.usgs.gov/of/2010/1179/>).
- Sleep, N.H. (1996). Lateral flow of hot plume material ponded at sublithospheric depths: *Journal of Geophysical Research*, v. 101, p. 28,065–28,083.
- Slement, C. (2010). The extent of fractionation and mineral accumulation in a 190 m thick Karoo sill, near the Gariep Dam. Unpublished Honours project. University of the Witwatersrand, South Africa, 41p.
- Sluzhenikin, S.F., Krivolutskaya, N.A., Rad'ko, V., Malitch, K.N., Distler, V.V., Fedorenko, V.A. (2014). Ultramafic–Mafic Intrusions, Volcanic Rocks and PGE–Cu–Ni Sulfide Deposits of the Noril'sk Province, Polar Siberia, Field Trip Guidebook. Institute of Geology of Ore Deposits, Petrography,

Mineralogy and Geochemistry, Yekaterinburg, Russia.

- Smallwood, J. R., and J. Maresh (2002). The properties, morphology and distribution of igneous sills: Modeling, borehole data and 3D seismic from the Faroe-Shetland area, in *The North Atlantic Igneous Province: Stratigraphy, Tectonic, Volcanic and Magmatic Processes*, Geological Society of London Special Publication, vol. 197, edited by D. Jolley and B. R. Bell, pp. 271–306, Geological Society, London.
- Sobolev, S.V., Sobolev, A.V., Kuzmin, D.R., Krivolutsкая, N.A., Petrunin, A.G., Arndt, N.T., Radko, V.A. & Vasiliev, Y.R. (2011). Linking mantle plumes, large igneous provinces and environmental catastrophes. *Nature*, 477: 312–316.
- Song, X.Y., Danyushevsky, L.V., Keays, R.R., Chen, L.-M., Wang, Y.-S. (2012). Structural, lithological, and geochemical constraints on the dynamic magma plumbing system of the Jinchuan Ni–Cu sulfide deposit, NW China. *Mineral. Deposita* 47, 277–297.
- Sparks, R.S.J., Huppert, H.E., & Turner, J.S. (1984). The fluid dynamics of evolving magma chambers. *Philosophical Transactions of the Royal Society of London*. A310, 511-34.
- St-Onge, M.R., Rayner, N.M., Liikane, D. and Chadwick, T. (2015a): Mafic, ultramafic and layered mafic-ultramafic sills, Meta Incognita Peninsula, southern Baffin Island, Nunavut; in *Summary of Activities 2014*, Canada-Nunavut Geoscience Office, p. 11–16.
- Svensen, H., Corfu, F., Polteau, S., Hammer, Ø., and Planke, S. (2012). Rapid

-
- magma emplacement in the Karoo Large Igneous Province: *Earth and Planetary Science Letters*, v. 325-326, p. 1–9, doi:10.1016/j.epsl.2012.01.015.
- Tang, Z. (1991). The metallogenetic model of the Jinchuan platinum-bearing copper–nickel sulphide deposit. *International Symposium on Sulfide Deposits, Academic Papers on Special Subjects, Jinchang, Gansu, China*, pp. 17–43.
- Tegner, C., Wilson, J.R. and Brooks, C.K. (1993). Intraplutonic quench zones in the Kap Edvard Holm layered gabbro complex, East Greenland. *Journal of Petrology*, 34, 681-710.
- Thomson, K., and D. Hutton (2004). Geometry and growth of sill complexes: Insights using 3D seismic from the North Rockall, *Bulletin of Volcanology*, 66, 364–375.
- Thomson, K. and Schofield, N. (2008). Lithological and structural controls on the emplacement and morphology of sills in sedimentary basins. In: Thomson, K. and Petford, N. (eds) *Structure and emplacement of high-level magmatic systems*. Geological Society, London, Special Publications, 302, 31–44.
- Upton, B.G.J., Parsons, I., Emeleus, C.H., & Hodson, M.E. (1996). Layered alkaline igneous rocks of the Gardar Province, South Greenland. In R.G. Cawthorn (ed.), *Layered Intrusions*. Amsterdam: Elsevier, pp. 331–363
- Upton, B.G.J. & Thomas, J.E. (1980). The Tugtutôq Younger Giant Dyke Complex, South Greenland: fractional crystallisation of transitional olivine

-
- basalt magma. *Journal of Petrology*, 21: 167–198.
- van Wijk, J.W., Huisman, R.S., Ter Voorde, M., and Cloetingh, S.A.P.L., (2001). Melt generation at volcanic continental margins: no need for a mantle plume? *Geophysical Research Letters*, 28: 3995–3998.
- Wager, L.R. and Brown, G.M. (1967). *Layered igneous rocks*. xv + 588 p., 278 figs., and 13 pls. (half-tone and coloured). Oliver and Boyd, Ltd., Edinburgh.
- Welford, J.K. and Clowes, R.M. (2004). Deep 3-D seismic reflection imaging of Precambrian sills in southwestern Alberta, Canada, *Tectonophysics*, 388, 161-172.
- Weller, O.M., Dyck, B.J., St-Onge, M.R., Rayner, N.M. and Tschirhart, V. (2015). Completing the bedrock mapping of southern Baffin Island, Nunavut: plutonic suites and regional stratigraphy; in *Summary of Activities 2015*, Canada-Nunavut Geoscience Office, p. 33–48.
- White, R.S. & McKenzie, D. (1989). Magmatism at rift zones – the generation of volcanic continental margins and flood basalts. *Journal of Geophysical Research: Solid Earth and Planets*, 94: 7685–7729.
- White, W.M. (2010). Oceanic island basalts and mantle plumes: the geochemical perspective. *Annual Review of Earth and Planetary Sciences*, 38: 133–160.
- Wilson, A.H. (1996). The Great Dyke of Zimbabwe. In R.G. Cawthorn (ed.), *Layered Intrusions*. Amsterdam: Elsevier, pp. 365–402.
- Wilson, M. (1989). *Igneous Petrogenesis: A Global Tectonic Approach*. Unwin

Hyman, London. <http://dx.doi.org/10.1007/978-1-4020-6788-4>

Wright, T.L., Peck, D.L. and Shaw, H.R. (1976). Kilauea lava lakes: natural laboratories for study of cooling, crystallization and differentiation of basaltic magma. In: G.H. Sutton, M.H. Manghani, R. Moberly and E.U. McAfee (Editors), *The Geophysics of the Pacific Ocean basin and its margin*, Geophysical Monograph Series, American Geophysical Union, D.C., 19, 375-390.

Yakubchuk, A., and Nikishin, A. (2004). Noril'sk-Talnakh Cu-Ni-PGE deposits: a review of the classic model. *Mineralium Deposita* v39, p 125-142

Zavala K., Leitch A.M. and Fischer G.W. (2011). Silicic segregations of the Ferrar dolerite sills. *Journal of Petrology*, 52, 1927-1964.

Zientek, M.L. (2012). Magmatic ore deposits in layered intrusions—Descriptive model for reef-type PGE and contact-type Cu-Ni-PGE deposits: U.S. Geological Survey Open-File Report 2012–1010, 48 p.

Figure 2.1: Karoo large igneous province (LIP) of South Africa. (a) Regional map showing the distribution of the saucer-shaped sills within the Karoo sedimentary basin (yellow units), as well as the associated Drakensberg Group lavas. Modified from [Ernst \(2014\)](#). Originally modified from [Svensen et al. \(2012\)](#).

Figure 2.2: Saucer-shaped geometry of mafic sills, typical of sills emplaced in sedimentary basins. Modified from [Ernst \(2014\)](#). Originally from [Chevallier and Woodford \(1999\)](#).

Figure 2.3: **a)** Ascending mantle plume beneath an Archean craton, within a few hundred kilometres of the original cratonic boundary. The plume head is mixture of high-temperature tail and entrained ambient upper mantle (Campbell *et al.*, 1989; Griffiths and Campbell, 1990). **b)** Plume head impinges on base of lithosphere and flattens. **c)** Plume head and tail are channeled to thinner lithosphere at craton margin, continental rifting is generated at original suture, and high-Mg, low-Ti melts are produced. It is beneath the cratonic margin where maximum melt production occurs, and this is controlled by flow of the plume along the sloping base of cratonic lithosphere (Begg *et al.*, 2010), and predominant flux of komatiite plume-tail melting is focused at cratonic margin. **d)** Both favourable and unfavourable environments for mineralisation are developed above the melting zone, as high flux “magma freeways” have potential for assimilation of crustal S, transport and deposition of magmatic sulphide ores. Modified Figure 1 from Barnes *et al.* (2015).

Chapter 3 – Regional Geology

3.1 Geological Framework

The bedrock geology of southern Baffin Island comprises polydeformed and metamorphosed plutonic and supracrustal rocks that form part of the northeastern (Québec-Baffin) segment of the Trans-Hudson orogen (THO); a broad, arcuate, Himalayan-scale collisional orogenic belt that extends from south-central to north-eastern North America ([Figure 3.1](#)) ([Hoffman, 1988](#); [Lewry and Collerson, 1990](#); [St-Onge et al., 2007](#); [Sanborn-Barrie et al., 2008](#)). The THO formed during the final collisional phase of Nuna supercontinent growth and records the suturing of the lower Superior and upper Churchill plates between 1.86 and 1.80 Ga, concomitantly with closure of the Manikewan Ocean ([Weller et al., 2015](#)). During the ca. 120 m.y. of plate convergence, the orogenic system experienced several accretionary events, involving several Archean crustal blocks and younger terranes ([Corrigan et al., 2009](#); [St-Onge et al., 2009](#)). The area affected by the THO in the eastern Arctic is outlined on the simplified geological map of [Figure 3.2](#).

Regional bedrock mapping has led to the subdivision of southern Baffin Island into four orogen-scale tectonically stacked domains ([Figure 3.3](#)), all separated by major deformation zones (Baffin, Soper River, and Bergeron sutures; [St-Onge et al., 2002, 2015f](#)). During the middle Paleoproterozoic, these tectonostratigraphic domains either accumulated on, or were accreted to the northern margin of the lower-plate Archean Superior craton ([St-Onge et al., 2000b, and references therein](#)). Collectively, the divergent and convergent margin tectonic activity occurred over a period greater than 200 million years.

From lowest to highest structural level, these domains include the Superior craton and margin, Narsajuaq terrane, Meta Incognita microcontinent, and Rae craton and margin.

The lower-plate Superior craton and margin (level 1; [Figure 3.3](#)) comprises Archean basement orthogneiss of the Superior craton, and the Povungnituk and Chukotat groups, middle Paleoproterozoic cover sequences that stratigraphically overlie, or are thrust imbricated above the Archean basement ([St-Onge et al., 1996](#); [St-Onge et al., 2000b](#)). The basement gneiss includes variably deformed granodioritic to syenogranitic plutonic bodies, emplaced in dominantly mafic tonalite-quartz diorite gneiss ([St-Onge et al., 2001](#)). Collectively, these have been interpreted to be the remnants of the plutonic root of several Archean-age magmatic arcs, with magmatism evolving temporally from calcic to more potassic ([Lucas and St-Onge, 1991](#)). Zircons from the tonalitic orthogneiss and granitoid plutons have been dated (U-Pb), yielding ages between $3220^{+32/-23}$ Ma and $2737^{±2}$ Ma ([Parrish, 1989, personal communication to M. St-Onge](#); [St-Onge et al., 1992](#); [Scott and St-Onge, 1995](#); [Wodicka and Scott, 1997](#)). The overlying Povungnituk Group comprises parautochthonous fluvial clastic rocks, carbonatitic volcanoclastic rocks, and a package of LREE-enriched continental tholeiitic flood basalts and rhyolites, which formed during the initial Paleoproterozoic rifting of the Superior Province ([St-Onge et al., 2001](#)) and are part of the Minto large igneous province (LIP) ([Buchan et al., 1998](#); [Hamilton et al. 2016](#); [Kastek et al. 2016](#)). The Chukotat Group comprises komatiitic to LREE-depleted tholeiitic basalts, which were extruded

along the northern Superior plate margin (St-Onge et al., 2001) and which are part of the plume-derived Circum-Superior LIP (e.g. Ernst, 2014; Minifie et al. 2013; Ciborowski et al. 2017). Rhyolite flows and mafic to ultramafic intrusions were dated to constrain the timing of the initial continental rifting events, yielding ages between $2038^{+4/-2}$ Ma and $1959^{+3.1/-2.7}$ Ma, and $1887^{+37/-11}$ Ma and 1870^{+4} Ma for the renewed rifting (Parrish, 1989; Machado et al., 1993; St-Onge et al., 1992; Wodicka et al., 2002a; R.R. Parish, unpublished data). Parautochthonous sedimentary and volcanic sequences are mostly preserved in the Cape Smith (thrust) Belt (Lucas, 1989a,b), where they are separated from allochthonous crustal elements of the Narsajuaq terrane by the Bergeron suture (St-Onge et al., 2001).

The Narsajuaq terrane (level 2; Figure 3.3) comprises: (1) an arc (Parent Group and Narsajuaq arc) of dominantly monzogranitic to granodioritic plutonic units, interpreted to be either a deformed magmatic arc terrane (Scott, 1997; St-Onge et al., 2009) or Narsajuaq-age intrusions (Corrigan et al. 2009); (2) the Watts group, a sequence of both mid-oceanic ridge and oceanic island basalts (Scott et al., 1992); and (3) the Spartan group, an undated fore-arc clastic apron (St-Onge et al., 1992). The magmatic arc includes two petrologically and temporally distinct suites: an older intra-oceanic suite and a younger Andean-style margin suite (Dunphy and Ludden, 1998). Calc-alkaline diorite-tonalite gneiss, basaltic andesite and rhyolite dominate the older suite, whereas cross-cutting monzodiorite to granite plutons comprise the younger suite (St-Onge et al., 1992). U-Pb ages from the intraoceanic suite range from 1863^{+2} Ma and

1845 $\pm$ 2 Ma ([Machado et al., 1993](#); [R.R. Parish, unpublished data](#)), and age from the Andean-style margin suite range between 1842 $^{+5/-3}$ Ma and 1820 $^{+4/-3}$ Ma ([Parrish, 1989, unpublished data](#); [Machado et al., 1993](#); [Scott, 1997](#); [Scott and Wodicka, 1998](#)). A zircon recovered from a mafic cumulate layer in the Watts Group ophiolite yielded a U-Pb age 1998 $\pm$ 2 Ma ([Parrish, 1989](#)). A number of kilometer-scale plutons, quartz diorite to monzogranite in composition, intrude into the Watts and Parent groups ([Dunphy, 1994](#)) and yield ages that temporally overlap the Narsajuaq arc proper. In addition, they share geochemical similarities, suggesting a lithotectonic correlation ([Dunphy and Ludden, 1998](#)). Thus, the intrusion of the Narsajuaq-like plutons into the Watts and Parent groups effectively stitched the allochthonous units of level 2 together ([St-Onge et al., 1992](#)). The allochthonous crustal elements of the Narsajuaq terrane are separated from the structurally overlying Archean and Paleoproterozoic rocks of the Meta Incognita microcontinent by the Soper River suture; the second south verging crustal suture found within southern Baffin Island ([St-Onge et al., 2001](#)).

The Meta Incognita microcontinent (level 3; [Figure 3.3](#)), the main focus of this study, comprises an Archean to Paleoproterozoic basement orthogneiss (Ramsay River orthogneiss; [Scott and Wodicka, 1998](#)) with its middle Paleoproterozoic (ca. 2100-1800 Ma) cover sequence (platformal-type continental margin) comprising psammite, semipelite, pelite, quartzite, marble and amphibolite strata (Lake Harbour Group; [Jackson and Taylor, 1972](#); [St-Onge et al., 2000](#)), a suggested overlying foreland basin succession (originally broken out from the Lake Harbour Group and referred to as the Blandford Bay

assemblage; [Scott et al., 1997](#)), and a suite of monzogranitic plutons (Cumberland Batholith; [Jackson et al., 1990](#)) that pervasively intrude all three of the aforementioned units ([St-Onge et al., 2001, 2007](#)). Additionally, there is a ca. 1900 Ma suite of sills, informally named the “Frobisher suite,” that extensively intrude the Lake Harbour Group metasedimentary rocks ([Liikane et al., 2015](#)). The age of the Ramsay river orthogneiss has been constrained between 3019 ± 5 Ma and 2784 ± 9 Ma ([Scott, 1998, 1999](#)), with a late-intruding monzogranite yielding an age of $1950^{+6/-4}$ Ma ([Scott and Wodicka, 1998](#)). Detrital zircons and crosscutting intrusions have been used to bracket the deposition age of both the shelf succession and foreland basin sequences of the Lake Harbour Group to between 1934 ± 2 and ca. 1880 Ma ([Figure 2.4](#); [Scott, 1997, unpublished data](#); [Scott et al., 1997, 2002](#)). The Cumberland Batholith plutonic units range in age between $1865^{+4/-2}$ Ma and $1848^{+/-2}$ Ma ([Jackson et al., 1990](#); [Wodicka and Scott, 1997](#); [Scott and Wodicka, 1998](#); [Scott, 1999](#)).

In central Baffin Island, the upper-plate Rae craton and margin (level 4; [Figure 3.3](#)) comprise Archean basement orthogneiss, overlain by middle Paleoproterozoic cover sequence of metasedimentary and metavolcanic units (Piling group; [Wodicka et al., 2014](#)), that exhibit stratigraphical similarities to the Hoare Bay Group on Cumberland Peninsula. The upper-plate Rae craton is comprised of felsic orthogneiss, metamorphosed supracrustal rocks, and younger felsic plutonic rocks ranging in age from $2868^{+13/-12}$ Ma and 2702 ± 3 Ma ([Jackson 2000](#); [Wodicka et al., 2002b](#); [Bethune and Scammel, 2003](#); [D.J. Scott, unpublished data](#); [N. Wodicka, unpublished data](#)). Overlying its southern margin,

the Paleoproterozoic Piling Group comprises rift, shelf margin and foredeep strata, which have ages constrained to between ca. 2159 ± 16 Ma and $1897^{+7/-4}$ Ma (Wodicka et al., 2002b; N. Wodicka, unpublished data, 2014). In addition to the Cumberland batholith, both basement and cover units are intruded by a felsic plutonic suite (the Qikiqtarjuaq suite) of middle Paleoproterozoic age (ca. 1.89 Ga); distinctly older than the 1.865-1.845 Ga Cumberland batholith (Rayner et al., 2012).

St-Onge (2009) has interpreted the Cumberland batholith to represent a period of Andean-style magmatism. Alternatively, Whalen et al. (2010) have interpreted the batholith to be the result of post-accretionary delamination of the lithosphere and subsequent mantle upwelling. All structural levels are crosscut by Franklin dykes that were emplaced during the plume-driven ca. 720 Ma Franklin LIP event, which was associated with the Rodinia supercontinent break-up (Fahrig, 1987; Heaman et al., 1992; Jowitt and Ernst, 2013; Ernst and Bleeker, 2010; Ernst et al., 2016). The 3rd and 4th levels are also unconformably overlain by limestone strata of Ordovician age (Blackadar, 1967).

3.2 Deformation Events and Metamorphism

Several tectonothermal magmatic events of Paleoproterozoic age have been recognized in southern Baffin Island, through integrated tectono-stratigraphic, structural and petrological studies (Figure 3.4; St-Onge et al.,

2007). This includes at least five regional episodes of compression (D_0 - D_4) and two thermal events (M_1 - M_2 ; [St-Onge et al., 2015f](#)).

The imbrication of basement and cover units resulted in map-scale D_0 tectonic contacts. The imbrication post-dates Lake Harbour Group deposition and Frobisher suite emplacement, and pre-dates Cumberland batholith emplacement ([St-Onge et al., 1999b](#)). Thrust imbrication best explains the field repetition of distinct tectonostratigraphic units, and is interpreted to result from accretion of the Meta Incognita microcontinent to the southern margin of the Rae craton and closure of the Baffin suture (BaS; [St-Onge et al., 2007](#)).

As convergence continued, the Narsajuaq arc was accreted to the Meta Incognita microcontinent ([St-Onge et al., 2007](#)). This D_{1a} event led to the closure of the Soper River suture (SRS) at the structural base of the Meta Incognita microcontinent, which occurred between 1845 ± 2 Ma ([Dunphy and Ludden, 1998](#)) and $1842^{+5/-3}$ Ma ([Scott, 1997](#)). These ages were obtained from the youngest unit of the Narsajuaq's intraoceanic phase and the oldest plutonic unit of the Narsajuaq's Andean-margin phase, respectively ([Dunphy and Ludden, 1998](#); [Scott, 1997](#)). Associated with the D_{1a} deformation event is a period of prograde granulite-facies metamorphism (M_{1a}) that is recorded in the Frobisher suite sills and host Lake Harbour Group metasedimentary rocks ([Skipton et al., 2016a,b](#)). The S_{1a} foliation formed during progressive metamorphism is generally shallowly dipping ($<45^\circ$) and parallel to supracrustal lithological contacts, and is interpreted to represent an enhancement of the primary bedding (S_0 ; [St-Onge et al., 2007](#); [Skipton et al., 2016a,b](#)). The result was near-complete to complete

recrystallization of primary assemblages, resulting in psammitic and pelitic units containing a subset of the granulite-facies assemblage garnet-cordierite-sillimanite-biotite-plagioclase-K-feldspar-quartz-melt, and mafic units containing a subset of the granulite-facies assemblage garnet-orthopyroxene-clinopyroxene-hornblende-biotite-plagioclase (St-Onge et al., 2007; Skipton et al., 2016a,b). Within the Frobisher suite, S_{1a} is locally defined by varying proportions of ferromagnesian minerals and plagioclase.

Following the granulite-facies metamorphic event, there was a high-T thermal perturbation event (M_{1b}) at ca. 1833-1829 Ma, related to the emplacement of felsic dykes and plugs (St-Onge et al., 2007). This event was recognized through the U-Pb dating of zircon and monazite from siliciclastic and gneiss samples (Scott & Gauthier, 1996; Wodicka & Scott, 1997; Scott & Wodicka, 1998; St-Onge et al., 2007). However, it is not clear whether this is a distinct thermal pulse or a protracted M_1 metamorphic event.

The final accretion event (D_2) marks the collision of the northern Churchill plate (i.e. Rae craton and accreted microcontinent, arc and ophiolitic units) with the Superior craton (St-Onge et al., 2007). This accretion event involved: (1) the closure of the Manikewan ocean, forming the Bergeron suture (BaS) and the resulting foreland thrust and fold belt within the Superior lower plate (Lucas, 1989; St-Onge et al., 2001); (2) the imbrication of tectonostratigraphic units in southern Baffin Island and further crustal shortening across the Soper River suture; (3) modification of pre-existing fabrics by D_2 - M_2 tectonothermal events. The terminal collision is bracketed between $1820^{+4/-3}$ Ma (Scott & Wodicka, 1998)

and 1795 ± 2 Ma (Wodicka & Scott, 1997); the ages of the youngest-dated plutonic rock in the suture's hanging wall and the age of a post-collisional syenogranite dyke cutting the Bergeron suture, respectively (St-Onge et al., 2007). The principal modifications of pre-existing fabrics resulting from D₂ deformation in southern Baffin Island, within the Bergeron suture's hanging wall, include a more finely layered S₂ foliation developed in metasedimentary units, and transposition of compositional layering within metaplutonic units (St-Onge et al., 2007). Associated with the D₂ deformation event was an M₂ retrograde amphibolite-facies metamorphism on southern Baffin Island, and prograde greenschist- to amphibolite-facies metamorphism within the cratonic margin of the lower-plate Superior craton margin (St-Onge et al., 2000b). This event is recorded in areas of D₂ deformation of psammitic and pelitic units of the Lake Harbour Group, as manifested by an S₂ fabric featuring the growth of biotite-sillimanite±garnet±cordierite at the expense of the previously described S_{1a} assemblages. The M₂ retrograde metamorphism occurred at ca. 1820 ± 1 to 1813 ± 2 Ma (Wodicka & Scott, 1997; Scott et al., 2002; St-Onge et al., 2007; Skipton et al., 2016a,b).

A set of regional north- to northwest-trending D₃ folds reorient preexisting D₁ and D₂ structures and fabrics (St-Onge et al., 2015f). These folds range from metre- to map-scale and consistently display west- to southwest-verging asymmetry. The D₃ folding involves all tectonostratigraphic units and can clearly be identified in the field and on air photos. In northern Quebec, the D₃ folding is constrained to ca. 1.76 Ga within structural levels 1 and 2 (Lucas and St-Onge, 1992). The orogen-parallel folds are proposed to have formed from continued

horizontal compression during post-collisional intra-continental shortening in the northeastern segment of the Trans-Hudson orogen (Lucas and Byrne, 1992).

Fold interference patterns exist as a result of D₄ upright folding deforming all other structural elements, with fold axis trending east-northeast (St-Onge et al., 2015f). D₄ generated dome-and-basin map patterns are best preserved in the southern and eastern portions of Meta Incognita Peninsula (Figure 3.5; St-Onge et al., 2015f). The structural relief generated by D₃ and D₄ folding allow for the crustal architecture of southern Baffin Island, and its bounding tectonic elements, to be studied. While D₁ and D₂ folds do contribute to map patterns, they contribute to a much lower degree.

In summary, five deformation events (D₀, D₁, D₂, D₃, D₄) occurred during the sequential north to south closure of three crustal sutures (Baffin suture, Soper River suture, and Bergeron suture, respectively). D₁ is associated with a regional granulite-facies metamorphic event M_{1a}, which is associated with the development of a dominant S_{1a} compositional foliation. A subsequent regional retrograde amphibolite-facies metamorphic event (M₂) is associated with the development of a penetrative schistose fabric (S₂). Two later localized thermal events (M_{1b} and post-D₂) occurred as a result of crustal magmatism, with the latter recorded in Frobisher suite titanite grains (St-Onge et al., 2007). Finally, regional D₃ and D₄ folding events generated interference patterns within the units of southern Baffin Island that are exposed on the present-day erosional surface.

3.3 Lake Harbour Group

The middle Paleoproterozoic Lake Harbour Group (LHG) of southern Baffin Island comprises psammite, semipelite, pelite, quartzite, marble and amphibolite strata (Figure 3.6a-c), which have been interpreted to represent a platform-type continental margin that formed along the microcontinent after ca. 2010 Ma, followed by foredeep sedimentation between 1915±8 Ma and ca. 1880 Ma (Blackadar, 1967; Jackson and Taylor, 1972; Scott, 1997; Wodicka et al., 2008). Basin closure and the deposition of foredeep sediments along margins appear to be related to the ca. 1.9 Ga Snowbird collisional orogen (Wodicka et al., 2008). The type locality for this succession is situated north of the community of Kimmirut (Figure 3.8; St-Onge et al., 1996, 1998; Scott et al., 1997). These rocks are interpreted to have been deposited between 2.01 and 1.90 Ga (Wodicka, 2010), and deformed at 1845 –1842 Ma (Dunphy and Ludden, 1998; Scott, 1997).

Metasedimentary rocks of the Lake Harbour Group were deformed during the Trans-Hudson orogen, and are now preserved in an obliquely exposed continental collisional zone. Relative to an undeformed, un-metamorphosed, “reference” microcontinent (i.e. at ca. 1900 Ma), the LHG sequence preserved on Meta Incognita Peninsula represents the more “proximal” facies (i.e. most inland-deposited sediments) and would have formed on the furthest from the margin of the microcontinent. The LHG sequence preserved on northern Hall Peninsula represents the more “distal” facies (i.e. deeper water, off-shelf strata, more-seaward sequence, closest to the edge of the microcontinent).

On Meta Incognita Peninsula ([Figure 3.8](#)), the Lake Harbour Group exists as two lithologically and geographically distinct sequences ([St-Onge et al., 2014](#)). The first sequence is found throughout Meta Incognita Peninsula to the west of York Sound ([Figure 3.8](#)) and comprises variable amounts of garnetiferous psammite, quartzite, and minor pelite and semipelite that are structurally overlain by laterally continuous and boudinaged bands of marble and calc-silicate strata ('Kimmirut sequence' of [Scott et al., 1997](#)). The second sequence is comprised of garnetiferous psammite interlayered with pelite/semipelite, and volumetrically dominates the far eastern part of the peninsula, including exposures on eastern islands and bluffs, and also around Markham Bay ([Figure 3.8](#); 'Markham Bay sequence' of [Scott et al., 1997](#)). Additionally, the latter areas are largely devoid of marble/calc-silicate layers.

On Hall Peninsula, the siliciclastic units bear a strong resemblance to the metasedimentary units of the Lake Harbour Group found throughout Meta Incognita Peninsula. Quartzite occurs as strongly recrystallized, well-layered, rusty weathering panels, up to hundreds of meters in thickness, and is overlain by interbedded psammite, semipelite and/or pelite. Layers within the quartzite vary compositionally from orthoquartzite to feldspathic quartzite; migmatitic psammite and semipelite contain biotite±garnet and biotite±garnet±sillimanite assemblages, respectively ([Weller, et al., 2015](#)). Horizons of pelite occur with the assemblage biotite-garnet-sillimanite-K-feldspar-melt. Siliciclastic units often contain gossanous horizons ([Figure 3.6d](#)), in which graphite, pyrite, and chalcopyrite are found ([Weller, et al., 2015](#)). Marble and calcsilicate units occur

as panels up to 100 m thick, and occur structurally above (faulted) the aforementioned siliciclastic units. The carbonate units are compositionally layered, due to varying proportions of medium to coarse-grained calcite, humite, forsterite, apatite, diopside, magnetite, phlogopite, scapolite, spodumene, titanite, and tremolite ([Weller, et al., 2015](#)).

Two additional stratigraphic sequences of metasedimentary rocks (the Lona Bay and Schooner Harbour sequences) have been recognized to the west of Meta Incognita peninsula on Foxe Peninsula, where they structurally overlie the Lake Harbour Group ([Sanborn-Barrie et al., 2008](#)). The Lona Bay sequence is a succession of relatively homogeneous arenaceous metasedimentary rocks centred around Lona Bay ([Figure 3.4](#)). The sequence typically has 10-30 cm thick beds, highlighted by cross-stratification, variations in grain size, and the heterogeneous distribution of porphyroblasts. The overlying Schooner Harbour sequence is a heterogeneous volcanic-bearing supracrustal belt of dominantly alkaline basaltic composition, extending for 100 km along southern Foxe peninsula, from Schooner Harbour eastward to the West Foxe Islands ([Figure 3.4; Wodicka et al., 2008](#)). The lower part of the succession contains at its base fine-grained, metapelitic, sulphidic slate and schist with garnet, staurolite or andalusite porphyroblasts, and gradually becomes more psammitic up-section ([Sanborn-Barrie et al., 2008](#)). Also in the lower sequence are orthoquartzite units 1 m thick are found alternating with small exposures of amphibolite, and discontinuous horizons of <1 m thick polymictic metaconglomerate are also present, locally. In the upper part of the Schooner Harbour sequence, volcanic

units are dominantly alkali basaltic to andesitic in composition, and contain some variolitic and lapilli-tuff units. In the vicinity of Schooner Harbour, extrusive ultramafic units are present in what appears to be komatiitic lapillistone (Sanborn-Barrie et al., 2008). Recent geochronology has constrained the age of the Schooner Harbour sequence to ca. 2099 Ma (Wodicka et al., 2010), suggesting that it is older than the Lake Harbour Group. The Schooner Harbour sequence has been interpreted to potentially represent initial rifting within the Meta Incognita microcontinent (Wodicka et al., 2008).

3.4 Frobisher Suite Mafic, Ultramafic and Mafic-Ultramafic Sills

The siliciclastic and carbonate strata of the Lake Harbour Group, and the orthoquartzite and semipelite of the lower Schooner Harbour sequence are intruded extensively by the mafic to ultramafic suite of sills herein referred to as the Frobisher suite (Liikane et al., 2015). The Frobisher suite (informally named after the suite's geographic proximity to Frobisher Bay; Figure 3.4) is a suite of mafic, ultramafic, and mafic-ultramafic sills (Figure 3.7a) found throughout most of the Meta Incognita microcontinent terrane of southern Baffin Island, Nunavut (Liikane et al., 2015). Their distribution throughout Meta Incognita Peninsula and south-central Baffin Island can be seen in Figures 3.8 and 3.9, respectively. However, it should be noted that given the reconnaissance-scale of mapping, it is likely that the majority of sills have yet to be mapped. The sills generally comprise concordant to slightly discordant sheets of variably altered coarse-grained metaperidotite, medium- to coarse-grained metaperidotite-metagabbro, and

medium-grained metagabbro emplaced into the siliciclastic strata of the Lake Harbour Group (Liikane et al., 2015), and lower Schooner Harbour sequence (Sanborn-Barrie et al., 2008). Their interpreted primary magmatic lithology (e.g. pyroxenite) is derived primarily from CIPW norms, estimated based on major element geochemistry. The layered mafic-ultramafic sills vary lithologically from metapyroxenite at the base, overlain by metaperidotite, and capped by metagabbro to metaleucogabbro (Figure 3.7b; Liikane et al., 2015). If the layering is a result of gravity-induced magmatic differentiation from a single pulse of magma, the ultramafic to mafic progression provides a way-up indicator (see section 2.2.3 for discussion on the mechanisms of layer formation in layered intrusions). All layering within the sills is parallel to the intrusive contacts (Figure 3.10), consistent with their emplacement as strata-bound intrusions. Evidence of partial melting (granophyric intrusive contacts; St-Onge et al., 2015a) or of enclaves of host strata are absent (Liikane et al., 2015). Numerous intrusions contain disseminated sulphide mineralization (see Chapter 5 for more details), and deep orange gossans occur in the host strata immediately adjacent to some of the sills (Liikane et al., 2015).

In map view, the sills typically appear elongate to ellipsoid or annular in shape (Figure 3.11). The apparent tightness of the acutely folded sills varies from gentle to tight, and the bluntness varies from rounded (in parallel folds) to angular (in similar folds). Apparent thicknesses of sills typically vary between 5 and 20 m - although a few exceed 100 m in thickness - and in most cases extend several kilometres along strike (Liikane et al., 2015).

The current erosional surface of southern Baffin Island exposes the Frobisher suite sills in such a way that it may give the impression of a map pattern characterized by primary saucer-shaped intrusions ([Figure 1.1](#)). However, consideration of host/intrusion field relationships suggests that in fact the geometry of the sills is best interpreted as a product of the complex regional deformation history in this part of southern Baffin Island. Specifically, [St-Onge et al., \(2015f\)](#) have documented that the D₃ and D₄ regional folding events are responsible for generating the pervasive fold interference patterns within the metasedimentary units of the Lake Harbour Group. The genetic deformational link between sill and host rock is suggested by (a) sill outcrop patterns that mimic the outcrop patterns of the enveloping host rocks, and (b) the concordant nature of intrusive contacts and compositional layering within both sill and adjacent host metasedimentary strata. This suggests that the emplacement of the sills predate these folding events.

The Frobisher suite sills must be younger than the Lake Harbour group (ca. 2.01 - 1.90 Ga), and older than the 1845-1842 Ma M_{1a} deformation event. More recent geochronological data arrived at a minimum crystallization age of ca. 1900 Ma for the Frobisher suite on southern Baffin Island ([Liikane et al., 2015](#)).

To determine the age of the Frobisher suite sills, a sample (14SAB-L017A; location shown in [Figure 3.8](#) as “Killapait sill”) collected from the leucogabbroic portion of the Killapait sill (discussed in 4.1.2) of south-central Meta Incognita Peninsula was submitted for U-Pb geochronology, with the sensitive high-resolution ion microprobe (SHRIMP) and with isotope dilution–thermal ionization

mass spectrometry (ID-TIMS), at the Geological Survey of Canada (Liikane et al., 2015). From this sample, abundant, high-quality, clear, colourless zircon grains were recovered (Liikane et al., 2015). The zircon grains are morphologically diverse: anhedral (Figure 3.12a), prismatic (Figure 3.12b) or rounded (not shown), and simply zoned (Figure 3.12c, d) with no apparent core/overgrowths (Liikane et al., 2015). Twenty-five analyses were obtained via SHRIMP, of which 22 zircon grains encompassing the full morphological range yielded ages ranging from 2004 to 1833 Ma (Figure 3.12e), with a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of $1922 \pm 12 \text{ Ma}$ (2σ , $n=25$, $\text{MSWD} = 1.9$; Liikane et al., 2015). Four single-grain ID-TIMS fractions were analyzed, yielding slightly discordant $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 1876.9 and 1896.6 Ma (Figure 3.12e, inset; Liikane et al., 2015). The results from TIMS indicate some cryptic complexity in the zircon microstructure or U-Pb systematics that only becomes apparent in high spatial resolution analyses (Liikane et al., 2015). All grains also have Th/U ratios greater than 1, which is a common attribute of zircon recovered from mafic lithologies (Wang, et al., 2011). The concentric oscillatory zoning is consistent with magmatic crystallization (Corfu, 2003) and the high Th/U signature suggests an igneous origin for the zircon (Heaman et al., 1990). Based on the collective data, these preliminary results suggest a minimum crystallization age of ca. 1900 Ma (Liikane et al., 2015).

3.4.1 Mafic units

The majority of mafic units within the mafic and mafic-ultramafic sills exhibit faint, to well-defined, centimetre- to metre-scale rhythmic layering (Figure 3.13a) defined by variations in the modal abundance of plagioclase, amphibole, clinopyroxene and orthopyroxene (Figure 3.13b; Liikane et al., 2015). In contrast, some of the sills lack any visible layering. These units are interpreted to represent metamorphosed gabbroic to noritic lithologies.

3.4.2 Ultramafic units

The ultramafic units with ultramafic and mafic-ultramafic intrusions comprise layered clinopyroxene-orthopyroxene-olivine metaperidotite (Figure 3.13c-d) and clinopyroxene-orthopyroxene metapyroxenite (Figure 3.13e-f; Liikane et al., 2015). As with the metagabbro intrusions, some of the ultramafic intrusions are homogeneous and lack any visible layering.

Petrographic observations of the sills are described in Chapter 5, mineral and bulk-rock geochemical characteristics are explored in Chapters 6 & 7, respectively, and their petrogenesis is addressed in Chapter 8.

References

- Blackadar, R.G. (1967). Geological reconnaissance, southern Baffin Island, District of Mackenzie. Geological Survey of Canada, Paper 66–47, p. 32.
- Bethune, K. M. and Scammell, R. J. (2003). Geology, geochronology, and geochemistry of Archean rocks in the Ege Bay area, north-central Baffin Island, Canada: constraints on the depositional and tectonic history of the Mary River Group of northeastern Rae Province. *Canadian Journal of Earth Sciences* 40, 1137-1167.
- Blackadar, R.G. (1967). Geological reconnaissance, southern Baffin Island, District of Mackenzie. Geological Survey of Canada, Paper 66–47, p. 32.
- Buchan, K.L., Mortensen, J.K., Card, K.D., and Percival, J.A. (1998). Paleomagnetism and U–Pb geochronology of diabase dyke swarms of Minto Block, Superior Province, Quebec, Canada. *Canadian Journal of Earth Sciences*, 35: 1054–1069.
- Ciborowski, T.J.R., Minifie, M.J. Kerr, A.C., Ernst, R.E., Baragar, B., Millar, I.L. (2017). A mantle plume origin for the Palaeoproterozoic Circum-Superior Large Igneous Province. *Precambrian Research*, v. 294, p. 189-213.
- Corrigan, D., Pehrsson, S., Wodicka, N. and de Kemp, E. (2009). The Palaeoproterozoic Trans Hudson Orogen: a prototype of modern accretionary processes; in *Ancient Orogens and Modern Analogues*, J.B. Murphy, J.D. Keppie and A.J. Hynes (ed.), The Geological Society, London, Special Publications, v. 327, p. 457–479. doi:10.1144/SP327.19.

- Dunphy, J.M. (1994). Evolution des roches plutoniques du domaine nord de la Fosse de l'Ungava. Ministère de l'Énergie et des Ressources du Québec manuscript brut, MB 94-58: 48 p.
- Dunphy, J.M., Ludden, J.N. (1998). Petrological and geochemical characteristics of a Paleoproterozoic magmatic arc (Narsajuaq terrane, Ungava Orogen, Canada) and comparisons to Superior Province granitoids. *Precambrian Res.* 91, 109–142.
- Ernst, R.E. (2014). *Large Igneous Provinces*. Cambridge University Press, 653 p.
- Ernst, R.E. and Bleeker, W. (2010). Large igneous provinces (LIPs), giant dyke swarms, and mantle plumes: significance for breakup events within Canada and adjacent regions from 2.5 Ga to the Present. *Canadian Journal of Earth Sciences*, 47: 695–739.
- Ernst, R.E., Hamilton, M.A., Söderlund, U., Hanes, J.A., Gladkochub, D.P., Okrugin, A.V., Kolotilina, T., Mekhonoshin, A.S., Bleeker, W., LeCheminant, A.N., Buchan, K.L., Chamberlain, K.R., Didenko, A.N., (2016a). Long-lived connection between southern Siberia and northern Laurentia in the Proterozoic. *Nature Geoscience*, 9:464–469.
- Fahrig, W.F. (1987). The tectonic setting of continental mafic dyke swarms: failed arm and early passive margin. In Halls, H.C. & Fahrig, W.F. (eds.), *Mafic Dyke Swarms*. St John's, NL: Geological Association of Canada, Special Publication 34, pp. 331–348.
- Hamilton, M.A., Goutier, J., Buchan, K.L. (2016). Minto Large Igneous Province: a 2.00 Ga mafic magmatic event in the eastern Superior craton based on

U-Pb baddeleyite geochronology and paleomagnetism, IDC 7 conference, Beijing China, August 2016.

Heaman, L.M., LeCheminant, A.N. and Rainbird, R.H. (1992). Nature and timing of Franklin igneous events, Canada: implications for a Late Proterozoic mantle plume and the breakup of Laurentia; *Earth and Planetary Science Letters*, v. 109, p. 117–131.

Hoffman, P.F. (1988): United Plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia; *Annual Reviews of Earth and Planetary Sciences*, v. 16, p. 543–603.

Jackson, G. D. (2000). Geology of the Clyde-Cockburn Land Map Area, North-Central Baffin Island, Nunavut. Geological Survey of Canada Memoir 440, 303 pp.

Jackson, G.D., Hunt, P.A., Loveridge, W.D., Parrish, R.R. (1990). Reconnaissance geochronology of Baffin Island, N.W.T. Geological Survey of Canada Paper 89-2, 123–148.

Jackson, G.D., Taylor, F.C. (1972). Correlation of major Archean rock units in the northeastern Canadian Shield. *Canadian Journal of Earth Sciences* 9, 1650–1669.

Jowitt, S.M., Ernst, R.E. (2013). Geochemical Assessment of the Metallogenic Potential of Proterozoic LIPs of Canada: *Lithos*, v. 174, p. 291-307.

Kastek, N., Ernst, R.E., Baragar, W.R.A., Söderlund, U., Kamo, S.L., Bleeker, W., Sylvester, P. (2016). U-Pb Geochronology and geochemistry of the Povungnituk Group of the Cape Smith belt: a part of a craton-scale circa

2.0 Ga large igneous province (LIP), northern Superior craton. GAC-MAC annual meeting 2016 June 2, Whitehorse, Yukon.

Lewry, J.F. and Collerson, K.D. (1990). The Trans-Hudson Orogen; extent, subdivisions, and problems; in *The Early Proterozoic Trans-Hudson Orogen of North America*, J.F. Lewry and M.R. Stauffer (ed.), Geological Association of Canada, Special Paper 37, p. 1–14.

Liikane, D.A., St-Onge, M.R., Kjarsgaard, B.A., Rayner, N.M., Ernst, R.E. and Kastek, N. (2015). Frobisher suite mafic, ultramafic and layered mafic-ultramafic sills, southern Baffin Island, Nunavut; in *Summary of Activities 2015*, Canada-Nunavut Geoscience Office, p. 21–32.

Lucas, S.B. (1989a). Structural evolution of the Cape Smith Thrust Belt and the role of out-of-sequence faulting in the thickening of mountain belts. *Tectonics* 8, 655–676.

Lucas, S.B. (1989b). Geometrical, rheological and mechanical evolution of continental thrust belts: examples from the Cape Smith Belt, northern Quebec, Canada. PhD thesis, Brown University, Providence, Rhode Island, 253 p.

Lucas, S.B. and Byrne, T. (1992). Footwall involvement during arc-continent collision, Ungava orogen, northern Canada; *Journal of the Geological Society of London*, v. 149, p. 237–248.

Lucas, S.B. and St-Onge, M.R. (1991). Evolution of Archean and early Proterozoic magmatic arcs in northeastern Ungava Peninsula, Quebec. *Geological of Survey Canada Paper* 91-1C, 109–119.

- Minifie, M. J., Kerr, A. C., Ernst, R. E., Hastie, A. R., Ciborowski, T. J. C., Desharnais, G. & Milliar, I. L. (2013). The northern and southern sections of the western ca. 1880Ma Circum- Superior Large Igneous Province, North America: The Pickle Crow dyke connection? *Lithos* 174, 217–235.
- Parrish, R.R. (1989). U–Pb geochronology of the Cape Smith Belt and Sugluk block. *Geoscience Canada* 16, 126–130.
- Rayner, N.M., Sanborn-Barrie, M., Young, M.D. and Whalen, J.B. (2012). U-Pb ages of Archean basement and Paleoproterozoic plutonic rocks, southern Cumberland Peninsula, eastern Baffin Island, Nunavut; Geological Survey of Canada, Current Research 2012-8, 24 p.
- Sanborn-Barrie, M., St-Onge, M.R., Young, M.D., and James, D.T. (2008). Bedrock geology of south-western Baffin Island, Nunavut: expanding the tectonostratigraphic framework with relevance to mineral resources; Geological Survey of Canada, Current Research 2008-6, 16p.
- Scott, D.J. (1997). Geology, U–Pb, and Pb–Pb geochronology of the Lake Harbour area, southern Baffin Island: implications for the Paleoproterozoic tectonic evolution of northeastern Laurentia. *Canadian Journal of Earth Sciences* 34, 140–155.
- Scott, D.J. (1998). U–Pb ages of Archean crust in the southeast arm of the Rae Province, southeastern Ungava Bay, Quebec. Geological Survey of Canada Paper 1998-F, 41–45.
- Scott, D. J. & Gauthier, G. (1996). Comparison of TIMS (U-Pb) and laser ablation microprobe ICP-MS (Pb) techniques for age determination of detrital

-
- zircon from Paleoproterozoic metasedimentary rocks from northeastern Laurentia, Canada, with tectonic implications. *Chemical Geology* 131, 127-142.
- Scott, D.J., Stern, R.A., St-Onge, M.R. and McMullen, S.M. (2002). U-Pb geochronology of detrital zircons in metasedimentary rocks from southern Baffin Island: implications for the Paleoproterozoic tectonic evolution of Northeastern Laurentia; *Canadian Journal of Earth Sciences*, v. 39, p. 611–623.
- Scott, D.J. and Wodicka, N. (1998). A second report on the U–Pb geochronology of southern Baffin Island, Northwest territories. *Geological Survey of Canada Paper* 1998-F, 47–57.
- Scott, D.J., Helmstaedt, H., Bickle, M.J. (1992). Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada: a reconstructed section of Early Proterozoic oceanic crust. *Geology* 20, 173–176
- Scott, D.J. (1997). Geology, U–Pb, and Pb–Pb geochronology of the Lake Harbour area, southern Baffin Island: implications for the Paleoproterozoic tectonic evolution of northeastern Laurentia. *Canadian Journal of Earth Science*, 34, 140–155.
- Scott, D.J. and St-Onge, M.R. (1995). Constraints on Pb closure temperature in titanite based on rocks from the Ungava orogen, Canada: implications for U–Pb geochronology and P-T-t path determinations. *Geology* 23, 1123–1126.

- Scott, D.J., St-Onge, M.R., Wodicka, N. and Hanmer, S. (1997). Geology of the Markham Bay Crooks Inlet area, southern Baffin Island, Northwest Territories; in Geological Survey of Canada, Current Research 1997-C, p. 157–166.
- Skipton, D.R., Schneider, D.A., McFarlane, C.R.M., St-Onge, M.R., and Jackson, S. (2016a). Multi-stage zircon and monazite growth revealed by depth-profiling and in situ U-Pb geochronology: resolving the Paleoproterozoic tectonics of the Trans-Hudson Orogen on southeastern Baffin Island, Canada” Submitted to Precambrian Research, 285, 272-298. <https://doi.org/10.1016/j.precamres.2016.09.002>
- Skipton, D.R., St-Onge, M.R., Schneider, D.A., and McFarlane, C.R.M. (2016b). Tectonothermal evolution of the middle crust in the Trans-Hudson Orogen, Baffin Island, Canada: evidence from petrology and monazite geochronology of sillimanite-bearing migmatites, *Journal of Petrology*, 57 (8): 1437-1462. DOI: <https://doi.org/10.1093/petrology/egw046>
- St-Onge, M.R., Hanmer, S. and Scott, D.J. (1996). Geology of the Meta Incognita Peninsula, south Baffin Island: tectonostratigraphic units and regional correlations; in Geological Survey of Canada, Current Research 1996-C, p. 63–72.
- St-Onge, M.R., Lucas, S.B., Parrish, R.R. (1992). Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 1: tectonostratigraphic assemblages and their tectonic implications. *Canadian Journal of Earth Sciences*, 29, 746–764.

-
- St-Onge, M.R., Scott, D.J. and Lucas, S.B. (2000). Early partitioning of Quebec: Microcontinent formation in the Paleoproterozoic; *Geology*, v. 28, p. 323–326.
- St-Onge, M.R., Scott, D.J. and Wodicka, N. (1999a). *Geology, Frobisher Bay, Nunavut*; Geological Survey of Canada, Map 1979A, 1:100 000 scale, doi: 10.4095/210833
- St-Onge, M.R., Scott, D.J. and Wodicka, N. (1999b). *Geology, Hidden Bay, Nunavut*; Geological Survey of Canada, 1980A, 1:100 000 scale, doi: 10.4095/210835
- St-Onge, M.R., Scott, D.J. and Wodicka, N. (1999c). *Geology, McKellar Bay, Nunavut*; Geological Survey of Canada, Map 1981A, 1:100 000 scale, doi: 10.4095/210836
- St-Onge, M.R., Scott, D.J. and Wodicka, N. (1999d). *Geology, Wright Inlet, Nunavut*; Geological Survey of Canada, Map 1982A, 1:100 000 scale, doi: 10.4095/210840
- St-Onge, M.R., Scott, D.J. and Wodicka, N. (1999e). *Geology, Blandford Bay, Nunavut*; Geological Survey of Canada, Map 1983A, 1:100 000 scale, doi: 10.4095/210837
- St-Onge, M.R., Scott, D.J. and Wodicka, N. (1999f). *Geology, Crooks Inlet, Nunavut*; Geological Survey of Canada, Map 1984A, 1:100 000 scale, doi: 10.4095/210838

- St-Onge, M.R., Scott, D.J. and Wodicka, N. (1999g). Geology, White Strait, Nunavut; Geological Survey of Canada, Map1985A, 1:100 000 scale, doi: 10.4095/210839
- St-Onge, M. R., Scott, D. J. and Wodicka, N. (1999h). Geological maps and descriptive notes and legend, parts of Meta Incognita Peninsula, Nunavut. Geological Survey of Canada Maps 1979A to 1985A, scale 1:100 000.
- St-Onge, M.R., Scott, D.J., and Wodicka, N. (2001). Terrane boundaries within Trans-Hudson Orogen (Quebec – Baffin segment), Canada: changing structural and metamorphic character from foreland to hinterland; *Precambrian Research*, v. 107, p. 75–91.
- St-Onge, M.R., Scott, D.J. and Wodicka, N. (2002). Review of crustal architecture and evolution in the Ungava Peninsula– Baffin Island area: connection to the Lithoprobe ECSOOT transect; *Canadian Journal of Earth Sciences*, v. 39, p. 589-610, doi:10.1139/E02-022.
- St-Onge, M.R., Scott, D.J., Wodicka, N. and Lucas, S.B. (1998). Geology of the McKellar Bay Wight Inlet–Frobisher Bay area, southern Baffin Island, Northwest Territories; in Geological Survey of Canada, *Current Research 1998-C*, p. 43–53.
- St-Onge, M.R., Rayner, N.M., Liikane, D. and Chadwick, T. (2015a). Mafic, ultramafic and layered mafic-ultramafic sills, Meta Incognita Peninsula, southern Baffin Island, Nunavut; in *Summary of Activities 2014*, Canada-Nunavut Geoscience Office, p. 11–16.

-
- St-Onge, M.R., Rayner, N.M., Steenkamp, H.M., and Gilbert, C. (2015b).
Geology, Grinnell Glacier, Baffin Island, Nunavut; Geological Survey of
Canada, Canadian Geoscience Map 217E (preliminary); Canada-Nunavut
Geoscience Office, Open File Map 2015-04E, scale 1:100 000. doi:
10.4095/296111.
- St-Onge, M.R., Rayner, N.M., Steenkamp, H.M., and Gilbert, C. (2015c).
Geology, Pritzler Harbour, Baffin Island, Nunavut; Geological Survey of
Canada, Canadian Geoscience Map 216E (preliminary); Canada-Nunavut
Geoscience Office, Open File Map 2015-03E, scale 1:100 000. doi:
10.4095/296109.
- St-Onge, M.R., Rayner, N.M., Steenkamp, H.M., and Gilbert, C. (2015d).
Geology, Terra Nivea, Baffin Island, Nunavut; Geological Survey of
Canada, Canadian Geoscience Map 215E (preliminary); Canada-Nunavut
Geoscience Office, Open File Map 2015-02E, scale 1:100 000. doi:
10.4095/296104.
- St-Onge, M.R., Rayner, N.M., Steenkamp, H.M., and Gilbert, C. (2015e).
Geology, Pritzler Harbour, Baffin Island, Nunavut; Geological Survey of
Canada, Canadian Geoscience Map 216E (preliminary); Canada-Nunavut
Geoscience Office, Open File Map 2015-03E, scale 1:100 000. doi:
10.4095/296109.
- St-Onge, M.R., Rayner, N.M., Steenkamp, H.M. and Skipton, D.R. (2015f).
Bedrock mapping of eastern Meta Incognita Peninsula, southern Baffin

-
- Island, Nunavut; in Summary of Activities 2014, Canada-Nunavut Geoscience Office, p. 105–118.
- St-Onge, M.R., Wodicka, N. and Ijewliw, O. (2007). Polymetamorphic evolution of the Trans Hudson Orogen, Baffin Island, Canada: Integration of petrological, structural and geochronological data; *Journal of Petrology*, v. 48, p. 271– 302. doi:10.1093/petrology/eg1060.
- St-Onge, M.R., Van Gool, J.A.M., Garde, A.A. and Scott, D.J. (2009). Correlation of Archaean and Palaeoproterozoic units between northeastern Canada and western Greenland: constraining the pre-collisional upper plate accretionary history of the Trans-Hudson orogen; in *Earth Accretionary Systems in Space and Time*, P.A. Cawood and A. Kroner, The Geological Society, London, Special Publications, v. 318, p. 193–235. doi:10.1144/SP318.7.
- St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016a). Geology, Sylvia Grinnell Lake (north), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 253; Canada-Nunavut Geoscience Office, Open File Map OFM2016-10, scale 1:100 000, 1 sheet.
- St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016b). Geology, Sylvia Grinnell Lake (south), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 254; Canada-Nunavut Geoscience Office, Open File Map OFM2016-11, scale 1:100 000, 1 sheet.

St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016c). Geology, Amittok Lake (north), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 255; Canada-Nunavut Geoscience Office, Open File Map OFM2016-12, scale 1:100 000, 1 sheet.

St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016d). Geology, Amittok Lake (south), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 256; Canada-Nunavut Geoscience Office, Open File Map OFM2016-13, scale 1:100 000, 1 sheet.

St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016e). Geology, Clearwater Fiord (north), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 257; Canada-Nunavut Geoscience Office, Open File Map OFM2016-14, scale 1:100 000, 1 sheet.

St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016f). Geology, Clearwater Fiord (south), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 258; Canada-Nunavut Geoscience Office, Open File Map OFM2016-15, scale 1:100 000, 1 sheet.

St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016g). Geology, McKeand River (north), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 259; Canada-

-
- Nunavut Geoscience Office, Open File Map OFM2016-16, scale 1:100 000, 1 sheet.
- St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016h). Geology, McKeand River (south), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 260; Canada-Nunavut Geoscience Office, Open File Map OFM2016-17, scale 1:100 000, 1 sheet.
- St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016i). Geology, Irvine Inlet (north), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 261; Canada-Nunavut Geoscience Office, Open File Map OFM2016-18 scale 1:100 000, 1 sheet.
- St-Onge, M.R., Weller, O.M., Dyck, B.J., Rayner, N.M., Chadwick T., and Liikane, D. (2016j). Geology, Irvine Inlet (south), Baffin Island, Nunavut. Geological Survey of Canada, Canadian Geoscience Map 262; Canada-Nunavut Geoscience Office, Open File Map OFM2016-19, scale 1:100 000, 1 sheet.
- Weller, O.M., Dyck, B.J., St-Onge, M.R., Rayner, N.M. and Tschirhart, V. (2015g). Completing the bedrock mapping of southern Baffin Island, Nunavut: plutonic suites and regional stratigraphy; in Summary of Activities 2015, Canada-Nunavut Geoscience Office, p. 33–48.
- Whalen, J.B., Wodicka, N., Taylor, B.E. and Jackson, G.D. (2010). Cumberland batholith, Trans- Hudson Orogen, Canada: petrogenesis and implications

-
- for Paleoproterozoic crustal and orogenic processes; *Lithos*, v. 117, p. 99–118, doi: 10.1016/j.lithos.2010.02.008.
- Wodicka, N., Madore, L., Larbi, Y. & Vicker, P. (2002a). Geochronologie U-Pb de filons-couches mafiques de la Ceinture de Cape Smith et de la Fosse du Labrador. Seminaire d'information sur la recherche geologique, Programme et resumes 2002. Ministere des Ressources naturelles, Quebec DV 2002-10, 48.
- Wodicka, N., Scott, D.J. (1997). A preliminary report on the U–Pb geochronology of the Meta Incognita Peninsula, southern Baffin Island, Northwest Territories. *Geological Survey of Canada Paper* 1997-C, 167–178.
- Wodicka, N., St-Onge, M.R., Corrigan, D., Scott, D.J. and Whalen, J.B. (2014). Did a proto-ocean basin form along the southeastern Rae cratonic margin? Evidence from U-Pb geochronology, geochemistry (Sm-Nd and whole-rock), and stratigraphy of the Paleoproterozoic Piling Group, northern Canada; *GSA Bulletin*, v. 126, p. 1625–1653. doi:10.1130/B31028.1
- Wodicka, N., St-Onge, M. R., Scott, D. J. and Corrigan, D. (2002b). Preliminary report on the U-Pb geochronology of the northern margin of the Trans-Hudson Orogen, central Baffin Island, Nunavut. *Geological Survey of Canada Paper* 2002-F7, , 12 pp.
- Wodicka, N., St-Onge, M. R., Whalen, J. B. (2008). Characteristics of two opposing continental margin successions in northeast Laurentia; *Geochimica et Cosmochimica Acta*, v. 72, p. A1030

Wodicka, N., Whalen, J. B., St-Onge, M. R., and Corrigan, D. (2010). Meta Incognita microcontinent revisited: insights from U-Pb geochronology and Nd isotopes. In Geological Association of Canada-Mineralogical Association of Canada, Abstract (Vol. 35).

Figure 3.1: Summary geological map of North America and Greenland (collectively termed Laurentia), and the western Baltic Shield, modified from [Hoffman \(1988\)](#). This map outlines the extent of the Trans-Hudson orogen and the location of bounding Archean cratons and crustal blocks. The collisional lower plate of Trans-Hudson orogen comprises the Archean Superior craton and its marginal Paleoproterozoic supracrustal cover. The collisional upper plate (the Churchill domain) includes the Rae craton, Slave craton, Wyoming craton, Hearne block, and North Atlantic (Nain) craton, together with their associated Palaeoproterozoic supracrustal sequences. M, Manitoba promontory; Q, Québec promontory. Meta Incognita Peninsula (2014) and south-central Baffin (2015) field areas are outlined in teal.

Figure 3.2: Simplified geological map of the Canadian Shield and west Greenland, with the pale colours southwest of thick hashed line outlining the extension of the western Precambrian domains beneath Phanerozoic cover. The minimum extent of the area affected by a Trans–Hudson orogen age tectonothermal overprint (1.86–1.80 Ga) is delineated by the thick red hashed line. The limit of the Reindeer Zone is defined by the thick stippled blue line. Terrane names and major faults: BH, Buffalo Head terrane; BK, Black Hills; C, Chinchaga belt; CZ, Core Zone; FS, Fort Simpson terrane; FFB, Foxe Fold Belt; GB, Great Bear; GSL, Great Slave Lake shear zone; H, Hottah terrane; KS, Ksituan belt; L, Lacombe belt; N, Nahanni belt; NO, New Quebec; NV, Nova terrane; RB, Rimbey Belt; STZ, Snowbird Tectonic Zone; T, Thorsby basin; TA, Tasiuyak domain; W, Wabamun belt. Meta Incognita Peninsula (2014) and south-central Baffin (2015) field areas are outlined in teal. Modified from [Corrigan et al. \(2009\)](#).

Figure 3.3: Geologic terrane map of the Québec-Baffin segment of the Trans-Hudson Orogen (modified from [St-Onge et al., 2007](#)), showing the four tectonically stacked domains of southern Baffin Island and northern Québec and the suture zones that separate the domains. Meta Incognita Peninsula (2014) and south-central Baffin (2015) field areas are outlined in teal.

Figure 3.4: Tectonic summary diagram for the Québec-Baffin segment of the Trans-Hudson Orogen (from St-Onge et al., 2007), documenting time intervals for emplacement and accumulation of major plutonic and tectonostratigraphic units on southern Baffin Island and surrounding area. Shaded rectangles indicate Archean or Paleoproterozoic basement units, and the unfilled rectangles indicate Paleoproterozoic supracrustal units and magmatic suites. The horizontal shaded bars (D_0 - D_2) correspond to crustal sutures, identifying their age and extent, while M_{1a} - M_2 represent regional metamorphic events that occurred on southern Baffin Island. Abbreviations: BaS, Baffin suture; BeS, Bergeron suture; SRS, Soper River suture; M_{1a} , prograde granulite facies metamorphism; M_{1b} , high-T contact metamorphism; M_2 , retrograde amphibolite metamorphism.

Figure 3.5: Basic bedrock map showing dome-and-basin map patterns, which represent interference patterns formed during the D₄ upright folding that deformed all other structural elements; these fold axes generally trend east-northeast (St-Onge et al., 2015f). The D₄ generated dome-and-basin map patterns are best preserved in the southern and eastern portions of Meta Incognita Peninsula, Baffin Island, Nunavut. Map area is located in central Meta Incognita Peninsula, adjacent to the east side of the Narsajuaq arc terrane boundary. Coordinates at image centre are: Lat: 62°39'6.23"N; Long:68°6'30.80"W (coordinates obtained through Google Earth).

Figure 3.6a-d: **a)** Gossan in Lake Harbour Group psammite adjacent to a layered mafic-ultramafic sill, Meta Incognita Peninsula, Baffin Island, Nunavut; hammer is 35 cm long; **b)** Diopside-phlogopite-spinel-apatite-quartz calcareous grit, Lake Harbour Group, Meta Incognita Peninsula, Baffin Island, Nunavut; **c)** Shallowly-dipping, well-layered, garnet-bearing orthoquartzite, Lake Harbour Group, Meta Incognita Peninsula, Baffin Island, Nunavut; geologist for scale is 2 m tall; **d)** Garnet-sillimanite-biotite-melt psammite, Lake Harbour Group, Meta Incognita Peninsula, Baffin Island, Nunavut. From [St-Onge et al. \(2015a,f\)](#).

Figure 3.7a-b: **a)** Mafic-ultramafic sill (dark; from [St-Onge et al., 2015a](#)) emplaced in Lake Harbour Group quartzite (white) and psammite (rusty brown) and subsequently folded, Meta Incognita Peninsula, Baffin Island, Nunavut. Width of field of view is ~800 m. **b)** The Chidliak sill (15SAB-L075A–15SAB-L081A), Hall Peninsula, Baffin Island ([from Liikane et al., 2015](#)). Detailed sampling was conducted across this structurally overturned layered mafic-ultramafic sill in order to determine the petrogenesis of the Frobisher magmatic suite from a regional perspective. The sill comprises a basal metapyroxenite, overlain by metaperidotite, overlain by metagabbro and capped by metaleucogabbro. Width of field of view is ~500 m.

Figure 3.8: Bedrock geology map of Meta Incognita Peninsula, Baffin Island (with legend on following page), showing locations of the mafic, ultramafic and layered mafic-ultramafic sills. Intrusions sampled for geochemical and/or geochronological analysis are shown by their respective group's symbol, described in Chapter 7, geochemistry. Maps modified from [St-Onge et al. \(1999a-h; 2015b-e\)](#).

Figure 3.9: Bedrock geology map of northern Hall Peninsula, southern Baffin Island, showing locations of the mafic, ultramafic and layered mafic-ultramafic sills sampled. Intrusions sampled for geochemical and/or geochronological analysis are shown by their respective group's symbol, described in Chapter 7, geochemistry. Maps from [Blackadar \(1967\)](#) and [St-Onge et al \(2016a-j\)](#). Faded legend labels indicate that those specific geochemical groups were not recognized in that map area.

Figure 3.10: Concordant intrusive contact between ultramafic sill (14SAB-L003) and orthoquartzite host rock, Meta Incognita Peninsula, Baffin Island, Nunavut; hammer is 75 cm in length.

Figure 3.11: Basic bedrock map illustrating the morphology of sills from Meta Incognita Peninsula, Baffin Island, Nunavut; appearing elongate to ellipsoid or annular in shape. This is a result of the D_3 and D_4 folding that is responsible for generating interference patterns within the metasedimentary units of the Lake Harbour Group ([St-Onge et al., 2015f](#)). Coordinates at image centre are: Lat: 63°20'38.88"N; Long: 70°21'42.24"W (coordinates obtained through Google Earth).

Figure 3.12a-e: Geochronology diagrams for a leucogabbroic sample (14SAB-L017A) collected from a mafic-ultramafic sill on Meta Incognita Peninsula, Baffin Island, Nunavut: **a)** transmitted light photomicrograph of anhedral zircon grains from 14SAB-L017A; **b)** transmitted light photomicrograph of prismatic zircon grains from 14SAB-L017A; **c), d)** cathodoluminescence images of zircon grains prior to SHRIMP analysis. White scale bar is 100 μm ; **e)** Concordia diagram of U-Pb sensitive high-resolution ion microprobe (SHRIMP) results from 14SAB-L017A. Ellipses are plotted at the 2σ confidence level, colour was selected to match the coding of the traceelement signature of the sample, white dashed box delineates the outline of the isotope dilution–thermal ionization mass spectrometry (ID-TIMS) concordia diagram. Inset: Concordia diagram of IDTIMS results from 14SAB-L017A. Ellipses are plotted at the 2σ confidence level. After [Liikane et al., 2015](#).

Figure 3.13a-f: a-b) Layered metagabbro (15SAB-L080A), Hall Peninsula, Baffin Island (from Liikane et al., 2015). Rhythmic layering is a result of differences in the modal abundances in plagioclase, clinopyroxene and hornblende; hammer is 75 cm and 35cm in length on the left and right, respectively. **c-d)** Layered metaperidotite (15SAB-L077A), Hall Peninsula, Baffin Island (from Liikane et al., 2015). Rhythmic layering is a result of differences in the modal abundances of clinopyroxene, orthopyroxene and olivine; hammer is 35 cm in length. **e-f)** Basal metapyroxenite (15SAB-L076A), Hall Peninsula, Baffin Island (from Liikane et al., 2015); hammer is 75 cm in length.

Chapter 4 – Methods

4.1 Fieldwork and Sample Collection

This M.Sc. thesis was one of the activities undertaken as part of the Geo-mapping for Energy and Minerals (GEM) program, led by Natural Resources Canada (NRCan) in conjunction with the Nunavut Geoscience Office (CNGO), in an effort to complete the regional bedrock mapping for southern Baffin Island, Nunavut ([Weller, et al., 2015](#)). The primary objective of GEM is to “unlock the mineral and energy potential of the north” ([Natural Resources Canada, 2016](#)), and our contribution to this goal is the compilation and publication of thirteen new geological maps of southern Baffin Island, along with their lithostratigraphic units, boundaries and structures. The synthesis provided by the new geological maps and related research, including this study, has contributed to an increased understanding and better constrained tectonic history of Baffin Island.

Fieldwork was carried out over two consecutive field seasons during the summers of 2014 and 2015. The mapping and sample collection of Meta Incognita Peninsula took place between July 15th and August 15th, 2014, covering all or parts of NTS maps 025G, I, J, K, N, & O ([Liikane et al., 2015](#); [St-Onge et al., 2015](#)). The mapping and sample collection on the western Hall Peninsula took place between June 21st and August 18th, 2015, covering all or parts of NTS maps 026B, C, F, G, J, & K ([Liikane et al., 2015](#); [St-Onge et al., 2015](#)). Collectively, the bedrock mapping expeditions involved participants from Nunavut Arctic College, Aboriginal Affairs and Northern Development Canada, the University of Ottawa, Carleton University (myself) and Oxford University.

4.1.1 Regional Sampling

Samples of the Frobisher suite sills were collected throughout the two summers of fieldwork, via 10-15 km foot traverses and targeted sill-sampling excursions *via* helicopter. Over one hundred sills were documented over an area of ~130,000 km² (Figures 3.8 & 3.9), and 80 fist-size, or larger, representative samples collected. Almost all of the samples used in this study were collected in situ, and taken from outcrops exposing the freshest and least altered rock. In a few situations samples were collected from broken rock, due to poor or insufficient outcrop exposure. In such cases, field constraints (size and angularity of blocks, monolithological population) always indicated little transport. In the field, samples were individually wrapped in fabric bags, placed together in steel pails, and transported *via* cargo plane back to the Geological Survey of Canada (GSC) in Ottawa, Canada.

4.1.2 Detailed Sampling

Detailed sampling was conducted across two of the thicker mafic-ultramafic sills. The first is a 470 m-thick, tightly folded sill located near Kimmirut on the southern coast of central Meta Incognita Peninsula (Liikane et al., 2015), and is herein referred to as the Killapait sill (informally named after the nearby traditional Inuit place name; Figure 3.8). The second is a 230 m-thick, folded, boudinaged, and overturned sill from northern Hall Peninsula (referred to as the “northern” sill in Steenkamp et al., 2014 and Liikane et al., 2015), and is herein informally named the Chidliak sill (based on its proximity to Chidliak Bay).

4.2 Sample Selection and Preparation

Mafic and ultramafic samples that were collected during the two field seasons were sorted and those from mafic to ultramafic units, hosted in Lake Harbour Group metasedimentary rocks, were selected for this study (total of 80 samples). The collection includes samples from 31 sills across southern Baffin Island; 15 sills from Meta Incognita Peninsula (Figure 3.8), and 16 sills from western Hall Peninsula (Figure 3.9). Every sample processed in a rock preparation lab (both at Carleton University and the GSC), where a fist-sized portion was removed using a circular saw with a segmented diamond-encrusted (in soft metal) steel saw blade. The fist-sized sample was trimmed of weathered surfaces, ensuring scuff marks left from the blade were kept to a minimum. From each remaining representative sample, two slabs and a thin section puck were produced. The slabs were cut to a ~1 cm thickness using the circular saw. One slab was kept for the sample archive and the other was used to produce the thin section puck. The thin section pucks were cut using a trim saw to the approximate dimensions 27 mm x 46 mm.

The fist-sized trimmed samples were inserted into labelled plastic sample bags, neatly packed into steel pails, and sent away to ALS Chemex for whole-rock geochemical analysis. Sample bags were labelled twice and an additional label was placed in each bag as a precaution. The thin section pucks were marked with an “X” on one side, indicating the “bad” side or the side not to be made into a thin section. Thin section pucks were then placed into separate small

labelled envelopes, packed in a cardboard box (in alphanumerical order), and sent to Vancouver Petrographics to be made into polished thin sections.

Rock samples submitted to ALS Minerals' Laboratory in Val D'or, Quebec, for sample preparation were first subjected to fine crushing to 70% passing 2 mm, using steel jaw crushers. This finely crushed material was then divided up using a riffle splitter, producing representative sub-samples. Each single representative sub-sample (up to 250 g) was then pulverized to 85% passing 75 μm , using a "ring and puck" style grinding mill, either composed of low-chrome steel (PUL-31; used for samples collected in 2015) or ceramic (PUL-41; used for samples collected in 2014). For quality control, one out of every 50 samples crushed or pulverized was selected at random for routine pulp QC tests (CRU-QC; PUL-QC). For these routine pulps tests, 85% of the pulp must pass through a 75 μm screen, as per pulverization specifications. Following sample preparation, the pulps were sent out to ALS Minerals' Vancouver laboratory in British Columbia for lithogeochemical analysis. Following a preliminary evaluation of the bulk-rock geochemical data, a subset of these samples were selected for electron microprobe (EMP) mineral chemistry analyses, scanning electron microscopy (SEM) and Sm-Nd isotope geochemistry; all conducted at Carleton University.

4.3 Petrography

Thin section pucks were sent to Vancouver Petrographics where polished thin sections (30 μm) were produced, however, a subset (i.e. ~20) were either too

thick, or too thin. This caused some difficulty in mineral identification under XPL, but minerals and their textures could be identified through a combination of PPL and cautious XPL work. Polished thin sections were examined at Carleton University with a Nikon Optiphot Pol petrographic microscope under plane polarized, cross polarized, and reflected light (PPL, XPL, and RL, respectively) on 4x, 10x, and 20x magnification. PPL and XPL were used to document the silicate phases of the samples, and RL was used to observe oxide and sulphide mineral phases.

4.4 Photography and Microphotography

Microphotographs in PPL, XPL and RL were taken at Carleton University using a ProgRes SpeedXTcore5 petrographic microscope camera and software. Field photographs were captured during the 2014 and 2015 mapping programs by various members of the bedrock mapping team. These images were taken on Canon PowerShot G11 cameras on various camera settings. Photographs of thin section pucks and slabs were taken on a Canon compact camera.

High-resolution whole thin section scans were acquired from the GSC's Epson Perfection V700 Photo Scanner in both PPL and XPL at 2400 dots per inch (dpi). These images were printed in colour and bound, for reference while probing and completing petrography.

In conjunction with electron beam analyses on Carleton University's Cameca Camebax, digital BSE (back-scattered electron) images were collected in areas where EMPA mineral chemical analyses were obtained. These images

were collected with an Electron Optic Services digital imaging system at 512 x 512 pixel resolution with a Lamont 4 element solid state BSE detector and BSE Quad Summing Amplifier. BSE images were interfaced to a 4Pi Analysis Inc. digital imaging and EDX x-ray system, and Power Macintosh computer running NIH image and NIST desktop spectrum analyzer programs.

BSE imaging was also conducted on mounted zircon grains (discussed in section 3.5.4) prior to U-Pb geochronological analysis on the sensitive high resolution ion microprobe (SHRIMP) at the GSC. This form of imaging allowed for the internal features of the zircons to be examined (e.g. zoning, structures, alteration). The imaging was conducted on a Zeiss EVO® 50 scanning electron microscope (SEM) at the GSC.

4.5 Analytical Methods

Rock powders prepared by ALS Chemex were used for both whole-rock and Sm-Nd isotope geochemistry, and polished thin sections were used for mineral chemistry *via* EMPA. All data were processed using a combination of Microsoft Excel and Igpet (2012; by Terrasoft). The former was used for calculations and data organization, while both were used for plotting major- and trace-element geochemical data.

4.5.1 Whole Rock Geochemistry

Eighty samples were analysed at ALS Chemex in Vancouver, British Columbia, for bulk-rock geochemical analysis by X-ray fluorescence (XRF) and mass spectrometry techniques. A minimum of one standard (Certified Reference Material) was analyzed using each method described below, to determine the precision and accuracy of results.

Major oxide determinations were undertaken using XRF (code ME-XRF06), whereby each sample (0.9 g) was calcinated or ignited and then added to a 9.0 g solution of lithium borate flux (50% $\text{Li}_2\text{B}_4\text{O}_7$ - 50% LiBO_2 ; WEI-GRA06). This solution was mixed thoroughly and fused in an auto fluxer between 1050 - 1100°C and then quenched. The flat glass disc was then analyzed by XRF. This method involves the glass sample disc being analyzed by an XRF spectrometer, where an X-ray incident beam strikes the glass disc, exciting the atoms within the sample and causes them to release energy spectra specific to each element contained within the sample. The intensities of each individual spectrum are proportional to their elemental abundance within the sample. These intensities are converted to oxide wt.% by the instrument's software and reported. Table 4.1a contains a list of the major oxides determined, along with their units and lower detection limits. All iron was reported as Fe_2O_3 , but was converted to FeO, when necessary, using the conversion equation:

$$\text{FeO}^{\text{T}}\text{wt.}\% = \text{Fe}_2\text{O}_3^{\text{T}}\text{ wt.}\% / 1.1111 \text{ (weight ratio between } \text{Fe}_2\text{O}_3 \text{ and FeO)} \quad (3.1)$$

Loss on ignition (LOI) was obtained through method OA-GRA05, whereby 1.0 g of prepared sample was placed in a furnace at 1000°C for an hour, cooled and then weighed. The difference in weight gave the weight percent loss on ignition. Table 4.1a contains the units and lower detection limits for LOI analyses.

Data for the 30 trace element package, including the rare-earth elements (REEs), were obtained through Inductively Coupled Plasma - Mass Spectrometry (ICP-MS) method ME-MS81, which involved the addition of a prepared 0.200 g sample to 0.90 g of lithium metaborate flux. The solution was then mixed and fused in a furnace at 1000°C. Following fusion, the resulting melt was cooled, dissolved in a 100 mL solution of 4% HNO₃ + 2% HCl₃, and then analyzed by ICP-MS. In ICP-MS, the sample is subjected to argon plasma that dissociates the molecules and creates charged atoms. The ions then move through a magnetic field, which separates them based on mass, and then the ions strike the precisely positioned detection plates. Upon impact, their energy produces an electric pulse that is amplified and then the intensity is compared with a standard curve to determine its concentration. Table 4.1c contains a list of the trace elements analyzed, along with their units and lower detection limits. Ti concentrations in ppm were not determined with the rest of the trace elements, and were instead calculated from the TiO₂ wt.% concentrations obtained through XRF, using the equation:

$$\text{Ti (ppm)} = \text{TiO}_2 \text{ wt.\%} \times 0.5993\% \text{ (percent of Ti in TiO}_2\text{)} \times 10000 \quad (3.2)$$

Base metal values were obtained using the Inductively Coupled Plasma - Atomic Emission spectroscopy (ICP-AES) method ME-MS81, where each sample is dissolved using four acid digestion (ME-4ACD81) and analyzed on an ICP-AES. A four acid digestion is typically preferred when dealing with more resistive mineralization (i.e. those associated with Ni and Co). [Table 4.1d](#) contains a list of the base metals analyzed, along with their units and lower detection limits.

Total sulphur (S) and carbon (C) was measured using the Leco sulphur analyzer (S-IR08). In this method, the sample (0.01 - 0.1 g) is heated in an induction furnace to a temperature of approximately 1350°C, while passing a stream of oxygen through the sample. The IR detection system then measures the sulphur dioxide released from the sample and provides the total sulphur value. [Table 4.1b](#) contains a list of the units and lower detection limits for S and C analyses.

Platinum (Pt), palladium (Pd) and gold (Au) analysis were obtained using standard lead oxide collection fire assay method PGM-ICP23 (for rocks with intermediate levels of precious metal mineralization), finishing with ICP-AES. In this method, 30 g of sample is fused with a mixture of lead oxide, sodium carbonate, borax and silica, inquarted (i.e. gold content in alloy was reduced to 25%) with 6 mg of gold-free silver and then cupelled to yield a precious metal bead. The bead was then digested at high power by microwave in dilute nitric acid. The solution is cooled, hydrochloric acid is added and then the solution digested at half power by microwave. The digested solution was then cooled,

diluted with hydrochloric acid, homogenized, and then analyzed for Au, Pt and Pd by ICP-MS analysis. Table 4.1e contains a list of the units and lower detection limits for Pt, Pd, and Au analyses.

Each of ALS Minerals' laboratories in North America are registered to ISO 9001:2008 for the "provision of assay and geochemical analytical services" by QMI Quality Registrars. The North America analytical laboratories are also accredited by the Standards Council of Canada (SCC) for specific tests listed in their Scope of Accreditation No. 579, http://palcan.scc.c-a/specs/pdf/677_e.pdf. This accreditation is based on ISO 17025:2005 international standards and involves extensive site audits and performance evaluations. Detailed descriptions of all the analytical methods can be found on ALS Global's website through the search bar using the method code (e.g. PGM-ICP23).

A total of six different standards were used throughout the two batches of geochemical analyses (2014 and 2015 datasets): WGB-1, TDB-1, UMT-1, WPR-1, WMG-1 and 10-LT-05. These samples were given southern Baffin sample code names from 14SAB-Z001 to 14SAB-Z008 and from 15SAB-Z001 to 15SAB-Z006. All standards, except for 10-LT-05, are certified reference materials from the Canadian Certified Reference Materials Project (CCRMP).

All standards show that the geochemical data for major and trace elements produced by ALS Chemex are accurate, as they generally fall within a 95% confidence interval of the certified (and provisional) values outlined by the Canadian Certified Reference Materials Project (CCRMP; see Figures 4.1, 4.2, 4.3). Due to the large sample size required for fire assay (minimum of 30 g),

there was insufficient standard powder to analyze for Au, Pt, and Pd. Consequently, the accuracy and precision of fire assay results could not be verified. All data from standards and submitted samples can be found in Appendices I-III.

4.5.2 Mineral Geochemistry

Quantitative mineral analyses were obtained on an automated 4 spectrometer Cameca Camebax MBX electron probe (EMP) at the Earth Sciences Department of Carleton University, under the supervision of Peter Jones. The group of samples (n=57) selected for EMP analyses were the least altered samples from mafic, ultramafic, and layered mafic-ultramafic sills from both field areas, spanning every significant geochemical group, and providing sufficient geographical coverage of the two field areas. Mineral compositional data were also obtained from the mafic-ultramafic Chidliak and Killapait sills. Every mineral phase in each sample analyzed was probed at least once, with the exception of retrograde metamorphic mineral phases (i.e. serpentine, epidote, sericite).

The polished thin sections of samples selected for mineral analyses were evaluated under a Nikon petrographic microscope and areas ~1 mm in diameter were marked and photographed using the circular petrographic marker and petrographic microscope camera, respectively. These circular areas outlined the locations of subsequent mineral analyses on the EMP.

Prior to analysis, the polished thin sections were coated with a thin film of carbon in a high-vacuum carbon coater. The carbon-coated thin sections were then placed in a vacuum sample chamber within the EMP and selected mineral phases were bombarded with an accelerated and focused electron beam. When the electron beam strikes the surface of the mineral being analyzed, the incident electrons collide inelastically with electrons in the inner shells of atoms within the mineral. The inner-shell electrons are ejected, creating a vacancy, and electrons from a higher orbital shell drop down into their place, shedding energy in the form of an X-ray; each quantized X-ray energy is characteristic of a particular element. The raw X-ray data is then converted to elemental weight % by the Cameca PAP matrix correction program. Additional calculations undertaken produce cation proportions and mineral formulae.

Operating conditions for the wavelength dispersive x-ray analysis method (WDX) were: focused (2 μm diameter) beam; 20 kv accelerating potential and a beam current of 20 nano-amperes (nA) for silicates and oxides, and 20 kv and 30 nA for sulfide minerals. Beam-sensitive minerals such as feldspar and mica were analyzed using a rastered electron beam of from 5x5 to 10x10 microns in size. Peak-counting times for analyzed elements were: 15-40 seconds or 40,000 accumulated counts. Background positions were chosen carefully to avoid interferences from adjacent elemental peaks. Background measurements were made at 50% peak counting time on each side of the analyzed peak. Analyses are accurate to 1-2 % relative for major elements (>10 wt %) and 3-10 % relative

for minor elements (<10 wt% to >0.5wt %). As detection limit is approached (<0.1 wt %), relative errors approach 100%.

A group of well-characterized natural and synthetic minerals and compounds were used as calibration standards for EMP mineralogical analysis. The elements calibrated prior to analyses are listed in Table 4.2, along with the X-ray lines and mineral standards used.

4.5.3 Isotope Geochemistry

Following the preliminary characterization of the Frobisher suite using bulk-rock geochemistry, a batch of 12 samples was selected for Sm-Nd isotope geochemical analysis. Samples were selected such that all eight geochemical groups (groups defined in Chapter 7) were covered, and samples were chosen from mafic, ultramafic, and mafic-ultramafic sills in both field areas; including the one that had been dated. A subsequent batch of 12 samples was selected based on the results from the first batch of analyses obtained. Preparation and analyses were completed at the Isotope Geochemistry and Geochronology Research Centre (IGGRC) at Carleton University. The Sm-Nd isotope data was acquired via a 9-cup ThermoFinnigan Triton TI multicollector thermal ionization mass spectrometer (MC-TIMS).

Silicate rock powders, were weighed, transferred into individual 15 mL Teflon screw-cap vials, and spiked with a mixed ^{148}Nd - ^{149}Sm spike. The need for a spike is due to the Paleoproterozoic age of the samples, and the Nd and Sm concentrations cannot be precisely determined without having added a known

concentration of ^{148}Nd - ^{149}Sm spike. The precise sample powder weight (between 30 and 200 mg) and spike weight for each sample were calculated on an Excel spreadsheet based on the measured Nd concentration (in ppm) obtained from the ICP-MS analysis conducted by ALS Chemex. The spiked sample powders were dissolved in an acidic solution (2-3 mL) of ~29 M HF and ~16 M HNO_3 . The samples were then dried down on the hotplate before being re-dissolved with 8 M HNO_3 and 6 M HCl, sequentially. The dried residue of samples were dissolved in 2.5 M HCl and subsequently loaded into 14-mL Bio-Rad borosilicate glass chromatography columns containing 3.0 mL of Dowex AG50W-X8 cation resin. Columns were then washed with 23 mL of 2.5 M HCl before the rare earth elements (REE) were eluted using 9 mL of 6 M HCl. The REE fractions were dissolved in 0.26 M HCl and then loaded onto Eichrom Ln Resin chromatographic columns containing Teflon powder coated with HDEHP [di(2-ethylhexyl) orthophosphoric acid; [Richard et al., 1976](#)). Nd was eluted using 4.5 mL of 0.26 M HCl, followed by eluting of Sm with 4 mL of 0.5 M HCl. Both Sm and Nd vials were then dried down.

The samples were loaded with 2 μL of H_3PO_4 in a double rhenium filament assembly. Samples with a higher Nd concentration were initially dissolved in 4 μL of H_3PO_4 prior to loading, leaving 2 μL behind in the vial for use as a re-run in the event of a failed initial run. The sample solutions were then partially dried by progressively increasing the current passing through the filament from 0.0 to 1.9 amperes (A), in 0.2 A stages. As the filaments heated up, they began to smoke (at ~1.7 A). After ~30 s of smoking, the current was cut and the sample and

ionization filaments were loaded on a 21-position sample wheel. The sample and ionization filaments were then carefully aligned and a metal cover was clipped on top of them. Finally, the sample wheel was loaded into the TIMS sample chamber.

The isotope ratios were measured using the ThermoFinnigan Triton TI MC-TIMS. The MC-TIMS operates by utilizing the thermal ionization effect, whereby the sample is heated to a point where the atoms ionize. These ions are then focused by electrostatic lenses, and the resulting beam passes through a magnet field which splits the ions into a number of beams based on their mass/charge ratios. The mass-resolved beam is directed into collectors, which convert the ion beam into voltage. Through the comparison of voltages corresponding to specific ions, precise isotope ratios can be determined.

Ions were ionized at temperatures between 1600 and 1800°C. Isotope ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$. IGGRC's in-house Nd standard was routinely measured with $^{143}\text{Nd}/^{144}\text{Nd} = 0.511831 \pm 0.000006$ (1σ) over a period of three years; this value is equivalent to $^{143}\text{Nd}/^{144}\text{Nd} = 0.511860$ for the La Jolla Nd standard. Sm and Nd concentrations were measured precisely within 1%, while $^{147}\text{Sm}/^{144}\text{Nd}$ ratios are reproducible to 0.5%. Analyses of the USGS standard BCR-2 yield Nd = 28.59 ppm, Sm = 6.647 ppm, and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512644 \pm 0.000006$ (1σ , $n=4$), consistent with accepted values. Total procedural blanks for Nd were consistently less than 50 picograms. Isotopic data is located in Appendix II.

Additional data including Sm and Nd concentrations, epsilon Nd values, initial Nd isotopic ratios, and depleted mantle model ages were calculated using an Excel program written by Drs. George Tilton and Brian Cousens of Carleton University. All samples were corrected for post-crystallization decay using the minimum crystallization age of 1900 Ma; an age obtained through the U-Pb geochronological dating of a leuco-gabbroic sample from the Killapait layered mafic-ultramafic sill (Liikane et al., 2015). Initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were calculated by the rearranging of equation 3.1.

$$\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{sample}} = \left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{initial}} + \left(\frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_{\text{sample}} \times (e^{\lambda t} - 1) \quad (3.1)$$

where: $\lambda = 6.54 \times 10^{-12} \text{y}^{-1}$

Epsilon-Nd values at time T were calculated using equation 3.2, and indicate whether rocks were derived from either a depleted or enriched mantle source (i.e. depleted or enriched in ^{143}Nd relative to CHUR, respectively; Faure and Mensing, 2005). The $^{143}\text{Nd}/^{144}\text{Nd}$ values were corrected for fractionation using $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$.

$$\epsilon_{\text{Nd,CHUR}} = \left[\frac{\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{sample}}}{\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{CHUR}}} - 1 \right] \times 10^4 \quad (3.2)$$

where: CHUR = Chondrite Uniform Reservoir

T = time of formation of the rock

The DePaolo depleted mantle model ages were calculated using equation 3.3, and the assumed upper mantle $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of 0.214 and $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.513115.

$$T_{DM}^{Nd} = \frac{1}{\lambda} \ln \left[\frac{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample, today}} - (^{143}\text{Nd}/^{144}\text{Nd})_{DM, \text{today}}}{(^{147}\text{Sm}/^{144}\text{Nd})_{\text{sample, today}} - (^{147}\text{Sm}/^{144}\text{Nd})_{DM, \text{today}}} + 1 \right] \quad (3.3)$$

Quality data was obtained from all samples except for one (15SAB-L076A01). The sample 15SAB-L076A01 failed to produce accurate data due to an insufficient amount of Nd on the filament; the Nd was exhausted during heating, before analysis could commence. This is believed to have resulted from a combination of issues. The first issue is that this particular sample has a very low Nd concentration (1.2 ppm). The second issue is that during sample dissolution in hydrofluoric acid (HF) the formation of calcium fluoride (CaF_2), which is known to accommodate Nd in its structure, likely left the remaining solution depleted in Nd. The result was a sample with too low a concentration to allow for proper analysis.

4.6 Potential Error in Geochemical Analyses

Geochemical error is an important issue when it comes to evaluating the accuracy of geochemical data. If the proper precautions aren't taken to minimize potential causes of error, inaccurate data may be produced and this may further lead to false results and interpretations of a geochemical dataset. Potential sources of error commonly include equipment error, human error (i.e. mislabelling samples, sample contamination, etc.), sample heterogeneity, redistribution of chemical components during metamorphism/alteration, etc., and can lead to misidentification, misinterpretation, and anomalous values. All of these sources of error can occur beginning at the time of sample collection.

In the field, there exists the possibility that the samples collected may not be geochemically representative of the sills from which they came. Even while textural differences throughout a sill are visually evident, other geochemical variations elusive to the naked eye often exist in layered intrusions; namely the presence of cryptic layering (e.g. the Eastern Lobe of the Bushveld Complex of South Africa; [Wager and Brown, 1968](#)). Unfortunately, there is no way to identify these variations in the field without a detailed geochemical profile of an intrusion. In light of this, much care was taken throughout the sampling process, but one cannot be fully certain of the homogeneity of any sill sampled under these conditions. Based on the “well-behaved” trace element patterns within each geochemical group, it appears that sampling error was not a significant issue. This may partial be to the consistently large samples that were collected where/when possible.

Prior to geochemical analysis, samples were divided into subsets numerous times as they progressed through the sample preparation process. After every spilt, there is a possibility that the specific subset of sample material chosen to move to the next stage of sample preparation (and eventually on to geochemical analysis) may not be representative of the original sample collected. Within this sample set specifically, the greatest causes of potential heterogeneity amongst the powdered sample sub-sets are coarse grain-sizes, relict phenocrysts, and mm- to cm-scale layering within samples. Although unlikely, there was also ample opportunity for samples to be mislabelled as they moved from one form (or vessel) to another during preparation.

Sample contamination is a major source of error in geochemical data. It can occur at any stage, and can occur simply by the sample coming into contact with another surface containing a chemical component of interest (i.e. carbon, iron, chromium, etc.). The main sources of these contaminants are those instruments used in sample collection and preparation, including sledgehammers, saw blades, jaw crushers and “puck-and-ring” style grinding mills. The chrome-steel ring mill used in the 2014 samples does not appear to have affected samples, as suggested by the lack of anomalous Cr values in plots in the geochemistry chapter (Chapter 7). The specific saw blade currently being used at the GSC, where these samples were cut and trimmed, contains “some or all of: Al_2O_3 , Co, Fe, Ni, SiC, W, B_4C , Cu, Mn, P, Ag, WC, Cr, C, Mo, steel” (pers. comm. [J. Pinard, 2017](#)). If any of these components were added to the sample, the effects of these components on the geochemical data cannot be recognized,

as there was no standards subjected to the saw blade (all standards were already in powdered form). However, these saw blades have been used in sample preparation for decades without any significant issues arising with inaccurate data.

During sample preparation, other potential sources of contamination include improper cleaning of equipment, rock dust/powder on preparation surfaces or pails. In the process of collecting REEs, Sm, and Nd for Sm-Nd for isotope work, contaminants can be introduced from vials or Teflon columns in the form of ions or sample residues. However, this is only likely to occur if lab procedures and methods deviate from accepted practice, or standard cleaning and preparation tasks are overlooked.

All along the way, every effort was made to avoid the possibility of sample contamination prior to submitting the samples to ALS Chemex for crushing, pulverizing and geochemical analysis. Representative samples collected from the field were trimmed of all weathered surfaces and veining. Additionally, large sample-sizes were submitted in the majority of cases, to offset the potential heterogeneity arising from coarser grain sizes.

Instrumental error is also a potential source of invalid data. However, this is typically easier to track and identify versus the contamination that occurs during preparation. However, slight unrecognized changes or fluctuations in machine settings may have occurred during XRF, ICP-MS, and ICP, AES geochemical analyses. However, data obtained from standards indicate that the

data are accurate. Samples analyzed for Sm-Nd isotopes were also run with standards to assess the functionality of equipment and result validity.

References

- Faure G. and Mensing T.M. (2005). *Isotopes. Principles and Applications*, 3rd Edition. John Wiley and Sons, New York. ISBN 0-471-38437-2.
- Liikane, D.A., St-Onge, M.R., Kjarsgaard, B.A., Rayner, N.M., Ernst, R.E. and Kastek, N. (2015). Frobisher suite mafic, ultramafic and layered mafic-ultramafic sills, southern Baffin Island, Nunavut; in *Summary of Activities 2015*, Canada-Nunavut Geoscience Office, p. 21–32.
- Richard, P., Shimizu, N. and Allègre, C.J. (1976). $^{143}\text{Nd}/^{146}\text{Nd}$, a natural tracer: an application to oceanic basalts, *Earth Planet. Sci. Lett.*, 31, 269-278.
- Stacey, J.S. and Kramers, J.D. (1975). Approximation of Terrestrial Lead Isotope Evolution by a 2-Stage Model. *Earth and Planetary Science Letters* 26(2): 207-221.
- St-Onge, M.R., Rayner, N.M., Steenkamp, H.M. and Skipton, D.R. 2015b: Bedrock mapping of eastern Meta Incognita Peninsula, southern Baffin Island, Nunavut; in *Summary of Activities 2014*, Canada-Nunavut Geoscience Office, p. 105–118.
- Wager, L.R. & Brown, G.M., 1967. *Layered igneous rocks*. xv + 588 p., 278 figs., and 13 pls. (half-tone and coloured). Oliver and Boyd, Ltd., Edinburgh.
- Weller, O.M., Dyck, B.J., St-Onge, M.R., Rayner, N.M. and Tschirhart, V. 2015: Completing the bedrock mapping of southern Baffin Island, Nunavut: plutonic suites and regional stratigraphy; in *Summary of Activities 2015*, Canada-Nunavut Geoscience Office, p. 33–48.

Figure 4.1a-c: Known-value-normalized major oxide data obtained from ALS Chemex for the gabbro standards WGB-1 (a) and WMG-1 (b) in the 2014 dataset, and for WGB-1 (c) in the 2015 dataset. Analyses compared to certified reference values are bold and in a larger font, while analyses compared to provisional reference values are italicized and in a smaller font.

Figure 4.2: Known-value-normalized trace element data obtained from ALS Chemex for the diabase standard TDB-1 in the 2014 dataset. Analyses compared to certified reference values are bold and in a larger font, while analyses compared to provisional reference values are italicized and in a smaller font.

Figure 4.3a-b: Known-value-normalized base metal data obtained from ALS Chemex for the diabase standard TDB-1 in the 2014 (a) and 2015 (b) datasets. Analyses compared to certified reference values are bold and in a larger font, while analyses compared to provisional reference values are italicized and in a smaller font.

Table 4.1a-e: List of analytes, their lower detection limits and units: a) Major oxides were digested in a lithium borate flux solution and analyzed *via* X-Ray Fluorescence (XRF). LOI was determined using a thermal decomposition furnace and measured gravimetrically; b) Sulphur (S) and carbon (C) were analyzed *via* the Leco Method in an induction furnace and measured gravimetrically; c) Trace elements were digested through lithium metaborate fusion and analyzed *via* Inductively Coupled Plasma - Mass Spectroscopy (ICP-MS); d) Base metals were digested using four acid digestion and analyzed *via* Inductively Coupled Plasma - Atomic Emission spectroscopy (ICP-AES); e) Precious metals (except silver) were determined using standard lead oxide collection fire assay, and finishing with ICP-AES.

a) Major Oxides			c) Traces Elements (including REEs)			d) Base Metals		
Analyte	Detection limit	Unit	Analyte	Detection limit	Unit	Analyte	Detection limit	Unit
Al ₂ O ₃	0.01	%	Ba	0.5	ppm	Ag	0.5	ppm
BaO	0.01	%	Ce	0.5	ppm	As	5	ppm
CaO	0.01	%	Cr	10	ppm	Cd	0.5	ppm
Cr ₂ O ₃	0.01	%	Cs	0.01	ppm	Co	1	ppm
Fe ₂ O ₃	0.01	%	Dy	0.05	ppm	Cu	1	ppm
K ₂ O	0.01	%	Er	0.03	ppm	Li	1	ppm
MgO	0.01	%	Eu	0.03	ppm	Mo	1	ppm
MnO	0.01	%	Ga	0.1	ppm	Ni	1	ppm
Na ₂ O	0.01	%	Gd	0.05	ppm	Pb	1	ppm
P ₂ O ₅	0.01	%	Hf	0.2	ppm	Sc	1	ppm
SiO ₂	0.01	%	Ho	0.01	ppm	Tl	1	ppm
SrO	0.01	%	La	0.5	ppm	Zn	2	ppm
TiO ₂	0.01	%	Lu	0.01	ppm	e) Precious Metals (with Ag)		
LOI	0.01	%	Nb	0.2	ppm	Analyte	Detection limit	Unit
Total	0.01	%	Nd	0.1	ppm	Pt	0.005	ppm
b) Total Sulphur and Carbon			Pr	0.03	ppm	Pd	0.001	ppm
Analyte	Detection limit	Unit	Rb	0.2	ppm	Au	0.001	ppm
S	0.01	%	Sm	0.03	ppm			
C	0.01	%	Sn	1	ppm			
			Sr	0.1	ppm			
			Ta	0.1	ppm			
			Tb	0.01	ppm			
			Th	0.05	ppm			
			Tm	0.01	ppm			
			U	0.05	ppm			
			V	5	ppm			
			W	1	ppm			
			Y	0.5	ppm			
			Yb	0.03	ppm			
			Zr	2	ppm			

Table 4.2: Elements, X-ray lines and standards used in electron

microprobe (EMP) standard calibration.

Silicate Standards			Sulphide Standards		
Element	Line	Standard	Element	Line	Standard
Si	K α	olivine	As	L α	NiAs
Al	K α	spinel (synthetic)	S	K α	pyrite
Mg	K α	olivine	Ni	K α	NiAs
Na	K α	albite	Fe	K α	pyrite
K	K α	microcline (USNM)	Cu	K α	chalcopyrite
Ca	K α	wollastonite	Mn	K α	MnS (synthetic)
Ti	K α	MnTiQ (synthetic)	Co	K α	Co
Cr	K α	C ϵ O $_3$ (synthetic)	Cd	L α	CdS
Mn	K α	MnTiQ (synthetic)	ZnS	K α	ZnS (synthetic)
Fe	K α	Fayalite (synthetic)			
Ni	K α	NiO (synthetic)			
Ba	L α	barite			
S	K α	barite			
Cl	K α	tugtupite			
F	K α	lithium fluoride (synthetic)			
Zn	K α	gahnite (USNM)			

Chapter 5 – Petrography

5.1 General Observations

Geochemical groups mentioned below are defined on the basis of whole-rock geochemistry (Chapter 7).

5.1.1 Mineral assemblages

Mineral assemblages found within units spanning all geochemical groups, which are defined in Chapter 7, are quite variable in terms of constituent phases, modal percentages and grain size. However, the mafic to ultramafic mineral assemblages are generally characteristic of amphibolite to granulite facies metamorphic conditions (see Chapter 2).

The majority of the mafic samples collected contain the assemblage plagioclase + hornblende + orthopyroxene + clinopyroxene, with a lesser number containing the assemblage plagioclase + hornblende. A significant number also contain the assemblage plagioclase + hornblende + clinopyroxene, or plagioclase + hornblende + orthopyroxene. Other samples contain a combination of hornblende, olivine, clinopyroxene, orthopyroxene and spinel, in variable abundances. Plagioclase is present in a subset of ultramafic samples, up to 10 modal%.

The metamorphic mineral assemblages replace the primary magmatic minerals and consequently few primary magmatic textures are preserved. However, in some cases, relict magmatic textures (e.g. porphyritic texture, relict magmatic layering) are preserved and thus provide information on the petrogenesis of this sills.

Rock samples were named on the basis of their modal mineral content (Figure 5.1) through the use of ternary diagrams adapted from the IUGS classification and nomenclature of igneous rocks (Le Bas & Streckeisen, 1991) to reflect the igneous protolith, despite the metamorphic origin of mineral assemblages. This method of classification was chosen in order to provide the maximum amount of information about the sample's modal mineralogical composition. Standard metamorphic naming schemes yield names that are too broad and generalized (e.g. granulite, amphibolite, gneiss, etc.) to highlight the significant variation that exists in the suite of mafic and ultramafic samples. Thus, these rocks are secondary, recrystallized gabbros, peridotites, pyroxenites and hornblendites.

5.1.2 Textures

Sample textures vary from well-layered (Figure 5.2b) to massive (Figure 5.3c). Massive samples show significant textural variability, with samples exhibiting both homogeneous and heterogeneous mineral grain distribution. Samples may be equigranular (e.g. granoblastic), or inequigranular [e.g. seriate (Figure 5.3a), porphyritic (Figure 5.2a)] in character. Some samples appear to have been annealed, as exhibited by the granoblastic textures and 120-degree triple junction grain boundaries. However, many samples have retained strain textures as amphibole-defined foliations. The granoblastic textures and 120-degree triple junction grain boundaries are suggestive of slow cooling at high temperatures, such as during regional metamorphism. In granoblastic samples,

grain boundaries are generally sharp and straight, however, rounded grain boundaries occur as well. Porphyritic samples contain megacrysts of either orthopyroxene (Figure 5.3a), olivine (Figure 5.2c), or plagioclase (Figure 5.2a). Matrix grains are often granular or polygonal, with crystal shapes dictated by close packing, rather than crystal structures. However, amphibole defines a weakly to moderately developed foliation fabric in many samples (Figure 5.3c). A very limited number of samples exhibit reaction textures (Figure 5.4a-b), and many samples contain pyroxene or amphibole grains that exhibit exsolution laminae (Figure 5.4c). Additionally, a relict cumulate texture (Figure 5.3b) can be observed in a few samples also. The average matrix mineral grain size is 1.0 mm, with the full range of grain size generally ranging between 0.3 and 3.0 mm.

5.1.3 Metamorphism

It is difficult to attribute many of the mineral assemblages to a specific metamorphic event (M_{1a} or M_2), given that Hbl + Pl occur well into the granulite facies in mafic rocks. Consequently, it is not clear whether these assemblages formed during the initial prograde granulite facies metamorphism, or whether they are part of the retrograde overprinting assemblage formed during the later upper amphibolite metamorphic event. However, the assemblages observed in metamorphosed ultramafic samples suggest that many sills preserved granulite facies (M_{1a}) metamorphic assemblages, highlighted by the coexistence of clinopyroxene, orthopyroxene, and spinel. The presence of spinel, as opposed to garnet or plagioclase, indicates an intermediate pressure of metamorphism.

Effects of greenschist-facies retrogression are apparent in select samples, where serpentine or sericite have partially broken down olivine or plagioclase, respectively. Overall, these observations are consistent with conclusions made through P-T studies on the host metasedimentary units of the Lake Harbour Group by [St-Onge et al. \(2007\)](#); see Chapter 3).

5.2 Mineralogical Characteristics of Geochemical Groups

With respect to mineral abundance descriptions, the percentages used are modal volume percentages. In the text below, the term “n” represents the number of samples studied, and “s” represents the number of sills with a mafic or ultramafic component that are part of a given geochemical group. Geochemical groups are described in detail later in Chapter 7 (Geochemistry). A comparison of the mineral contents in each of the geochemical groups is found in [Table 5.1](#).

5.2.1 Group A1

Group A1 (n=6) is limited to ultramafic units sampled from ultramafic (s=3) and mafic-ultramafic (s=2) sills. Mineral assemblages are generally dominated by hornblende (39-80%) and olivine (15-60%), with some samples containing orthopyroxene and spinel, up to 25% and 1%, respectively. Plagioclase and clinopyroxene are absent in samples from this group. Magnetite+ilmenite contents are <1%, and all samples contain trace amounts of sulphide. Sulphides are dominantly pyrrhotite and chalcopyrite. There is one occurrence of pentlandite, and two contain only pyrite.

5.2.2 Group A2

Group A2 (n=17) comprises mafic and ultramafic units from mafic (s=7), ultramafic (s=1), and mafic-ultramafic (s=3) sills. Ultramafic samples (n=3) are dominantly composed of hornblende (20-70%) and olivine (14-70%), with orthopyroxene up to 16%. One ultramafic unit contains 4% spinel. Trace amounts of magnetite+ilmenite are also present, and the majority of samples contain trace sulphides. Mafic samples (n=14) from Group A2 are primarily composed of hornblende (2-70%) and plagioclase (20-50%), and >60% of samples contain clinopyroxene and orthopyroxene in variable abundances of 10-53% and 5-25%, respectively. A few samples collected contain up to 4% biotite, and one sample (15SAB-L056B01) contains a garnet porphyroblast. Most samples contain trace amounts of magnetite+ilmenite and sulphides, with the latter dominantly comprising pyrrhotite or pyrite, both with lesser chalcopyrite and one occurrence of pentlandite. Four samples are sulphide-barren.

5.2.3 Group B

Group B (n=5) comprises mafic and ultramafic units from mafic (s=1) and mafic-ultramafic (s=2) sills. The single ultramafic sample (15SAB-L113B01) contains mostly hornblende (84%) with minor amounts of clinopyroxene (8%) and orthopyroxene (8%). Trace amounts of sulphide, but no magnetite+ilmenite are present. The mafic samples (n=4) are plagioclase-rich, and contain variable proportions of hornblende (15-30%), clinopyroxene (15-25%) and orthopyroxene

(0-20%). One of the samples (15SAB-D196B01) contains 5% olivine. The samples generally contain trace amounts of sulphide and trace to 4% magnetite+ilmenite. One sample (15SAB-M005A01) contains 2% sulphide. Sulphides are dominantly pyrrhotite with lesser chalcopyrite with two samples containing pentlandite.

5.2.4 Group C1

Group C1 (n=8) comprises mafic and ultramafic samples from mafic (s=1), ultramafic (s=1) and mafic-ultramafic (s=1) sills. Ultramafic samples (n=5) are hornblende-rich (20-88%), and contain variable amounts of clinopyroxene (0-35%), orthopyroxene (0-28%) and olivine (0-40%). Samples contain 0 to 10% spinel, and one sample contains 5% plagioclase (15SAB-F013B01). Most samples contain trace sulphides, and some contain trace magnetite+ilmenite. Mafic samples (n=3) contain approximately equal proportions of plagioclase (40-55%) and hornblende (45-55%), and are devoid of clinopyroxene, orthopyroxene or olivine. These samples are relatively enriched in sulphides (1 and 5% sulphide) compared to other sample groups, and have magnetite+ilmenite contents up to 4%. Sulphides are dominantly either pyrrhotite with minor chalcopyrite and pyrite, or pyrite with minor chalcopyrite. The majority of the Group C1 samples were collected during a detailed sampling traverse across the Killapait sill, located on the south coast of central Meta Incognita Peninsula.

5.2.5 Group C2

Group C2 (n=12) comprises mafic and ultramafic samples from mafic (s=1) and mafic-ultramafic (s=2) sills. The ultramafic samples (n=3) contain olivine (55-70%), hornblende (20-35%), and up to 5% spinel. One sample contains 5% orthopyroxene (14SAB-L014A01). The samples also contain trace sulphides and magnetite+ilmenite. The mafic samples contain plagioclase (13-96%) and hornblende (4-50%), with a few also featuring orthopyroxene (0-20%), olivine (0-30%), biotite (0-5%), and spinel (0-1%). Clinopyroxene (30%) was only identified in one sample (14SAB-L011A02). Most samples contain at least trace sulphides, while magnetite+ilmenite were only identified in a couple. A large number of samples contain only pyrite, while the others contain pyrrhotite and/or pyrite with chalcopyrite, with the latter in lesser amounts. The majority of these samples were collected during a detailed sampling traverse across the Killapait sill (the same sill that hosts the bulk of samples of Group C2).

5.2.6 Group D

Group D (n=17) comprises mafic and ultramafic samples from mafic (s=1), ultramafic (s=5) and mafic-ultramafic (s=2) sills. The ultramafic samples (n=11) contain hornblende (23-80%), and variable amounts of orthopyroxene (0-52%) and olivine (0-40%). Three ultramafic samples contain between 15 and 50% clinopyroxene, two contain 5-15% biotite, and another two contain 1-2% spinel. Almost all ultramafic samples contain at least trace amounts of sulphide, with four containing 1-10% sulphide (14SAB-L004A01; 14SAB-L009A01; 14SAB-

L009A02; 14SAB-L009B02), and typically ~1% magnetite+ilmenite. Mafic samples (n=6) from Group D are composed mostly of hornblende (30-65%), a narrow range of clinopyroxene (20-30%), and variable plagioclase (4-34%). The sample from this group that is the most mineralized (10% sulphide) contains 48% plagioclase, 39% clinopyroxene and 3% spinel. All other mafic samples contain only trace amounts of sulphide and up to 2% magnetite+ilmenite, when present. Sulphides include pyrrhotite and/or pyrite with minor chalcopyrite. Some samples contain variable amounts of pentlandite as well.

5.2.7 Group E

Group E (n=4) comprises mafic samples from mafic (s=2) sills. The samples contain variable amounts of plagioclase (28-30%), hornblende (24-32%), clinopyroxene (18-20%) and orthopyroxene (5-23%), and some samples contain trace levels of sulphides and magnetite+ilmenite. One mineralized sample (15SAB-M003A01) from this group contains 70% plagioclase, 23% orthopyroxene, 5% sulphides and 2% magnetite+ilmenite. Two samples contain pyrrhotite and/or pyrite with chalcopyrite, and one sample contains pyrite as the only sulphide phase.

5.2.8 Group F

Group F (n=11) comprises mafic and ultramafic samples from ultramafic (s=1) and mafic-ultramafic (s=1) sills. Ultramafic samples (n=8) contain hornblende (50-90%) and variable amounts of orthopyroxene (0-40%), olivine

(0-46%) and spinel (0-7%). The two pyroxene-free ultramafic samples each contain 10% plagioclase. Mafic samples (n=3) contain hornblende (45-60%) and plagioclase (40-45%). In addition, one sample contains 10% orthopyroxene. Most samples contain trace sulphides, are generally devoid of magnetite+ilmenite, with one exception containing ilmenite. Sulphides within these samples include variable amounts of pyrrhotite, pentlandite and pyrite. Most of the samples were collected during a detailed sampling traverse across the Chidliak sill, located in southernmost portion of the field area on Hall Peninsula.

5.3 Silicate Mineralogy

Given that this is a reconnaissance study, the occurrences, abundances and characteristics of the main mineral constituents are the focus of the petrographic portion of the study, and therefore, the minerals present in trace abundances throughout the Frobisher suite are not described.

5.3.1 Olivine

Olivine typically occurs as anhedral grains that are transected by serpentine-filled interconnecting arcuate fractures, with the degree of serpentinization varying from weak to extreme. In most areas where olivine is completely serpentinized, serpentine pseudomorphs retain relict olivine crystal shapes. Also associated with retrograde serpentinization is fine-grained magnetite. Olivine grains often occur in clusters, where they are commonly

associated with spinel (proper) or hercynite. Select samples from Groups A1, A2, and F contain anhedral to subhedral megacrysts of olivine (Figure 5.2c), which may represent relict primary igneous grains. In general, olivine grains are considered to be metamorphic, given that they coexist with amphibole grains. If the amphibole formed via metamorphic reactions involving the introduction of H₂O, it is unlikely primary olivine would persist through these conditions. In one sample, serpentinized olivine is surrounded by a reaction rim of orthopyroxene, which is then rimmed by hornblende (Figure 5.4a).

5.3.2 Clinopyroxene

Clinopyroxene generally occurs as equant, anhedral grains that display a granoblastic texture. The grains are often fractured, and in rare instances, serpentine fills the fractures. Clinopyroxene from Groups A2, B, D, and E displays exsolution textures with thin (~2 µm) orthopyroxene exsolution laminae (Figure 5.4c). This texture is common in clinopyroxene from sills in the south-central Baffin area (n=7), as opposed to those from Meta Incognita Peninsula (n=2). In a few samples, hornblende is observed overgrowing clinopyroxene grains (Figure 5.4b). Clinopyroxenes lack oscillatory zoning.

5.3.3 Orthopyroxene

Orthopyroxene grains are anhedral, sometimes ovoid in shape, and display red-green pleochroism under plane-polarized light (PPL). Orthopyroxene megacrysts (Figure 5.3a) are observed in samples from Groups B, D, and C2. In

some cases, they exhibit poikilitic textures, with chadacrysts of hornblende and/or ilmenite. Orthopyroxenes lack oscillatory zoning.

5.3.4 Amphibole

Amphibole is anhedral and defines a foliation fabric (Figure 5.3c) in numerous samples. The grains can vary greatly in size within a single sample, creating a seriate texture. Amphibole is of the hornblende variety and appears green to green-brown to brown in PPL, corresponding to an increase in Ti-content with increasing metamorphic grade. Through BSE microscopy, very-thin exsolution laminae can be observed within hornblende in select samples. The nature of the laminae could not be identified due to their sub-micrometer ($< 1 \mu\text{m}$) thickness. Amphiboles lack oscillatory zoning, and are interpreted to have replaced pyroxene at high temperature, catalyzed by the ingress of H_2O during metamorphism and deformation.

5.3.5 Plagioclase

Plagioclase occurs as anhedral grains, displaying characteristic polysynthetic twinning and occasional Carlsbad twinning. Within a subsection of the ultramafic portion of the Killapait sill, plagioclase occurs as round megacrysts (Figure 5.2a). The megacrystic plagioclase is commonly variably sericitized. Plagioclase grains lack oscillatory zoning.

5.3.6 Phlogopite

Phlogopite occurs as subhedral grains and appears medium brown in PPL. It occurs most commonly in Groups D and C2, generally between 15 and 5 modal %. Phlogopite grains often form weak to strong foliations, and outline micro-folding in a few samples.

5.4 Oxide Mineralogy

5.4.1 Spinel

Spinel occurs as dark-green (in PPL; [Figure 5.2a](#)), anhedral grains that typically form clusters. A minority display arcuate fracturing. Spinel is found in ultramafic samples and is associated with either olivine or orthopyroxene. Its presence is not confined to any particular geochemical group, though it is relatively common and in higher abundances (up to 10%) in samples from Groups A1, C2, and F. Sample 15SAB-L112A01 contains spinel exhibiting a vermicular texture, characteristic of formation through metamorphic growth. Additionally, spinel (proper) is known to form almost exclusively through metamorphism.

5.4.2 Magnetite and ilmenite

Magnetite and/or ilmenite ([Figure 5.5g](#)) are found in almost all samples, where these magnetite-ilmenite are generally distributed evenly in trace amounts, and in some cases occur up to 4 wt.%. Abundance is highest in mafic

units. Occasionally, magnetite+ilmenite exhibit network textures associated with the serpentinization of olivine, and in some samples have crystallized along fractures and grain boundaries. In a number of samples from the Killapait sill, magnetite exsolution laminae were observed in ilmenite grains. No magnetite or ilmenite were geochemically analyzed for this study.

5.5 Ore Mineralogy

The majority of samples from each group contain at least trace amounts of sulphide, which is typically pyrrhotite intergrown with minor amounts of chalcopyrite and/or pentlandite (a typical magmatic assemblage). The sulphides are typically anhedral, though euhedral grains are observed. Some samples exhibit remobilization of sulphide along grain boundaries or fractures, and in a few samples the sulphides appear round or ovoid in shape, perhaps preserving their primary shape from formation as immiscible sulphide droplets. The various textures and ore mineral associates are shown in [Figure 5.5a-h](#).

5.5.1 Pyrrhotite

Pyrrhotite is the most common sulphide mineral ([Figure 5.5a-g](#)), and by far the most abundant. The Fe-rich end-member of pyrrhotite, troilite, was identified in two samples; one from an intensely folded, and locally brecciated, layered mafic-ultramafic sill from the west coast of Cumberland Sound (15SAB-D196B01 - Group B), and the other from an ultramafic sill on Hall Peninsula (14SUB-

H029B2 - Group F).

5.5.2 Chalcopyrite

Chalcopyrite commonly occurs intergrown with pyrrhotite and occasionally pentlandite (Figure 5.5a-f,h); however, it also occurs as a single-mineral phase.

5.5.3 Pentlandite

Pentlandite was observed in samples from Groups B, D, and F (Figure 5.5c,e-f,h).

5.5.4 Pyrite

Pyrite generally occurs in samples from Groups C1 and C2, with few exceptions. Sample 15SAB-F013B01 contains pyrite with exsolution laminae of pyrrhotite.

5.6 Mineralogy of Detailed Sample Transects Across Two Mafic-Ultramafic Sills

5.6.1 Chidliak mafic-ultramafic sill

Mineralogical variation throughout the Chidliak sill is illustrated in Figure 5.6, with thin section scan and microphotographs shown in Figure 5.7. From the base to the top of the sill, there is an increase the % of hornblende, while other mineral phases (i.e. orthopyroxene, olivine and plagioclase) demonstrate variable modal

vol.%.

5.6.2 Killapait mafic-ultramafic sill

Mineralogical variation throughout the Killapait sill is illustrated in [Figure 5.8](#), with thin section scan and microphotographs shown in [Figure 5.9](#). From the base to the top of the sill, there is an increase the mod. al vol.% of plagioclase, relatively consistent olivine content, and variable modal vol.% of orthopyroxene, spinel and hornblende.

References

- Le Bas, M.J. and Streckeisen, A. (1991). The IUGS systematics of igneous rocks. *Journal of the Geological Society*, 148, 825-833.
- St-Onge, M.R., Wodicka, N. and Ijewliw, O. (2007). Polymetamorphic evolution of the Trans Hudson Orogen, Baffin Island, Canada: Integration of petrological, structural and geochronological data; *Journal of Petrology*, v. 48, p. 271– 302. doi:10.1093/petrology/eg1060.
- Whitney, D.L. and Evans, B.W. (2010). *American Mineralogist*, Volume 95, pages 185–187.

Figure 5.1: Ternary diagrams for the classification of rock samples based on the IUGS classification scheme for igneous rocks.

Figure 5.2a-c: Textures shown in rock slabs and thin section scans. **a)** Plagioclase megacrysts in a hornblende troctolite (14SAB-L013A01); **b)** Relict igneous layering in an olivine-bearing hornblende pyroxene gabbro norite (14SAB-L011A02), controlled by varying amounts of Hbl, Cpx and Pl; **c)** Olivine megacryst in an orthopyroxene olivine hornblende (14SAB-L003A02). Mineral abbreviations after [Whitney and Evans \(2010\)](#).

Figure 5.3a-c: Textures shown in thin sections (left) and micrographs (right). **a)** Orthopyroxene megacryst in an olivine orthopyroxene hornblendite (14SAB-H080A01) displaying seriate texture - thin section scan on left and micrograph (x5 magnification) on right; **b)** Relict orthocumulate texture (interpreted) preserved in an olivine hornblendite (14SAB-H080B01), where metamorphic hornblende represents the cumulate phase, while olivine represents the groundmass; **c)** Weak metamorphic foliation, outlined by the hashed red line, created by oriented hornblende in a pyroxene hornblende mela-gabbro norite (14SAB-L003B01). Mineral abbreviations after [Whitney and Evans \(2010\)](#).

Figure 5.4a-c: Reaction textures in PPL on left and XPL on right. **a)** Corona texture in an olivine-bearing hornblende pyroxene gabbro norite from the Killapait sill (14SAB-L012A02). The sequence of minerals is: completely serpentinized olivine (center), orthopyroxene, hornblende (outermost); **b)** Hornblende replacing zoned clinopyroxene (uralitization) in an olivine-bearing hornblende pyroxene gabbro norite (14SAB-L012A02) from the Killapait sill; **c)** Calcic clinopyroxene displaying exsolution laminae of orthopyroxene (14SAB-L011A02). Mineral abbreviations after [Whitney and Evans \(2010\)](#).

Figure 5.5a-h: Photomicrographs of ore minerals and associations taken at 10x magnification [except for h, taken at 5x magnification]. a) Anhedral pyrrhotite grain intergrown with minor chalcopyrite (14SAB-L010A02); b) Pyrrhotite grains intergrown with pentlandite and chalcopyrite (15SAB-L076A02); c) Pyrrhotite grain intergrown with chalcopyrite and pentlandite, and remobilized sulphide just above grain (15SAB-D196B01); d) Anhedral pyrrhotite intergrown with chalcopyrite (15SAB-L080A03); e) Anhedral pyrrhotite grain intergrown with chalcopyrite and pentlandite (14SAB-L007A01); f) Anhedral pyrrhotite grains intergrown with pentlandite and minor chalcopyrite (14SAB-L009A02); g) Intergrown ilmenite and magnetite with minor pyrrhotite (15SAB-M005A01); h) Round (top-left) and euhedral (bottom-right) grains of pentlandite intergrown with chalcopyrite (15SAB-L078A02).

Figure 5.6: Mineralogical changes (mineral chemistry and modal mineralogy) documented in the Chidliak sill of Meta Incognita Peninsula, Baffin Island, Nunavut, going up section. Symbols indicate (defined in Chapter 7) which geochemical group the sample belongs to, and correspond to the groups outlined in **Figure 5.1**.

Figure 5.7: Mineralogical and textural variations through the Chidliak sill, illustrated by representative hand sample photos, thin section scans, and microphotographs from each station, from stratigraphic base to top. Symbols indicate which geochemical group (defined in Chapter 7) the sample belongs to, and correspond to the groups outlined in **Figure 5.1**.

Figure 5.8: Mineralogical changes (mineral chemistry and modal mineralogy) documented in the Killapait sill of Meta Incognita Peninsula, Baffin Island, Nunavut, going up section. Symbols indicate which geochemical group (defined in Chapter 7) the sample belongs to, and correspond to the groups outlined in **Figure 5.1**.

Figure 5.9: Mineralogical and textural variations through the Chidliak sill, illustrated by representative hand sample photos, thin section scans, and microphotographs from each station, from stratigraphic base to top. Symbols indicate which geochemical group (defined in Chapter 7) the sample belongs to, and correspond to the groups outlined in **Figure 5.1**.

Table 5.1: Comparison of the average and range of mineral contents and within all eight geochemical groups (groups defined in Chapter 7).

Average	Group A1	Group A2	Group D	Group B	Group C1	Group C2	Group E	Group F
Plagioclase	0	26	8	33	25	45	42	13
Clinopyroxene	0	18	15	19	6	3	15	0
Orthopyroxene	7	12	17	12	5	4	17	10
Olivine	38	7	9	1	10	20	0	11
Amphibole	56	37	44	35	52	26	20	63
Biotite	0	0	2	0	0	1	0	0
Spinel	0	0	0	0	2	1	0	2
Range	Group A1	Group A2	Group D	Group B	Group C1	Group C2	Group E	Group F
Plagioclase	0	0-50	0-48	0-49	0-55	0-96	28-70	0-45
Clinopyroxene	0	0-53	0-50	8-25	0-35	0-30	0-20	0
Orthopyroxene	0-25	0-25	0-52	0-20	0-15	0-20	5-23	0-40
Olivine	15-60	0-70	0-40	0-5	0-40	0-70	0	0-46
Amphibole	39-80	2-70	0-70	15-84	20-90	4-50	0-32	45-80
Biotite	0	0-4	0-15	0	0-trace	0-5	0	0
Spinel	0-1	0-4	0-2	0-3	0-10	0-5	0	0-7

Chapter 6 – Mineral Chemistry

6.3 Mineral Chemistry

Geochemical groups mentioned below are defined on the basis of whole-rock geochemistry (Chapter 7).

6.3.1 Silicate Mineral Chemistry

6.3.1.1 Olivine

Olivine compositions show limited variation within the same geochemical group (up to 9% Fo variation, where “Fo” content equals the molar $\text{Mg}/(\text{Mg}+\text{Fe})$ ratio of olivine analyses). Within individual samples, olivine grains show little variation, with olivine compositions falling within a maximum of 1% Fo-content range. Groups D and C1 contain olivine with relatively low Fo content (64-73 Fo), group A1 contains olivine with relatively high Fo content (77-84 Fo), and olivine from Groups A2, C2, and F contain an intermediate Fo content (70-81 Fo). A summary of the ranges in Fo content within olivine from the various groups can be found in [Table 6.1](#). Olivine analyses are reported in Appendix III.

Ni concentration does not correlate with Fo content, however, olivine analyses from each group form a distinct field in Fo-Ni space (illustrated in [Figure 6.1](#)). Groups D and A1 contain relatively high-Ni olivine (1780-3020 ppm), Groups C1 and C2 contain relatively low-Ni olivine (190-1330 ppm), and group F contains olivine with intermediate Ni contents (1370-1850 ppm). Olivine analyzed from Group A2 falls in all three categories of Ni content. Olivine grains that have compositions that fall outside the field of igneous olivine may be: (1) of metamorphic origin; (2) derived from magmas in which sulphide saturation was

achieved prior to olivine crystallization; or (3) derived from magmas which re-equilibrated with Ni-sulphide prior to olivine fractionation. This includes all olivine analyzed from group C2 (metamorphic or crystallization after sulphide saturation), most olivine from Group D (metamorphic or crystallization after sulphide re-equilibration), and two olivine from Group A2 (metamorphic or crystallization after sulphide saturation).

MnO negatively correlates with Fo content in olivine (Figure 6.2), as is expected with fractionation as Mn^{2+} substitutes readily for Fe^{2+} in minerals (Ure and Berrow, 1982). Groups D and C1 contain relatively high MnO content (0.31-0.47 wt.%), Group A1 contains relatively low MnO content (0.25-0.34 wt.%), and Groups F and C2 contain intermediate MnO content (0.28-0.39 wt.%). Group A2 spans the three divisions (0.22-0.42 wt.%).

6.3.1.2 Clinopyroxene

Clinopyroxene is Ca-rich and almost exclusively diopside (Figure 6.3). The diopside (Di) content ranges from 49-89 %Di. Cr contents generally range from 0 to 920 ppm, and as high as 3810 ppm. All clinopyroxene grains contain low, but variable amounts of Al_2O_3 (0.65-3.95 wt.%), TiO_2 (0.03-0.57 wt.%), and Na_2O (<0.51 wt.%). The diopside and chromium content of clinopyroxene in the various geochemical groups is provided in Table 6.2. Clinopyroxene analyses are reported in Appendix III.

6.3.1.3 Orthopyroxene

The enstatite (En) content of orthopyroxene ranges from 45-81 %En (Figure 6.4). The pyroxene is characterized by a small range in CaO (0.17-0.87 wt.%) and Al₂O₃ (0.67-3.73 wt.%), and MnO content generally varies between 0.25-0.87 wt.%. Enstatite also has low TiO₂ (<0.23 wt.%) and Na₂O (<0.04 wt.%), with no clear correlation with Mg# (not shown). The enstatite and chromium content of orthopyroxene from the various geochemical groups are presented in Table 6.3. Orthopyroxene analyses are reported in Appendix III.

6.3.1.4 Amphibole

Amphibole forms two separate trends on the Leake (1978) amphibole classification diagram (Figure 6.5). Amphibole from mafic rocks trends in composition from magnesiohornblende to tschermakite. Amphibole from ultramafic rocks forms a trend at a higher Mg#, from actinolite through magnesiohornblende to tschermakite, and from actinolite through magnesiohornblende, ferrohornblende to ferrotschermakite. The Si content ranges from 6.03 to 7.66 [atoms per formula unit (apfu)], and Mg# ranges from greatly from 33-89. Cr₂O₃ in hornblende is generally <0.78 wt.%, except for Group A1, which shows a range from 0.52 to 0.78 wt.%. TiO₂ shows a negative correlation with Mg#, ranging from 0.07 up to 2.85 wt.%. Amphibole analyses are reported in Appendix III.

6.3.1.5 Plagioclase

Anorthite (An) content ranges from 38 (andesine) to 99% (pure anorthite), however, the majority of plagioclase grains analyzed have labradorite, bytownite, and anorthite compositions ([Figure 6.6](#)). In differentiated basic rocks, the range in An-content of plagioclase is normally restricted between An₃₀ and An₈₅ ([Gribble & Hall, 1992](#)), suggesting that the more anorthitic compositions are metamorphic. [Phinney & Halls \(2001\)](#) noted that high-An content (An₈₀₋₈₅) of plagioclase megacrysts indicates the melts experienced a protracted stay in crustal magma chambers <10 kbar. This suggests that plagioclase in select samples from Groups A2, C1, C2, D and F are may be igneous in origin, however, they are most likely reflect the high degree of metamorphism achieved. Groups A2 and D exhibit the greatest variability in An-content, spanning compositions from andesine to anorthite. Group C2 spans compositions from labradorite through anorthite. Group B is confined to labradorite compositions, Group F grains all fall within the bytownite field, and Group C1 is almost exclusively anorthitic. The An content content of plagioclase from the various geochemical groups, is presented in [Table 6.4](#). Plagioclase analyses are reported in Appendix III.

6.3.1.6 Phlogopite

Phlogopite grains analyzed ([Figure 6.7](#)) in a sample from Group C2 are distinctly richer in FeO (~13 wt.%) compared to the sample from Group D (~7 wt. %, respectively). Based on the classification from [Deer et al. \(1992\)](#), the micas fall in the eastonite field, an aluminous variety of phlogopite. Phlogopite analyses

are reported in Appendix III.

6.3.2 Oxide Mineralogy

6.3.2.1 Spinel

The spinel grains are Al-rich and have an MgO content ranging between 11-17 wt.%, as per [Figure 6.8](#). Most grains plot in the spinel field, while four grains from Groups A1 and D are classified as hercynite (less Mg-rich). The Al-rich character ([Figure 6.9](#)) of spinel is controlled by Al exchange with coexisting pyroxene. The low $\text{Cr}/(\text{Cr}+\text{Al})$ ratio may result from spinel-clinopyroxene subsolidus exchange, which effectively exsolves significant amounts of Al from clinopyroxene resulting in highly aluminous spinel mineral phases ([Sinton 1977](#); [Barnes, 2001](#)).

A positive correlation of Al_2O_3 with MgO content is observed. There is a narrow range in Al_2O_3 from 57-64 wt.%, with the exception of Group A1 spinel which has Al_2O_3 content as low as 48 wt.%. Similarly, there is a relatively narrow range in Cr_2O_3 (0.02-4.26 wt.%), with the exception of Group A1 spinel which has Cr_2O_3 content as high as 16 wt.%. $\text{Cr}/(\text{Cr}+\text{Al})$ of spinel ranges between ~0.00-0.05. Spinel analyses are reported in Appendix III.

6.3.3 Ore Mineralogy

Sulphide analyses are reported in Appendix III.

6.3.3.1 Pyrrhotite

Fe and S contents of pyrrhotite range between 57-64 and 33-41 wt. %, respectively. Ni values are highest in Groups D and C2, up to 1.23 wt.%. Across all groups, trace amounts of Cu (<1550 ppm), Zn (<340 ppm), As (<200 ppm), and Mn (<270 ppm) are also present.

6.3.3.2 Chalcopyrite

Fe, S and Cu contents in chalcopyrite range between 29-31, 33-35, and 33-36 wt.%, respectively. Grains contain Co and Ni up to 1010 and 2690 ppm, respectively. Trace amounts of Zn, As, and Mn reach levels of 2120, 990, and 220 ppm, respectively.

6.3.3.3 Pentlandite

Fe and S contents in pentlandite range between 28-41% and 31-34%, respectively, and Ni varies between 25-38 wt. %. Pentlandite grains from Groups D and F contain the highest % of Ni. Co and Cu content is generally <2.15 wt.% and <1700 ppm, respectively. Grains from Group B yield Co and Cu contents up to 13 wt.% and 5140 ppm. Pentlandite grains also contain trace Zn (<520 ppm), As (<330 ppm), and Mn (<220 ppm).

6.3.3.4 Pyrite

Fe and S contents in pyrite range between 45-48% and 52-54%, respectively. Cobalt (Co) is generally low (<1280 ppm), but in the Killapait sill,

pyrite grains are typically more Co-rich, up to 1.12 wt.% Co. All pyrite grains contain at least some amount of Ni (<1610 ppm), and half contain some amount of Cu (<280 ppm). There are also trace amounts of Zn (<240 ppm), As (<100 ppm), and Mn (<60 ppm). Pyrite is devoid of Pb, Sb and Cd.

6.3.4 Mineralogy of Detailed Sample Transects Across Two Mafic-Ultramafic Sills

6.3.4.1 Chidliak mafic-ultramafic sill

Mineral chemistry variation throughout the Chidliak sill is illustrated in [Figure 6.10](#). Going up through the magmatic section, assuming a basal pyroxenite overlain by peridotite and capped by gabbro, there is a marked increase in the En-content of orthopyroxene, Fo-content of olivine, and Mg# of spinel. Conversely, there is a progressive decrease in the An-content of plagioclase. The Mg# of amphibole increases within the basal pyroxenite unit, decreases through the transition into the overlying layered peridotite + pyroxenite unit, and continues to increase briefly followed by a decline into the hornblende gabbroic + noritic portion of the sill. Given that these assemblages are metamorphic, the chemistry of any single mineral phase is likely reflective of the overall bulk chemistry of the sample, as opposed to being representative of the chemistry of primary mineral phases crystallizing from a magma. Thus, mineral chemistry variation with height in the sill is suggested to merely reflect bulk geochemistry.

6.3.4.2 Killapait mafic-ultramafic sill

Mineral chemistry variation throughout the Killapait sill is illustrated in [Figure 6.11](#). Going up magmatic section, there is an overall decrease in the Fo-content of olivine, An-content of plagioclase and Mg# of amphibole. There is an increase in the En-component of orthopyroxene. There is no consistent trend apparent in the variation of the Mg# of spinel and Di-component of clinopyroxene.

Where geochemical trends with height in the sill are apparent, these are suggested to be coincidental, based on the interpretation that the Killapait sill formed through multiple intrusions of geochemically distinct magma types.

References

- Bucher, K. and Grapes, R. (2011). *Petrogenesis of Metamorphic Rocks*. Springer-Verlag Berlin Heidelberg: pp. 191-224 & 339-393. ISBN: 978-3-540-74168-8.
- Deer, W. A., Howie, R. A., and Zussman J. (1963). *Rock-forming Minerals*, 5, Non-silicates. Longmans, London.
- Deer WA, Howie A and Zussman J. (1986). *An introduction to rock-forming Minerals*, 17th edition. Longman Ltd, UK, 528.
- Deer, W.A.; Howie, R.A.; Zussman, J. (1992). *An introduction to the rock forming minerals*. Engleand Editors. Longamn Scientific and Technical: 695 p. Harlow.
- Gribble, C.D. and Hall, A.J. (1992). *Optical Mineralogy: Principles and Practice*. UCL Press, London, 303 pp.
- Leake, B.E., Woolley, A.R., Arps, C.E.S., Birch, W.D., Gilbert, M.C., Grice, J.D., Hawthorne, F.C., Kato, A., Kisch, H.J., Krivovischev, V.G., Linthout, K., Laird, J., Mandarino, J.A., Maresch, W.V., Nickel, E.H., Rock, N.M.S., Schumacher, J.C., Smith, D.C., Stephenson, N.C.N., Ungaretti, L., Whittaker, E.J.W. & Youzhi, G. (1997). *Nomenclature of amphiboles: Report of the subcommittee on amphiboles on the International Mineralogical Association, Commission on new minerals and mineral names*. *American Mineralogist* 82, 1019–1037.
- Morimoto, N., Fabries, J., Ferguson, A. K., Ginzburg, I. V., Ross, M., Seifert, F. A., Zussman, J., Aoki, K. & Gottardi, G. (1988). *Nomenclature of pyroxenes*:

Report of the Subcommittee on Pyroxenes of the International Mineralogical Association, Commission on New Minerals and Mineral Names. *American Mineralogist* 73, 1123-1133.

Phinney, Wm.C., and Halls, H.C. (2001). Petrogenesis of the Early Proterozoic Matachewan dyke swarm, Canada, and implications for magma emplacement and subsequent deformation. *Canadian Journal of Earth Sciences* 38: 1541-1563.

Simkin, T. and Smith, J.V. (1970). Minor element distribution in olivine. *Journal of Geology*, 78, 304-325.

St-Onge, M.R., Wodicka, N. and Ijewliw, O. (2007). Polymetamorphic evolution of the Trans Hudson Orogen, Baffin Island, Canada: Integration of petrological, structural and geochronological data; *Journal of Petrology*, v. 48, p. 271– 302. doi:10.1093/petrology/eg1060.

Ure, A.M. and Berrow, M.L. (1982). The elemental constituents of soils. In: Bowen HJM (ed) *Environmental chemistry*. Royal Society of Chemistry, London, pp 94–203.

Whitney, D.L., and Evans, B.W. (2010). *American Mineralogist*, Volume 95, pages 185–187.

Figure 6.1: Bivariate plot illustrating the changes in Ni content (ppm) with respect to forsterite (Fo) content in olivine. Overlain on this plot is the [Simkin and Smith \(1970\)](#) field for olivine in layered intrusions. Symbols indicate which geochemical group (groups defined in Chapter 7) the olivine grain belongs to.

Figure 6.2: Bivariate plot illustrating the increase in MnO content (wt.%) with decreasing forsterite (Fo) content in olivine. Symbols indicate which geochemical group (groups defined in Chapter 7) the olivine grain belongs to.

Figure 6.3: Compositional range of calcic clinopyroxene from sills sampled throughout southern Baffin Island (sample locations provided on [Figures 2.8 & 2.9](#)). Symbols indicate which geochemical group (groups defined in Chapter 7) the clinopyroxene grain belongs to. Diagram from [Morimoto et al. \(1988\)](#).

Figure 6.4: Compositional range of the orthopyroxene from sills sampled throughout southern Baffin Island. Symbols indicate which geochemical group (groups defined in Chapter 7) the orthopyroxene belongs to. Diagram from [Morimoto et al. \(1988\)](#).

Figure 6.5: Classification of amphibole from [Leake et al. \(1997\)](#). Amphibole compositions define two separate trends: amphibole from mafic rocks trend from magnesiohornblende to tschermakite. Amphibole from ultramafic rocks define a higher Mg# trend from actinolite through magnesiohornblende to tschermakite, and from actinolite through magnesiohornblende, ferrohornblende to ferrotschermakite. Symbols indicate which geochemical group (groups defined in Chapter 7) the amphibole grain belongs to.

Figure 6.6: Classification of plagioclase from sills sampled through southern Baffin Island (after [Deer et al., 1963](#)). Symbols indicate which geochemical group (groups defined in Chapter 7) the plagioclase belongs to.

Figure 6.7: Compositional classification of phlogopite (after [Deer et al., 1992](#)).

Phlogopite shows an eastonite affinity (Al-rich phlogopite). Symbols indicate which geochemical group (groups defined in Chapter 7) the phlogopite grain belongs to.

Figure 6.8: Classification of spinel from sills sampled throughout southern Baffin Island (after [Deer et al., 1992](#)), showing the spinel to hercynitic character. Symbols indicate which geochemical group (groups defined in Chapter 7) the spinel belongs to.

Figure 6.9: Cr- Fe^{3+} -Al ternary plot of spinel mineral chemistry from samples of sills collected from southern Baffin Island, illustrating their Al-rich character. Symbols indicate which geochemical group (groups defined in Chapter 7) the spinel grain belongs to.

Table 6.1: Ranges in forsterite (%) and Ni (ppm) content in olivine from mafic and ultramafic rocks for each geochemical groups (groups defined in Chapter 7). “N/A” (not available) signifies olivine absent; “Not Analyzed” indicates that the mineral is present but was not analyzed.

Mineral	Geochemical Group	Mafic		Ultramafic	
Olivine		Fo (%)	Ni (ppm)	Fo (%)	Ni (ppm)
	Group A1	N/A	N/A	77-84 (n=5)	1780-3020 (n=5)
	Group A2	Not Analyzed	Not Analyzed	72-81 (n=3)	1040-2370 (n=3)
	Group B	Not Analyzed	Not Analyzed	N/A	N/A
	Group C1	N/A	N/A	67-68 (n=1)	930-1330 (n=1)
	Group C2	70-73 (n=2)	190-730 (n=2)	77-78 (n=1)	760-880 (n=1)
	Group D	N/A	N/A	64-73 (n=6)	1850-2600 (n=6)
	Group E	N/A	N/A	N/A	N/A
	Group F	N/A	N/A	74-78 (n=4)	1370-1850 (n=4)

Table 6.2: Range in diopside (%) and Cr (ppm) content of calcic clinopyroxene grains from mafic and ultramafic rocks of each geochemical groups (groups defined in Chapter 7). “N/A” signifies clinopyroxene absent; “Not Analyzed” indicates that the mineral is present but was not analyzed.

Mineral	Geochemical Group	Mafic		Ultramafic	
Clinopyroxene		Di (%)	Cr (ppm)	Di (%)	Cr (ppm)
	Group A1	N/A	N/A	N/A	N/A
	Group A2	60-74 (n=3)	40-780 (n=3)	N/A	N/A
	Group B	59-74 (n=2)	0-220 (n=2)	N/A	N/A
	Group C1	N/A	N/A	81-85 (n=2)	1100-3810 (n=2)
	Group C2	61-85 (n=1)	50-260 (n=1)	89 (n=1)	20 (n=1)
	Group D	42-69 (n=3)	0-920 (n=3)	84-87 (n=2)	220-670 (n=2)
	Group E	Not Analyzed	Not Analyzed	N/A	N/A
	Group F	N/A	N/A	N/A	N/A

Table 6.3: Range in enstatite (En%) and Cr (ppm) content of orthopyroxene from mafic and ultramafic rocks from each geochemical groups (groups defined in Chapter 7). “N/A” signifies orthopyroxene absent; “Not Analyzed” indicates that the mineral is present but was not analyzed.

Mineral	Geochemical Group	Mafic		Ultramafic	
Orthopyroxene		En (%)	Cr (ppm)	En (%)	Cr (ppm)
	Group A1	N/A	N/A	81 (n=1)	1010 (n=1)
	Group A2	46-65 (n=4)	0-560 (n=4)	76-81 (n=3)	0-1020 (n=3)
	Group B	45-61 (n=2)	0-100 (n=2)	N/A	N/A
	Group C1	Not Analyzed	Not Analyzed	75 (n=2)	100-180 (n=2)
	Group C2	47-76 (n=3)	0-230 (n=3)	79-80 (n=1)	270-450 (n=2)
	Group D	52 (n=1)	160 (n=1)	71-77 (n=6)	290-950 (n=6)
	Group E	Not Analyzed	Not Analyzed	N/A	N/A
	Group F	Not Analyzed	Not Analyzed	78-81 (n=4)	0-300 (n=4)

Table 6.4: Ranges in anorthite (%) content of plagioclase from mafic and ultramafic rocks from each geochemical group (groups defined in Chapter 7). “N/A” signifies plagioclase absent; “Not Analyzed” indicates that the mineral is present but was not analyzed.

Mineral	Geochemical Group	Mafic	Ultramafic
Plagioclase		An (%)	An (%)
	Group A1	N/A	N/A
	Group A2	39-86 (n=4)	N/A
	Group B	53-61 (n=2)	N/A
	Group C1	90-96 (n=1)	N/A
	Group C2	53-99 (n=6)	N/A
	Group D	38-92 (n=2)	Not Analyzed
	Group E	Not Analyzed	N/A
	Group F	81-99 (n=2)	99 (n=1)

Chapter 7 – Geochemistry

7.1 Whole Rock Geochemistry

The sills of southern Baffin Island have had a long history of metamorphism and deformation (Chapter 3). As a result, the rock units have been recrystallized, forming polygonal grains in most cases, and primary igneous textures have been obliterated (Chapter 5), with a few rare exceptions. This makes recognizing cumulate vs non-cumulate units extremely difficult. However, certain trace elements can be used to infer accumulation or removal of specific mineral phases from the magma, based on anomalies present in normalized trace element diagrams. To negate the effects of fractional crystallization, trace element ratios and patterns are primarily utilized for geochemical discrimination, from which eight geochemical groups are defined (see legend in [Figure 7.1](#), and geochemical discrimination in [Figure 7.3](#))

A significant number of samples (i.e. 79 samples) were utilized in the study of the suite of sills, however, as exhibited to by the layered Killapait sill, there is potential for vast differences in the trace element geochemistry within a single sill. Given that only one or two samples were collected from the majority of sills sampled, there is a relatively high probability that many of the layered sills contain additional geochemical signatures that were not recognized due to lack of detailed sampling. For these reasons, it is quite challenging to understand the relationship between the different geochemical groups and the magmatic processes that occurred on a sill-scale.

7.1.1 Element mobility

Given that the supracrustal rocks of southern Baffin Island have experienced amphibolite to granulite facies metamorphism, it is essential to consider the effects of secondary remobilization of elements and oxides. To assess element mobility, the method of [Cann \(1970\)](#) was implemented, whereby the elements to be evaluated are plotted in separate bivariate diagrams against a generally accepted immobile element (e.g., Zr used in this study). If both variables are incompatible during fractionation, and remain immobile through metamorphism, samples from suites of igneous rocks derived from the same mantle source should produce straight lines. A large degree of scatter indicates that one of the elements was mobile and secondary processes have altered its concentration. Alternatively, scatter may indicate the samples are unrelated, or that one of the samples represents a cumulate, with higher or lower concentrations of that particular element.

Incompatible elements generally considered to be immobile (e.g., Ti, Y, P) form coherent trends with small degrees of scatter ([Figure 7.2a-c](#)). However, most large-ion lithophile elements (LILE; e.g., Ba, Rb, Pb^{2+} , Sr, Cs, K) show significant scatter ([Figure 7.2d-f](#)). The coherent patterns observed in chondrite-normalized rare-earth elements (REE) diagrams ([Figure 7.3a](#)) and primitive mantle-normalized spider diagrams ([Figure 7.3b](#)) further support the limited mobility of the aforementioned immobile geochemical constituents. This suggests limited remobilization of these elements due to metamorphism or metasomatism.

These observations are consistent with numerous other geochemical studies on high-grade Archean rocks that indicate the major elements Al, Ti, Fe and P, the high-field-strength elements (HFSE; Th,Nb,Ta,Zr,Hf), the REE (except Eu), transition metals (Cr,Ni,Sc,V), and Y are relatively immobile during metamorphism and alteration (e.g. Winchester and Floyd, 1976; Sun and Nesbitt, 1978; Ludden et al., 1982; Arndt and Nesbitt, 1982, Jochum et al., 1991; Lafleche et al., 1992a; Arndt, 1994; Condie, 1994; Barley et al., 2000; Hanski et al., 2001; Shimizu et al., 2004; Arndt, 2008). A number of studies also demonstrate that, in contrast, LILE (e.g., Cs,Na,K,Rb,Ba,Ca,Pb,Sr) are often mobile during metamorphism and alteration (Leshner et al., 2001; Arndt, 2008). Accordingly, this study will focus mostly on elements and oxides that appear to have been immobile during the high-grade regional metamorphism.

7.1.2 Classification through chondrite-normalized rare-earth element plots

Samples from the mafic and ultramafic units of various sills throughout most of southern Baffin Island were discriminated into eight unique geochemical groups (Figure 7.1) on the basis of a variety of geochemical diagrams, but primarily by chondrite-normalized rare-earth element diagrams (Figure 7.3a). The few samples that displayed a unique chondrite-normalized REE patterns were removed from the dataset. Further refinement was conducted on the basis of sample relationships on a number of immobile-element-based bivariate and trivariate diagrams (i.e. >50), some of which are shown. This refinement removed samples that consistently did not plot with, or form trends with, the other samples

designated to same geochemical group. Furthermore, every group, contains both mafic and ultramafic samples, with the exception of Groups A1 and E, which contain only ultramafic and mafic units, respectively. In many cases, multiple geochemical groups exist in the same sill. [Table 7.1](#) contains information on the occurrences of the various geochemical groups within the different types of sills.

Based on number of occurrences and their spatial distribution, Group D (high-TiO₂ group) and Groups A1 and A2 (low-TiO₂ groups) are considered to be major constituents of Frobisher suite magmatism, whereas Groups B, C1, C2, E and F are minor components. The samples within each group exhibit very similar patterns, indicating crystallization from magmas with broadly similar composition. Throughout the rest of this thesis, the samples will be described with respect to their assigned geochemical group.

The rare-earth elements (REEs) are considered to be immobile during metamorphism, but remobilization of the light rare-earth elements (LREE) may occur during high-flux passage of metasomatic fluids through rocks during metamorphism. However, metasomatism does not appear to have significantly affected LREE, given the general lack of scatter in the fractionation trends of most REE ([Figure 7.2](#); most REE not shown). The resilience of this subgroup of high-field-strength elements (HFSE) is also illustrated in the consistency of patterns in chondrite-normalized REE diagrams ([Figure 7.3a](#)).

Group A1 exhibits a relatively flat REE chondrite-normalized REE pattern (very weakly LREE-enriched), similar to chondrite. Total REE contents vary from ~5x to 8x chondrite, with normalized La/Yb ratios varying from ~1.27 to 1.70.

Two samples have very minor negative Eu anomalies. Lithologically, this group is comprised of hornblendite (n=4) and peridotite (n=2) units from ultramafic and mafic-ultramafic sills.

Group A2, similar to A1, has a flat heavy rare-earth element (HREE) pattern, and exhibits increasing fractionation from middle rare-earth elements (MREEs) to LREEs. This pattern resembles E-MORB ([Schilling, 1973](#)). Total REE contents vary from ~10x to 39x chondrite, with La/Gd and Gd/Yb ratios varying from ~1.18 to 2.36 and ~1.17 to 1.62, respectively. Half of the samples contain minor negative Eu anomalies. Lithologically, this group contains gabbro (n=5), norite (n=3), gabbro-norite (n=6), hornblendite (n=2) and peridotite (n=1) units from mafic, ultramafic and mafic-ultramafic sills.

Group D has a sinusoidal, enriched REE pattern with weakly fractionated LREE, and significantly more fractionated MREE to HREE, similar to average OIB. Total REE contents vary from ~18x to 87x chondrite with La/Gd and Gd/Yb ratios ~1.63 to 3.68 and ~2.02 to 3.74, respectively. Half of samples in this group exhibit either weak positive or weak negative Eu anomalies. Lithologically, this group contains gabbro (n=2), gabbro-norite (n=2), pyroxenites (n=2), peridotites (n=1), and hornblendites (n=7) from mafic, ultramafic, and mafic-ultramafic sills.

Group B's enriched REE pattern is characterized by a relatively flat HREE and moderately enriched LREE and MREE. Total REE contents vary from ~5x to 23x chondrite, with La/Gd and Gd/Yb ratios varying from ~1.33 to 1.63 and ~2.86 to 3.41, respectively. Three samples have moderate negative Eu anomalies, which vary from weak to moderate. Lithologically, this group contains

gabbro (n=1) and hornblendite (n=1) units from mafic and mafic-ultramafic sills. Most samples from this group exhibit positive Eu anomalies, indicating plagioclase accumulation.

Group C1 is characterized by its highly enriched REE pattern, where LREE to MREE are steep and the HREE are moderately inclined and undulating. Total REE contents vary from ~12x to 53x chondrite, with La/Gd and Gd/Yb ratios varying from ~3.16 to 6.74 and ~2.32 to 3.23, respectively. Generally, the samples with higher REE concentrations contain negative Eu anomalies, and those with lower REE contents contain positive Eu anomalies. Lithologically, it contains gabbro (n=5), gabbro (n=6), hornblendite (n=2) and peridotite (n=1) units found within ultramafic and mafic-ultramafic sills.

Group C2, similar to C1, is also highly LREE enriched. Total REE contents vary from ~3x to 29x chondrite, with La/Gd and Gd/Yb ratios varying from ~5.87 to 16.65 and ~1.19 to 3.24, respectively. In some samples, HREE concentrations are below chondrite. All samples contain positive Eu anomalies, with the large majority with $\text{Eu}/\text{Eu}^* > 1.6$ (up to 3.2). As the magma becomes more evolved, this anomaly becomes less pronounced. Lithologically, it comprises anorthosite (n=1), gabbro (n=6), norite (n=2), gabbro (n=1), and peridotite (n=2) units from mafic and mafic-ultramafic sills.

Group E has a relatively flat, MORB-like REE pattern, all with distinctive major negative europium anomalies between 0.6-0.5 Eu/Eu^* . Total REE contents vary from ~13x to 20x chondrite, with La/Yb ratio ~0.93 to 1.57. Lithologically, this

group contains only gabbro (n=3) units within mafic and mafic-ultramafic sills. Samples in this group exhibit significant Eu anomalies.

Group F has a REE pattern in which samples become weakly enriched from LREE to HREE, typical of komatiitic magmas (e.g. [Condie, 1976, 1981](#); [Zhai & Santosh, 2011](#)). Total REE contents vary from ~2x to 5x chondrite (lowest of all groups), with La/Yb ratios ~-0.50 to -1.29. A number of samples display weak positive Eu anomalies and one is weakly negative. Lithologically, it contains gabbro (n=2), norite (n=1), hornblendite (n=7), and peridotite (n=1) units, and is found within mafic-ultramafic and ultramafic sills.

The relationships between the different patterns is illustrated in [Figure 7.4](#), where the $(La/Sm)_N$ ratio reflects the slope of the LREE and $(Gd/Yb)_N$ reflects the slope of the MREE to HREE. In general, Groups C1, C2 and D have relatively high Gd/Yb ratios (~2 to 3.3), though Group C2 samples fall as low as 1.2. Within this range, there is a significant increase in La/Sm ratios from group D through Group C1 into Group C2, with a progressive drop in Gd/Yb ratio and increasing spread with increasing La/Sm (the latter may be due to contamination, which is discussed later in the chapter). The remaining groups (i.e. A1,A2,B,E,F) have relatively low Gd/Yb ratios, where Groups A1, A2 and B show a very similar range in Gd/Yb ratios (~1.2 to 1.7), and Groups E and F together show a similar spread. Within these ranges, there is a distinct increase in La/Sm ratios from Group A1 through Group A2 to Group B, and from Group F to Group E.

7.1.3 Major elements

7.1.3.1 Major and minor oxides versus Mg#

Major element variations [SiO_2 (a), CaO (b), Al_2O_3 (c), Cr_2O_3 (d), P_2O_5 (e) & TiO_2 (f)] with Mg# [$100 \cdot \text{MgO} / (\text{MgO} + \text{FeO})$] as an index of fractionation are shown in Figure 7.5. Mg# appears to be the best proxy for index of fractionation, compared to other proxies (e.g., SiO_2 , MgO , Zr). Within all Mg# diagrams (with FeO calculated directly from Fe_2O_3 , representing total Fe), Groups B, C1 and E are the only groups that do not consistently demonstrate coherent trends, and consequently, are not be described in terms of geochemical trends in all cases. For Groups B and E, the apparent lack of trends may partially be due to a lack of data. Additionally, Group A1 has too narrow of an Mg# range to decipher trends. Major and trace element diagrams reveal significant scatter in large-ion lithophile elements (LILE) amongst all groups through the range of Mg#, with the exception of Group F. This indicates that the LILE became mobile during metamorphism-related metasomatism, supporting previous observations of LILE plotted against Zr.

Groups A2 and D show a very broad range in Mg# (27-66 and 22-65, respectively) but are mostly concentrated between Mg#s 35-50 and 57-65, respectively. Group A1 has a narrow range in Mg# (between 68 and 71), with the exception of one sample containing an Mg# of 77. Group C2 is restricted to high Mg#s (>48), with samples containing Mg# as high as 70. Group C1 contains a wide range of Mg# from 32 to 69, and Group E samples range from 47 to 58. Samples from group F have high Mg#s, ranging from 60 to 73.

The Cr_2O_3 contents of the sills decrease with increasing Mg#, indicating removal of either Cr-bearing spinel, clinopyroxene or olivine during fractionation. Group A1 forms a cluster at high Cr_2O_3 and Mg#. Groups A2 and D form similar trends with similar concentrations of Cr_2O_3 at identical Mg#, while Groups B, C1 and C2 generally contain significantly less Cr_2O_3 contents than the rest of the groups, at constant Mg#. Group F contains intermediate Cr_2O_3 contents.

TiO_2 , P_2O_5 , and Al_2O_3 contents in all groups increase with decreasing Mg#, with Group D containing greater concentrations of TiO_2 , P_2O_5 and lower concentration of Al_2O_3 , at similar Mg#. This indicates that these oxides remained incompatible during fractionation and suggests that fractionation of Fe-Ti-oxides, such as ilmenite or titanomagnetite (also titanite or rutile, in some cases), and phosphate-bearing minerals (e.g. apatite, monazite) were not significant fractionating mineral phases during the evolution of the parental magmas to all groups. Al_2O_3 and CaO increase with fractionation until they level out at around an Mg# of 60, suggesting the initiation of plagioclase fractionation. This is consistent with a change in lithology from ultramafic to mafic as Mg# in samples decreases through Mg# 60. Furthermore, concentrations of these three major element oxides are distinctly lower in Groups B, C2, E and F as compared to A (and D).

In numerous diagrams, trends extrapolated from Group A2 intersect the “cloud” of Group A1 samples (e.g. [Figure 7.5a,c,e-f](#)), suggesting the two groups may be co-genetic and that common petrogenetic processes (e.g. fractional crystallization or cumulate addition) control their chemical compositions.

7.1.3.2 High- and low-Ti groups

Many sill provinces, and LIPs in general, have been subdivided into high-Ti and low-Ti magma types (e.g. Emeishan, Karoo-Ferrar LIPs; [Duncan et al., 1997](#)). The high-Ti/low-Ti division reflects the groups' overall incompatible element or oxide levels, and is primarily discriminated based on differing TiO_2 and P_2O_5 concentrations ([Hooper, 1982](#)). Based on a TiO_2 versus Mg# diagram ([Figure 7.6](#)), the geochemical groups for the Frobisher suite can be subdivided into high-Ti and low-Ti groups. Group D is considered high-Ti (also richest in P), while the remainder of the groups are low-Ti. However, given the high abundance of sills containing geochemical signatures of Groups A1, A2 and D, and the rest of the groups existing as minor components, the term “low-Ti” will be applied strictly to Groups A1 and A2. Furthermore, P_2O_5 occurs at trace levels in all or most samples from Groups A1, A2, B, C1, C2, E, and F, and TiO_2 occurs at trace levels in most group C2 samples.

In continental flood basalts of LIPs, a low-Ti character of tholeiitic basaltic magmas has been interpreted to reflect either mantle melting conditions, such as shallow depths of melting or higher degrees of partial melting in the upper mantle ([Arndt et al., 1993](#); [Xu et al., 2004a](#)) or crustal contamination by continental crust and/or subcontinental lithospheric mantle ([Carlson, 1991](#); [Peate, 1997](#); [Ewart et al., 1998, 2004](#)). High-Ti character, on the other hand, show geochemical and isotopic similarities to OIBs (i.e. intraplate geochemistry), and are interpreted to contain an asthenospheric mantle or plume component ([Arndt et al., 1993](#); [Zhao et al., 1994](#); [Ewart et al., 1998, 2004a](#)).

Similar to the continental flood-basalts (CFBs) in the Afro-Arabian LIP (e.g. [Arndt and Menzies, 2005](#)), the low-Ti groups exhibit relatively flat chondrite-normalized REE patterns, and the more alkaline samples (i.e. high-Ti Group D; discussed in 7.1.2) contain higher incompatible element concentrations at similar Mg# and more fractionated REE patterns. Causes for the difference in TiO₂ concentration at similar Mg# may reflect either a heterogeneous source, difference in degrees of partial melting, or contamination.

7.1.3.3 FeO versus MgO

The variation in FeO with respect to decreasing MgO ([Figure 7.7](#)), an oxide that typically decreases as magmas evolve, illustrates different trends between certain geochemical groups. Groups A1, A2, and D show weak (tholeiitic) Fe-enrichment trend, as FeO content increases with decreasing MgO. In contrast, Groups C2 and F demonstrate decreasing FeO with decreasing MgO, suggestive of Fe-saturation attainment (moving towards a ‘calc-alkaline trend’). However, Group F is tholeiitic and its Ti concentrations continue to increase with decreasing Mg#, indicating that this is not the case in this group, as Fe + Ti would normally be removed from the magma together. Groups B, C1 and E do not display clear trends, however, FeO contents at constant MgO in Group C1 are very similar to that of Groups A2 and D while Group B bears a stronger resemblance to Groups C2 and F.

7.1.4 Trace elements

7.1.4.1 Trace elements versus Mg#

Trace element variations with Mg# as an index of fractionation are shown in Figure 7.8. Included in these diagrams are highly compatible trace elements (Co, Ni), high-field strength elements (Th, La, Sm, Y, Zr, Sc and Nb), and total REE. Incompatible trace elements which have distinctly higher concentrations in the high-Ti Group D compared to all other groups include Ce, Eu, Ga, Gd, Hf, La, Nb, Nd, Pr, Sm, Ta, Zr, and Hf, as well as the compatible element Ni (some elements are not shown in Figure 7.8).

The Ni concentrations in Group D are high, decreasing in a linear trend from Mg# ~62 to 23. Group A1 (highest Ni contents) shows a steep trend, while Group A2 has lower values, forming a shallower trend decreasing with decreasing Mg#. Ni contents in Groups C2 and F rapidly decline from Mg# ~70 to 65, from ~750 ppm to ~200 ppm, suggesting partitioning into early olivine, or a potentially an immiscible sulphide liquid phase. Groups B and C1 Ni values decrease only slightly through the fractionation index, but correlate with the Group C2 and F data. The inflection point at Mg# ~65 for these groups may indicate that olivine is no longer a fractionating phase. Groups A2 and D show shallow linear trends of decreasing Co with decreasing Mg# to Co values of ~50 ppm. Groups C2 and F show steep linear trends, with Group C2 decreasing more steeply with decreasing Mg# than Group F, to low values of Co (5 and 35 ppm, respectively). The depletion of Co in both Groups C2 and F suggests very

significant olivine fractionation, or depletion resulting from scavenging of chalcophile elements into an immiscible sulphide liquid.

Th contents in Group D and A1 increase with decreasing Mg#, while all other groups show significant scatter, with the exception of Group F which maintains relatively constant concentrations. La, Sm, and Y contents in all groups generally increase with decreasing Mg#, though Group C1 shows a high degree of scatter in La and Sm. Group C2 shows some significant scatter in La. Zr contents in all groups generally increase with decreasing Mg#, those of Group C1 show significant scatter. The V values in all groups increase with decreasing Mg#, with the exception of those values for Group C2 which remain relatively constant. Nb increases with decreasing Mg# in Groups A1, A2, D, and E, remain relatively constant Groups C2 and F, and show significant scatter in Group C1. Sc contents are highest in Groups A2, B, and E. Groups D and A2 show significant scatter in Sc concentrations, which very weakly increase with decreasing Mg#. Group F decreases with Mg# until Mg# 69, and increases steeply. Group C2 shows significant scatter and a weak negative trend with decreasing Mg#.

The HFSE concentrations in group D are significantly greater than those in Groups A1 and A2 (and all other groups) at constant Mg#. The only exceptions are V and Y, where Group D shows very similar concentrations to Groups A1 and A2 at constant Mg#. Groups B, C2 and F generally contain lower concentrations of HFSE compared to the rest of the groups. Concentrations in Group E are

variable with relation to the other groups. Again, Group C1 often displays a high degree of scatter and thus, highly variable concentrations as Mg# decreases.

7.1.4.2 Primitive mantle-normalized spider diagrams

Primitive mantle-normalized spider diagrams (Figure 7.3b) reveal some characteristics consistent across a number of geochemical groups, as well as some that are unique to a single geochemical group. A subset of incompatible elements were used in these diagrams to reduce scatter and to express the variations between the geochemical groups.

Group A1 displays a flat PM-normalized trace element pattern with prominent negative P and positive Zr-Hf anomalies that become less pronounced as total REE increases. La and Yb contents are ~2x primitive mantle.

Similar to Group A1, all Group A2 samples also contain strong negative P anomalies that become less prominent as total REE increases, however, Group A2 exhibits weak enrichment in the most incompatible trace elements. Nb-Ta anomalies in Group A2 vary from strongly negative (i.e. typical of lower crustal contamination) to weakly positive, and there are two minor positive and one moderately negative Ti anomaly. Additionally, there are very strongly positive, but mostly negative, Th anomalies. La and Yb contents vary from 5-11x and 2-8x primitive mantle, respectively.

Group B exhibits a moderate enrichment in the most incompatible trace elements, strong negative P anomalies and moderate negative Zr-Hf anomalies. The only samples with Nb and Ta concentrations above the detection limit have a

strong negative Nb-Ta anomaly. One sample has a moderate negative Ti anomaly, and another has a positive Ti anomaly of the same size. La and Yb contents vary from 2-11x and 0.5-3x primitive mantle, respectively, and overall these patterns exhibit variable (scattered) incompatible trace elements.

Group C1 exhibits a moderate to strong enrichment in the most incompatible trace elements. All samples show either positive or negative Ti anomalies that vary from weak to strong, and weak positive or negative Eu anomalies. Samples with the lowest trace element concentrations show moderate to strong negative P anomalies, while those with higher concentrations of trace elements are weak to moderate positive. Weak to moderate negative Zr-Hf anomalies are associated with the negative P anomalies, decreasing in size with increasing trace element abundance. Most samples with Nb-Ta abundances above detection limit have sizeable negative Nb-Ta anomalies. La and Yb contents vary from 7-40x and 0.6-2x primitive mantle, respectively. This group also shows less parallelism between patterns in the most incompatible trace elements.

Group C2 shows high enrichment in the most incompatible trace elements, and moderate to strong P anomalies in almost all samples. Samples with the lowest trace element abundance display moderate negative Zr-Hf anomalies. Samples with Nb-Ta abundances above detection limit have moderately to strongly negative Nb-Ta anomalies. La and Yb contents vary from 3-30x and 0.1-1x primitive mantle, respectively. Similar to what is observed in

Groups B and C1, these patterns are less parallel in the most incompatible trace elements.

Group D contains weak to moderate negative P anomalies in all but one sample, and also consistent weak to moderate positive Ti anomalies that become less pronounced as total REE contents increase. A highly distinguishing feature of Group D is the weak to moderate positive Nb-Ta anomalies. The Ti and Nb-Ta anomalies may reflect ilmenite cumulates, however, evidence of ilmenite accumulation is not clear in thin section. La and Yb contents vary from 8-60x and 2-5x primitive mantle, respectively.

Group E is unique in the sense that it displays only a range of different negative anomalies. These include moderate to strong negative Ti, Zr-Hf, Eu, P, and Ta (the latter only when above the detection limit). La and Yb contents vary from 5-6x and 4-7x primitive mantle, respectively.

Most Group F samples show weak negative Ti anomalies, and weak to moderate negative P anomalies. The presence of Nb-Ta anomalies cannot be determined due to the values for these elements falling beneath their respective detection limits. La and Yb contents vary from 0.4-1x and 0.7-2x primitive mantle, respectively.

7.1.4.3 MORB-normalized plots

Mid-ocean ridge basalt (MORB) is the most common mafic rock on Earth, and is often utilized for comparative purposes. MORB-normalized spider diagrams (Figure 7.3c) provide a different view of the relative abundance of

elements within the various groups. However, the MORB-normalized diagrams demonstrate very similar patterns to the primitive mantle-normalized diagrams, with anomalies approximately of the same magnitudes, with the exception of P anomalies, which disappear. The only notable differences are the lower degrees of overall enrichment relative to the normalizing values and greater enrichment in the most incompatible trace elements, since the MORB normalizing values are greater than primitive mantle normalizing values.

7.1.4.4 Immobile trace element ratio discrimination

Application of trace element ratios of the most immobile trace elements (HFSE) and REE, in the form of bivariate plots or ternary diagrams distinguishes the various geochemical groups. As previously stated, HFSE are considered generally immobile during metamorphism. Ratios of HFSE are not significantly modified by processes of fractional crystallization and partial melting (e.g. [Luttinen and Furnes, 2000](#)). Ternary diagrams in [Figure 7.9a-b](#) demonstrate the differences between the various geochemical groups. [Figure 7.9a](#) utilizes chondrite-normalized REEs (i.e. La, Gd, and Yb) as the discriminating variables, which more or less reflect the differences in the LREE (La/Sm), HREE (Gd/Yb) and overall slopes (La/Yb) of the REE. [Figure 7.9b](#) further illustrates the differences in geochemical composition between the eight geochemical groups, using Nb, Y and La. The latter diagram utilized the variables chosen for the [Cabanis and Lecolle \(1989\)](#) basalt-discrimination diagram. The fields from this

diagram are not used, as some samples do not necessarily correspond to them, due to either their ultramafic composition or cumulate nature.

In [Figure 7.9a](#), the trend from Group D up through to Group F may represent depth/degree of partial melting, based on the decrease in enrichment, apparent in chondrite-normalized REE diagrams ([Figure 7.3a](#)) and trace element bivariate diagrams ([Figure 7.8](#)). Closer to the La corner of the plot are Groups B, C1, and C2. Their separation from the other groups likely reflects contamination, as La would be expected to increase significantly and Nb would decrease, relatively, consistent with the generation of a negative Nb-Ta anomaly. Trends observed in [Figure 7.9b](#) also distinguish the different geochemical groups and reflect these changes in degrees of melting and contamination.

7.1.4.5 Alkaline versus tholeiitic

Typically, LIP-related tholeiitic magmas are produced from high-degrees of partial melting (often 20-30%) of near-anhydrous peridotitic mantle sources, while alkaline magmas are produced from small-degree partial melts of an enriched source or a source containing a high proportion of eclogitic or basaltic material (i.e. subducted oceanic slab components; [Ernst, 2014](#)). Eclogite geochemistry can have great influence on the geochemistry of low-degree partial melts, when present, since it melts far easier than peridotite ([Ernst, 2014](#)). Alkalinity is typically attributed to low degrees of partial melting, or source-enrichment. However, it may also be attributed to high-pressure melting where the opx-ol peritectic reaction may not occur ([J.Mungall, pers. comm., 2017](#))

The Nb/Y ratio is considered a reliable chemical separator of alkaline (>1) and tholeiitic (<1) magmas (Sun and McDonough, 1989). The Nb/Y diagram in Figure 7.10 (modified from Greenough et al., 2005) demonstrates difference in alkalinity between the different groups. The diagram indicates that all groups are tholeiitic, with the exception of Group D, which appears to be transitional from tholeiitic to mildly alkaline in nature. The observation of lower Mg# at similar elemental concentrations compared to other groups is consistent with the more alkaline nature of Group D, given that lower-degree partial melts are more FeO-rich relative to MgO. It should be noted that the Nb/Y ratio preserved in these rock may be affected by ilmenite accumulation (increasing the Nb/Y ratio), however, the lack of clear evidence of ilmenite accumulation suggests that the high Nb/Y ratio is a parental magmatic feature.

Experimental studies, such as those by Hirose and Kushiro (1993), have shown that at depths > 3 GPa, small degree melts (<5 wt.%) of peridotite can produce primary alkalic, basaltic magmas, and at larger degree of partial melting of the same source at shallow depths can produce tholeiitic series magmas. It may be suggested that based on the slight alkalinity of Group D, relative to the tholeiitic composition of the rest of the groups, the primary magma to the Group D samples may have been derived at a greater depth, with a lower degree of partial melting. This notion is further investigated below (see section 7.4.4).

7.2 Sm-Nd Isotopes Geochemistry

To assess the source characteristics of the parental magmas of the various geochemical groups, Sm-Nd isotope data were collected. The Sm-Nd isotopic system is comprised of parent ^{147}Sm that decays through alpha-emission to ^{143}Nd . Both Sm and Nd are incompatible elements that are part of the LREE, meaning that their concentrations increase with increasing degree of fractionation (Faure and Mensing, 2005). However, Nd is slightly more incompatible than Sm during melting due to its slightly larger ionic radius ($1.08\text{\AA}$ compared to $1.04\text{\AA}$). As a result, Nd becomes more concentrated in the melt relative to Sm during crystal fractionation, and this has led to generally low Sm/Nd ratios in crustal rocks compared to rocks derived from the upper mantle. Over time, crustal rocks have developed increasing lower $^{143}\text{Nd}/^{144}\text{Nd}$ ratios compared to the upper mantle. The reason this isotopic ratio works is because isotopes of the same element (e.g. ^{143}Nd , ^{144}Nd) cannot be fractionated through processes of crystal fractionation or partial melting, hence the ratios between them should reflect characteristics of the source from which the primary melts were extracted. However, the information extracted from this data is relatively limited compared to what can be revealed in conjunction with a second isotopic system (e.g. Rb-Sr, Pb-Pb).

$^{143}\text{Nd}/^{144}\text{Nd}$ isotopic ratios are most often expressed in epsilon (ϵ) units, which is a measure of the difference between the initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of the sample, and the corresponding ratio of the chondritic uniform reservoir (CHUR)

at the time of crystallization. The ϵ notation (DePaulo and Wasserburg, 1976) is used to resolve the issue of inter-laboratory differences in normalization procedures. When ϵNd values are positive, the $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic ratios are radiogenic, indicating that the magmas were derived from the depleted mantle. In contrast, negative ϵNd values indicate non-radiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic ratios, meaning that the magmas were derived from enriched mantle sources or continental crust. Crustal contamination of depleted mantle derived basaltic magmas will drive ϵNd to more negative values.

7.2.1 Epsilon-Nd (ϵNd)

Epsilon-Nd (ϵNd) values (at 1900 Ma) for Groups A1 (1.23 to 3.69), A2 (0.25 to 2.79), D (0.89 to 2.39) and F (4.58 to 6.13) are juvenile and reveal that they were derived from a depleted asthenospheric mantle. The exception is a Group A2 sample with ϵNd of -6.64 (15SAB-L057B02), suggesting it is significantly crustally-contaminated. The primitive mantle-normalized spider diagram of all samples collected from that particular sill supports this reasoning, as a significant negative Nb-Ta anomaly is observed in only that sample (Figure 7.11a). The ϵNd values suggest there is a common or similar source for Groups A1, A2 and D, while the source Group F appears to be even more depleted in nature. Samples from Groups B, C1, and C2 have evolved, less radiogenic ϵNd values -3.00, -4.31, and -3.99, respectively, suggesting that these rocks were derived from an enriched mantle source or had significant interaction with continental crust. The three ϵNd values from Groups B, C1, and C2 fall within a

close enough range to potentially suggest a common or similar source, however, their source characteristics are obscured by contamination from either crustal material, or metasomatized subcontinental lithosphere.

7.2.2 ϵNd versus $^{147}\text{Sm}/^{144}\text{Nd}$ diagram

In the ϵNd versus $^{147}\text{Sm}/^{144}\text{Nd}$ diagram (Figure 7.11b), there are two trends apparent amongst Groups A1, A2, and D. The first is interpreted to reflect changes in partial melting of the source, a process that influences the $^{147}\text{Sm}/^{144}\text{Nd}$ ratio. The $^{147}\text{Sm}/^{144}\text{Nd}$ ratios increase from Group D to A2 which is higher than in Group A1 because of differences in partial melting (lower degree of melting = less Sm in melt phase). ϵNd may be increasing with Sm/Nd because Group A1 is from a more depleted source. The second trend may reflect crustal contamination by the upper crustal rocks of the Meta Incognita microcontinent. This inference is supported by the ϵNd -Ti/Yb diagram (Figure 7.12) adapted from Leeman & Hawkesworth, (1986), which indicates that the low ϵNd values are a result of contamination, as opposed to source enrichment. Furthermore, samples apparently trending toward the fields of the Meta Incognita microcontinent upper crustal units (data from Theriault et al., 2001 and R.J. Theriault, unpublished data) in the ϵNd -Ti/Yb diagram exhibit greater La/Nb values on the ϵNd versus La/Nb diagram (Figure 7.17b), which is a proxy for crustal contamination (Lightfoot et al., 1993).

7.3 Geochemistry of Detailed Sample Transects Across Two Mafic-Ultramafic Sills

In the detailed sample sets collected from the Killapait and Chidliak sills, attempting to identify the phases controlling fractionation is challenging, as the primary minerals and textures have been subjected to subsequent amphibolite to granulite facies metamorphism. The limited number of samples also make it difficult to distinguish trends, however, some trends are observed that may represent fractionation of certain mineral phases.

In a study on the Golden Valley sill of the Karoo LIP ([Galerne et al., 2010](#)), it was revealed that a single saucer-shaped sill can exhibit significant geochemical variation through the stratigraphic thickness of an individual sill. This was proposed to be the result of post-emplacement porous melt flow driven by forces related to the saucer shape geometry. Consequently, if the sills were not emplaced in flat sheets, and exhibited more of a saucer shape, the following compositional profiles may be unique to the transect randomly chosen.

The geochemical variation through the Killapait sill is shown in [Figure 7.13](#), and the geochemical variation through the Chidliak sill is illustrated in [Figure 7.14](#) (both described in 7.4.1). The Killapait sills contains 4 different magma types (Groups A1, A2, C1 and C2), whereas the Chidliak sill only contains one (Group F). As a whole, there are no trends that appear to represent in-situ differentiation of a single magma batch, however, within certain lithological units some trends are exhibited that may have resulted from fractional

crystallization processes. However, the non-systematic change in geochemical signature through the stratigraphy of the sill suggests these sills represent either composite or a vertically-stacked package of sills (i.e. distinct sills emplaced within a host sill or distinct sills which have underplated one another, respectively).

7.4 Interpretation

7.4.1 Single injection versus multiple injection origin of layered sills

The occurrence of various combinations of geochemical groups within, what appeared to be individual differentiated sills while in the field, poses the question of whether these sills formed through the injection of a single sill which later differentiated, or whether they represent a much more complex history, whereby multiple sills were injected.

Recent geochronology has revealed small age differences between ultramafic and more mafic portions of both the Bushveld Complex ([Scoates et al., 2012](#); [Mungall et al., 2017](#)) and Stillwater Complex ([Wall et al., 2012](#)). Age differences between the two portions of the sill in each case are ~2-5 Ma. In the case of the Frobisher suite, sills that exhibit distinctly different trace-element signatures may represent the injection of multiple pulses of distinct magma batches, of unique magmatic lineage.

Comparative geochemistry between the mafic and ultramafic portions of individual sills, initially thought to have been derived through in situ differentiation of single magma batches, reveal geochemical variations that do not reflect

change through simple crystal fractionation. The chondrite-normalized REE patterns for the metapyroxenitic and metaperidotitic units of two individual layered ultramafic sills are presented (Figure 7.15a), along with the REE patterns for two individual non-layered ultramafic sills. What this diagram reveals is that two different geochemical signatures are present within these layered ultramafic sills (Groups A1 and D). The process of fractional crystallization, which only slightly increases LREE contents relative to HREE contents as the magma evolves, cannot explain the transition from the relatively flat Group A1 pattern to the highly LREE-enriched Group D pattern without significantly greater Yb contents in Group D, which is not observed. It seems highly plausible that these layered ultramafic sills were generated through at least two injections of distinct geochemically different magmas.

In Figure 7.15b, the chondrite-normalized REE patterns for the layered mafic-ultramafic Killapait sill are shown, along with sample location with respect to stratigraphic height within the sill. In a layered sill derived from closed-system in-situ crystal fractionation from a single magma pulse, it would be expected that HREE contents would generally increase from the base upwards through the sill and the chondrite-normalized REE patterns would remain same, with the exception of progressive, but minor, increases in the LREE slope as the samples evolve. Essentially, the lowest abundance of HREE through to the highest abundance of HREE should reflect the sample location from stratigraphically lowest to highest in the sill. Instead, what is observed are four different geochemical groups, with unique geochemical characteristics, and no specific

correlation between increasing HREE concentrations and height in sill stratigraphy. For example, a Group A2 sample collected ~40 m above the contact has significantly higher HREE contents than the rest of the samples collected from the sill. The samples collected at the basal contact (~5 m) and proximal to an obscure upper contact (~445 m), 14SAB-R010A01 and 14SAB-L017A01, respectively, contain intermediate HREE contents with respect to the rest of the samples, and appear to represent two distinctly different magma types. The sample with the lowest HREE content is situated ~90 m upwards through magmatic stratigraphy from the basal contact. Furthermore, the Group A2 signature is found at a higher stratigraphic level within the ultramafic portion of Killapait sill than the Group A1 signature, which occurs at the basal contact. Given the significantly higher Yb values of the Group A2 sample, the difference in patterns may simply reflect fractional crystallization. However, stratigraphically higher in the sill, the mafic portion of the sill contains the geochemical signatures of Groups C1 and C2. The Yb contents of the samples selected from these two groups are generally lower than that of the Group A1 sample found at the basal contact, precluding simple fractional crystallization as an origin of the former signatures. However, if significant assimilation occurred within Group C1 and C2 parental magmas, there is an argument for their derivation through assimilation-fractional-crystallization (AFC) processes.

Furthermore, while mafic and ultramafic components of a single sill may exhibit similar geochemical signatures, this does not indicate that it was differentiated in situ. For example, the metaperidotitic and gabbroic components

of the Chidliak sill exhibit REE patterns that are very similar (Figure 7.15c), with the gabbroic unit containing slightly higher Yb (and overall REE) contents than the metaperidotitic sample, consistent with the gabbro's origin through fractional crystallization. However, the Sm-Nd isotopic data obtained from both units suggest that they were derived from isotopically distinct magma batches, as the epsilon Nd values of the metaperidotitic and metagabbroic units are 4.58 and 6.13, respectively. If derived from the same source, minor contamination of the parental magma to the metaperidotite, which would reduce the ϵNd , could explain the different ϵNd values. Nb anomalies are present in both of these samples, however, they are similar in magnitude. A difference in the isotopic composition of the respective contaminants could potentially explain this. Another possibility is that the parental magmas to the two units have similar trace element characteristics, while representing two distinct sources.

All the geochemical evidence presented supports the notion that at least some of the layered intrusions of Frobisher suite may have been formed through multiple injections of geochemically similar or distinct magma types, as opposed to a single injection, that may or may not have differentiated in-situ into mafic and ultramafic components.

7.4.2 Mineral fractionation

The crystallization of minerals has a significant impact on the composition of any primary magma, leading to either significant accumulation or depletion of specific geochemical entities within the magma. In mafic or ultramafic magmas,

there is typically a decrease in MgO content, and concomitant increase in FeO, as Mg-rich ferromagnesian minerals are crystallized (decreasing the Mg#). A number of fractionating mineral phases have high partition coefficients ($\gg 1$) for certain elements (e.g. Ni in olivine or sulphides, or Eu^{2+} in plagioclase), which leads to the depletion of these elements as the magma evolves and Mg# decreases (Figures 7.5 and 7.8). For example, the low Mg#s and relatively low compatible element contents characteristic of Group A2 suggest that the compositions are far removed from that of the primary magma, presumably due to extensive fractional crystallization.

In the determination of mineral phases controlling fractionation, we are handicapped by the effects of regional high-grade metamorphism, which appears to have obliterated primary minerals and primary igneous textures. However, a significant amount of information can be obtained by studying the geochemistry of the samples alone.

7.4.2.1 Olivine and pyroxene fractionation

Samples with the highest Mg#s (> 70), which includes primarily samples from Group A1 and F, may represent cumulates of olivine or pyroxene because of the very high Ni and Co contents, and low total REE contents (Figure 7.8). This may partially explain the large gap in Mg# between Group A1 and A2. The petrographic observation of a relict pyroxene (pseudomorphed by hornblende) cumulate texture in sample 14SAB-H080B01 supports this notion (Figure 5.3b). Cumulates usually have higher MgO and low incompatible trace elements than

their parental magma (Maier, 2015) Additionally, these samples are limited to an ultramafic composition. This high Ni is not attributed to the presence of Ni-sulphide, as only one of six Group A1 samples contains any amount (i.e. <1%) of Ni-sulphide (i.e. pentlandite). The high Mg# Group A1 samples have a narrow range in Mg# and compatible elements Ni and Cr, which may possibly be due to high-MgO olivine or pyroxene of similar composition.

High Cr contents of some samples in other groups may indicate accumulation of orthopyroxene, clinopyroxene or chromite. In a similar manner, high concentrations of Ni in some samples may indicate accumulation of olivine or Ni-sulphide.

7.4.2.2 Plagioclase fractionation

The REE are all trivalent (3+), with the exception of Ce²⁺ and Eu²⁺ which form under reducing conditions. When Eu is in its divalent state (2+), it is preferentially incorporated into plagioclase, where it substitutes for Ca²⁺. As a result, when plagioclase is accumulated within a magma or crystallized and removed from a magma, positive or negative Eu anomalies will be present in chondrite-normalized diagrams. Furthermore, a negative Eu anomaly may indicate that the melt was in equilibrium with plagioclase at one time, however, it cannot distinguish whether plagioclase phenocrysts were removed from the magma or whether plagioclase remained in the source during melting; the latter only occurs at very low pressures. Another typically incompatible element that accumulates in plagioclase is Sr.

An additional explanation for the presence of minor Eu anomalies is regional metamorphism. While Eu^{3+} is considered largely immobile, Eu^{2+} is considered a LILE and thus, can be remobilized relatively easily compared to its 3+ charge counterpart. Its effects could be most easily recognized in samples where inconsistencies exist, such as plagioclase-barren samples with positive Eu anomalies. In this example, the positive Eu anomaly would indicate accumulated plagioclase. However, the absence of plagioclase in this hypothetical situation suggests that the anomaly was generated by other processes (e.g. high-grade metamorphism). Furthermore, care must be exercised in the attribution of these anomalies to metamorphism and element mobility, as secondary mineral assemblages can mask the presence of primary plagioclase in the now non-existent magmatic mineral assemblage.

Samples from Groups A1, A2, and D show weak positive and negative Eu anomalies with Eu/Eu^* typically between 0.80 and 1.20 (Figure 7.16), indicating that the fractionation of plagioclase was not particularly important in their petrogenesis. In contrast, samples from Groups B, C1 and C2 show significant, mostly positive, europium anomalies. Group C2 samples have the highest positive europium anomalies, with Eu/Eu^* up to 3.25. The variation in Eu/Eu^* values versus $\text{Mg\#}$ is illustrated in Figure 7.16. These significant positive anomalies coincide with a notable accumulation of plagioclase. The most magnesian Group C2 and B samples contain incompatible element concentrations lower than primitive mantle, while maintaining enriched LREE patterns, suggesting they contain a cumulate component (Sills et al., 1982).

While Sr is a LILE and shows some evidence of remobilization (not shown), the appearance of coexisting positive Sr and Eu anomalies and coexistence of negative Sr and Eu anomalies supports the interpretation that Eu anomalies represent the accumulation or removal of plagioclase within the parental magmas to these samples.

Within the Killapait sill, all Group C1 and C2 samples, with the exception of one, display significant positive Eu anomalies (Figure 7.15b). The interval through which these samples were collected represents the > ~90% of the sill's total volume, assuming uniform unit thickness during its formation, suggesting that the majority of the magma injected into the sill had accumulated plagioclase, possibly through entrainment of a plagioclase crystal mush en route to the site of final emplacement, or through plagioclase accumulation during in situ crystal fractionation within a crustal magma chamber.

7.4.2.3 Apatite fractionation

All samples from all groups, with one exception, exhibit negative P anomalies in PM-normalized diagrams that vary from weak to strong (Figure 7.3b). However, P anomalies are typically absent in MORB-normalized diagrams of Groups A1, A2, D, and F (Figure 7.3c). As samples become more evolved, there is a progressive decrease in the P anomalies exhibited by samples. Possible explanations for negative anomalies were explored and synthesized by McCuaig et al. (1994), and include: (1) crustal contamination; (2) hydrous metasomatism of depleted mantle; (3) crystal fractionation of P-bearing phase(s)

from the magma; and (4) retention of P-bearing phases in the mantle source region during partial melting.

If the negative P anomalies were derived from either (1) crustal contamination or (2) hydrous metasomatism of the depleted mantle, negative Nb-Ta and Ti anomalies would be expected to coincide with these negative P anomalies. Group D contains negative P anomalies with coexisting positive Nb-Ta and Ti anomalies, and Groups A1 and A2 do contain some Nb-Ta anomalies but most often lack significant Ti anomalies. Additionally, ϵNd values of samples from Groups A1, A2 and D would all be negative (enriched source), however, isotopic data indicates they are derived from a depleted source. Consequently, the possibility of (1) and (2) being responsible for negative P anomalies in Groups A1, A2 and D is quite low. Groups C1, C2, and F display coexisting negative Nb, and often negative Ti anomalies, with negative P anomalies, suggesting that crustal contamination or metasomatized mantle may have led to these anomalies. Group E displays negative Ta and Ti anomalies with negative phosphorous anomalies, also suggesting a similar processes to (1) and (2).

The remaining possibilities for Groups A1, A2 and D are now (3) and (4), but it is difficult to decipher between fractional crystallization of P-bearing phases and retention of P in the mantle source. The most significant P-bearing repositories in high-MgO rocks and the upper mantle include fluoroapatite and whitlockite ([Beswick & Carmichael, 1978](#); [Watson, 1980](#)). However, apatite is one of the last minerals to crystallize from a mafic magma and it is nearly impossible to retain apatite in the mantle source during partial melting of the source

([Watson, 1980](#)). An alternative explanation may be that the P data is simply just bad data, as the concentrations fall very close to the detection limit.

7.4.2.4 Fe-Ti-oxide fractionation

Increasing TiO_2 and V with fractionation (i.e., decreasing Mg#) is observed in all groups except for Group C2 (not shown), and is characteristic of tholeiitic suites ([Myashiro and Shido, 1975](#)). It is interpreted to reflect low oxygen fugacities in the magma which suppresses magnetite and ilmenite fractionation ([Osborn, 1962](#)). However, there are still many samples that bear Ti anomalies.

Those samples exhibiting negative Ti anomalies (all or some of Groups A2, B, C1, C2, E and F) in spider diagrams provides potential evidence for removal of (Fe-)Ti-oxides (e.g. rutile) or Ti-bearing amphibole, whereas those with positive anomalies (all or some of Group A2, C1, and D) may have accumulated Fe-Ti-oxides and Ti-bearing amphibole (e.g., [Ionov and Hofmann, 1995](#); [Tiepolo et al., 2000,2001](#); [Klemme et al., 2005, 2006](#); [Coltorti et al., 2007](#);). Alternatively, a negative Ti anomaly could reflect residual (Fe-)Ti-oxides or Ti-bearing amphibole in the source. Group A1 lacks a Ti anomaly and hence does not appear to have accumulated or fractionated away Fe-Ti-oxides. All samples from high-Ti Group D contain weak positive Ti anomalies, which may reflect presence of magnetite or other (Fe-)Ti-oxides. Select samples from Groups B and C1 also contain moderate positive Ti anomalies.

Another possibility is that the partial melting may have been water-saturated, as this would allow for rutile to be stable in the source, accounting for the HFSE depletion (Prouteau et al., 1999) in some but not all cases.

According to Pearce (1996), Ti anomalies are an effective indicator of the extent of fractionation, as the negative Ti anomaly typically increases with progressive fractionation. Most crystallizing phases in basaltic magma do not influence Ti/Zr, however, when magnetite becomes a fractionating phase, Ti is progressively depleted while Zr continues to increase (decreasing Ti/Zr ratio). The latter only seems to be the case for samples from Groups B and C1.

The relatively constant values of V with decreasing Mg# in Groups C1 and C2 could be due to fractionation of rutile, clinopyroxene, or magnetite.

7.4.3 Contamination

The presence of a negative Nb anomaly, along with other negative HFSE anomalies (e.g. Hf, Zr, Ta), is a well-established feature of many LIP magmas (Bertrand, 1991; Arndt et al., 1993; Molzahn et al., 1996; Puffer, 2001; Iacumin et al., 2003; Ewart et al., 2004). These anomalies are interpreted to reflect either (1) crustal assimilation by mantle-derived magmas (Arndt et al., 1993; Taylor & McLennan, 1995); (2) a mantle source that was previously modified by metasomatic enrichment induced by the percolation of hydrous fluids or silicate melt, which may reflect ancient subduction (Duncan et al., 1984; Puffer et al., 2001; Ewart et al., 2004); or (3) interaction of plume- or asthenosphere-derived magmas with lithospheric mantle (Arndt & Christensen, 1992). It is important to

note that in all three scenarios, the Nb anomaly was inherited within the lithosphere. Given that sills (and dykes) represent the plumbing system to LIPs, these same processes should account for these anomalies observed in mafic to ultramafic intrusions. In addition to the isotopic evidence for contamination already presented (Figures 7.11 & 7.12), certain trace element ratios reveal further evidence for the influence of lithospheric material on the chemical compositions of most groups, and potential contaminant sources [i.e. sub-continental lithospheric mantle (SCLM) versus continental crust].

The ratio La/Nb is a well-established proxy for the contamination of mafic-ultramafic magmas. Pearce (2008) also highlighted its usefulness for monitoring crustal contamination, as well as Ti/Yb for depth of melting. These two ratios were used in Figure 7.17a to identify samples that have been contaminated. Samples with La/Nb <1 are considered to have been uncontaminated and this field is typical of OIB, kimberlites, and continental alkali basalts, while those samples plotting above the line are considered to have experienced variable amounts of contamination; LIPs plot between 0.5 to 7 (Wilson, 1989). It is apparent that all samples from all groups, with the exception of most samples of Group D, some of Group A2 and one of Group F, experienced at least some degree of contamination. The contamination in Groups A1, A2, E, and F appears to be relative minor (La/Nb = ~1-3) compared to that of Groups B, C1, and C2 (La/Nb = ~3-20). Based on the relatively high La/Nb ratios in the ϵNd versus La/Nb diagram (Figure 7.17b), it is suggested the low ϵNd values of Groups B, C1 and C2 are due to contamination.

To demonstrate that the LREE-enriched character of Group D is due to source enrichment (i.e. ancient subducted oceanic crust component), rather than crustal contamination, the diagram $(\text{Th}/\text{Yb})_{\text{PM}}$ versus $(\text{Nb}/\text{Yb})_{\text{PM}}$ is used (Figure 7.18), which segregates enrichment arising from crustal contamination, from enrichment due to enrichment of the source. The result is Group D samples demonstrating a clear trend of source enrichment, while samples from Groups B, C1 and C2 exhibit a trend typically caused by crustal contamination.

In attempting to distinguish between crustal and SCLM contamination, the ratios Ti/Yb and Nb/La , when plotted against La/Sm (Figures 7.19a & 7.19b, respectively), should produce a correlation if the development of Ti and Nb-Ta anomalies are due to lithospheric contributions (Condie, 1997). The correlation results from the negative Nb and Ti anomalies that accompany crustal contamination-induced LREE-enrichment. The weak negative correlation produced by Group C2 suggests that the geochemical signature may have been produced through crustal contamination, whereas Groups B and C1 exhibit constant La/Sm with decreasing Ti/Yb and Nb/La , perhaps indicating lithospheric contamination. Furthermore, Groups A2 and F do produce a slight correlation consistent with the crustal contamination of these two groups. The decrease in Nb/La ratio with constant La/Sm (a ratio that should increase with increasing upper crustal contamination or moderate-degree lower crust contamination) suggests Group A2 incorporated various quantities of lithospheric material. The trends exhibited by Groups B and C1 also cannot be produced by upper crustal

contamination. The correlation of near-constant La/Sm with decreasing Nb/La is interpreted to reflect assimilation of SCLM melts depleted in HFSE.

Additionally, if the Nb depletion in these samples is due to crustal contamination, then a high La/Sm ratio without depletion in Yb would be expected. What is observed in Groups B, C1 and C2 is a depletion in Yb relative to the rest of the groups; Yb = 0.4 to 9x chondrite in Groups B, C1 and C2, compared to Yb = 7 to 10x chondrite in all other groups. This may suggest that the latter three groups contain a SCLM component.

Crustal contamination may also be indicated by negative Nb-Ta anomalies in the absence of Zr-Hf anomalies, along with positive (La-Yb)_N (Condie, 2001). This is observed in the PM-normalized diagrams (Figure 7.3b) of Groups A1, A2 and F, but negative Zr-Hf anomalies are present in most samples from Groups B, C1, C2 and E.

In contrast, positive Nb anomalies, may be derived from a residual eclogitic component of a partially-melted subducting oceanic slab (which releases LREE and Th into the overlying mantle; Saunders, 1996; Sobolev et al., 2011). This would infer the recycling of an eclogitic slab in its mantle source.

Furthermore, the strong resemblance between Group C2 and both the Ramsay River Orthogneiss and Lake Harbour Group (Figure 7.20) support the idea that this group may have been subjected to significant crustal contamination. However, the SiO₂ contents of this group are generally quite low (~35-50 wt.%), which opposes the suggestion that significant amounts of SiO₂-rich crustal material were incorporated into the Group C2 parental magma. The

low SiO_2 character could also be generated by olivine cumulates, which compositionally have low SiO_2 . However, mafic units also exhibit this pattern, meaning it cannot be the result of olivine cumulates. Therefore, Group C2 is considered to either represent (or contain a significant amount of) direct melt from metasomatized subcontinental lithospheric mantle (SCLM).

7.4.4 Depth and degree of partial melting

In order to understand the petrogenesis of the magmas which fed the Frobisher suite mafic to ultramafic sills, a general idea of the depth and degree of melting must be known. Certain ratios have been utilized as proxies for depth and degree of partial melting, including Gd/Yb and Ti/Yb , and La/Sm , respectively. These ratios essentially reflect the slope of the most compatible, incompatible trace elements or REE, which changes with depth (and degree) of melting. Groups B, C1, and C2 are not plotted in this section due to the fact they appear to be contaminated.

The $(\text{Gd/Yb})_N$ versus $(\text{La/Sm})_N$ diagram (Figure 7.21) from Álvero et al. (2014) defines garnet- and spinel-stability fields based on the primitive mantle-normalized HREE ratio of Gd/Yb , which increases with increasing depth. All Group D samples plot within the garnet-stability field, with the exception of 1 sample. This indicates that the primary magmas to this group melted at great depths, where garnet is stable as the aluminous phase within mantle peridotite (consistent with other figures, e.g. chondrite-normalized REE diagrams). All samples from Groups A1, A2, E and F plot well-within the spinel stability field,

indicating a much a higher degree of melting at shallower depths in comparison to the Group D samples.

An alternate diagram of $(\text{Tb/Yb})_{\text{PM}}$ versus $(\text{La/Sm})_{\text{PM}}$ (Figure 7.22) by Wang (2002) further supports these observations in the aforementioned diagram, however, significantly fewer samples from Group D plot within the garnet-stability field. But nonetheless, it supports the indication of a greater depth melting in the petrogenesis of this group, compared to the groups that fall strictly within the spinel-stability field.

The $(\text{La/Sm})_{\text{PM}}$ ratios of Group D are higher than those of Groups A1 and A2, indicating these melts were derived from a lower-degree of partial melting. The absence of Nb-Ta anomalies in Group D confirm that the higher $(\text{La/Sm})_{\text{PM}}$ ratio is not a result of contamination. Furthermore, Group A2 has higher $(\text{La/Sm})_{\text{PM}}$ ratios than Group A1, but same $(\text{Gd/Yb})_{\text{PM}}$, which supports a higher degree of melting in the generation of Group A1 magmas compared to Group A2. Alternatively, this may suggest Group A2 is a more evolved, and thus more fractionated, magma of Group A1.

The distinct difference between the major low-Ti (Groups A1 & A2) and high-Ti (Group D) groups in their contrasting trace element patterns (Figure 7.3a-c), specifically, the fractionation systematics of MREE/HREE, should be highlighted. While the two groups show considerable overlap in LREE/MREE ratios (Figure 7.4), the MREE/HREE ratios of the high-Ti group are systematically higher and exhibit much more variation $[(\text{Sm/Yb})_{\text{N}} = \sim 2 - 3.8]$ than those of the low-Ti group $[(\text{Sm/Yb})_{\text{N}} = \sim 1.2 - 2.7]$. The $(\text{Sm/Yb})_{\text{N}}$ ratios do not correlate with

Mg#, indicating that such variations cannot be explained by crustal contamination or fractional crystallization, which would have a very minor influence on this ratio. Furthermore, these ratios indicate different depths of melting, where the high $(\text{Sm/Yb})_N$ ratio of the high-Ti group reflects the presence of significant amounts of garnet in the source, whereas the low $(\text{Sm/Yb})_N$ ratios (nearly flat trend) of the low-Ti group excludes garnet in the source, indicating a spinel-bearing source. These differences in $(\text{Sm/Yb})_N$ correspond to deeper (>90 km) and shallower (<90 km) depths of melting, respectively. This is because Sm is incompatible in mineral phases in spinel and garnet peridotite; however, Yb is compatible ($K_d > 3$) in garnet. (Thirlwall et al., 1994). Furthermore, these interpretations are supported by diagrams from Wang (2002) and Álvaro et al (2014), which outline the garnet and spinel peridotite melting fields.

The interpretation of a lower degree of partial melting for Group D (and Groups C1 & C2) is also supported by the enrichment in LREE and other incompatible elements (Figures 7.3, 7.4, and 7.7), which is typical of magmas derived from lower degrees of partial melting. In a similar manner, higher degrees of partial melting lead to more even abundances of LREE and HREE, or even a depletion in LREE relative to HREE, which is observed in Groups A1 and A2 (and Groups B, E & F). However, relatively flat REE patterns may be produced by low-degree partial melts of a depleted source (i.e. MORB).

While none of the eight geochemical groups are likely representative of a primary magma (i.e. direct melt of mantle source), these trace element ratios should not have changed significantly through crystal fractionation processes in

the crust. Note that contamination may have affected the $(\text{La/Yb})_N$ ratios of Groups B, C1 and C2 to a higher degree, but the depth/degree of melting results are still consistent with the rest of the observations, except for isotope data.

7.4.5 Magma mixing

7.4.5.1 Group A1-like Group A2 samples

Basic modelling (Figure 7.24) was undertaken in order to determine the origin of the only two ultramafic Group A2 samples (assigned based on similar chondrite-normalized REE patterns (Figure 7.3a; 14SAB-S130A01 & 14SAB-L001A01) that have an apparent affinity to Group A1 geochemistry, including concentrations of compatible and incompatible elements and degree of fractionation, but are distinct from the rest of the Group A2 samples, which are mafic in composition. Mixing was conducted to produce the pattern of 14SAB-L001A01, while the same origin is suggested for 14SAB-S130A01 by their similar chemistries (see all previous figures).

There is evidence for some mixing between Groups A1 and D, which may have produced two samples that were assigned to Group A2 on the basis of their similar chondrite-normalized REE patterns. For example, within the ϵNd versus $^{147}\text{Sm}/^{144}\text{Nd}$ diagram, 14SAB-S130A01 plots within the field of Group D samples (Figure 7.11a), whereas the rest of Group A2 and A1 samples display greater $^{147}\text{Sm}/^{144}\text{Nd}$. The degree of LREE enrichment compared to Group A1 samples of similar composition is not likely to have been produced by crystal fractionation, considering the very small difference in Mg# (<4). In the La/Sm versus Yb

diagram (Figure 7.23), these two samples have similar Yb values to Group A1 samples, which are significantly lower in Yb than the rest of Group A2 samples. However, the La/Sm ratios are higher and very similar to Group A2 samples in these samples.

Another possibility for this enrichment, given that it is similar to Group A1 samples, is a difference in partial melting. This would likely result in different Gd/Yb ratios from those samples in Group A1, and marginally higher Gd/Yb ratios are observed in these two samples.

Using the mixing feature in Igpet, 70% of Group A1 sample 14SAB-L003A01 was mixed with 30% of Group D sample 14SAB-L009A02, which produced a chondrite-normalized REE pattern very similar to Group A2 sample 14SAB-L001A01 (Figure 7.24). This is not conclusive evidence, but it illustrates the plausibility that the high-MgO-Cr-Ni Group A2 samples were produced by mixing between magma of Groups A1 and D.

References

- Álvaro, J. J., Pouclet, A., Ezzouhairi, H., Soulaïmani, A., Bouougri, E. H., Imaz, A. G., and Fekkak, A. (2014). Early Neoproterozoic rift-related magmatism in the Anti-Atlas margin of the West African craton, Morocco: Precambrian Research, v. 255, Part 1, p. 433-442. <http://dx.doi.org/10.1016/j.precamres.2014.10.008>.
- Arndt, N.T. (1994). Archean komatiites. In: Condie, K.C. (Ed.), Archean Crustal Evolution, Developments in Precambrian Geology, vol. 11. Elsevier, Amsterdam, pp. 11–44.
- Arndt, N.T. (2008). Komatiite. Cambridge University Press, Cambridge.
- Arndt, N.T. & Christensen, U. (1992). Role of lithospheric mantle in continental volcanism: thermal and geochemical constraints. Journal of Geophysical Research, 97: 10,967–10,981.
- Arndt, N. T., Czamanske, G. K., Wooden, J. L. & Fedorenko, V. A. (1993). Mantle and crustal contributions to continental flood volcanism. Tectonophysics 223, 39-52.
- Arndt, N., Menzies, M.A. (2005). The Ethiopian Large Igneous Province. January 2005 LIP of the Month. <http://www.largeigneousprovinces.org/05jan>.
- Arndt, N.T., Nesbitt, R.W. (1982). Geochemistry of Munro Township basalts. In: Arndt, N.T., Nisbet, E.G. (Eds.), Komatiites. George Allen and Unwin, London, pp. 19–28.
- Barley, M.E., Kerrich, R., Reudavy, I., Xie, Q. (2000). Late Archean Ti-rich, Al-depleted komatiites and komatiitic volcanoclastic rocks from the Murchison

-
- Terrane in Western Australia. *Australian Journal of Earth Sciences* 47, 873–883.
- Bertrand, H. (1991). The Mesozoic tholeiitic province of Northwest Africa: a volcano-tectonic record of the early opening of Central Atlantic. In: Kampunzu, A. B. & Lubala, R. T. (eds) *Magmatism in extensional structural settings*. Berlin: Springer, pp. 147-188.
- Beswick, A.E. and Carmichael, I.S.E. (1978). Constraints on mantle source compositions imposed by phosphorous and the rare-earth elements, *Contributions to Mineralogy and Petrology* 67, pp. 317-330.
- Cabanis, B. and Lecolle, M. (1989). Le diagramme La/10-Y/15-Nb/8: un outil pour la discrimination des series volcaniques et la mise en evidence des processus de melange et /ou de contamination crustale. *Comptes Rendus de l'Academie des Sciences Series II* 309, 2023-2029.
- Cann, J.R. (1970). Rb, Sr, Y, Zr and Nb in some ocean floor basaltic rocks, *Earth and Planetary Science Letters*, vol.10, pp. 7-11.
- Carlson, R.W. (1991). Physical and chemical evidence on the cause and source characteristics of flood basalt volcanism. *Australian Journal of Earth Sciences*, 38: 525–544.
- Coltorti, M., Bonadiman, C., Faccini, B., Grégoire, M., O'Reilly, S.Y. and Powell, W. (2007). Amphiboles from suprasubduction and intraplate lithospheric mantle. *Lithos* 99, 68–84.
- Condie, K.C. (1976). Trace-element geochemistry of Archean greenstone belts. *Earth-Science Reviews*, 12, 393–417.

-
- Condie, K.C. (1981a). *Archean Greenstone Belts*. Elsevier, Amsterdam–Oxford–New York. 275–311 pp.
- Condie, K.C. (1981b). *Archean Greenstone Belts*. Elsevier, Amsterdam, pp. 434.
- Condie, K.C. (1994). Greenstones through time. In: Condie, K.C. (Ed.), *Archean Crustal Evolution*. Elsevier, Amsterdam, pp. 85–120 (Chapter 3).
- Condie, K.C. (1997). Sources of Proterozoic mafic dyke swarms: constraints from Th/Ta and La/Yb ratios. *Precambrian Research*. 81: 3–14.
- Condie, K.C., 2001, *Mantle plumes and their record in Earth history*: Cambridge, Cambridge University Press, pp. 306
- DePaolo, D.J. and Wasserburg, G.J. (1976). Nd isotopic variation and petrogenetic models. *Geophysical Research Letters* 3, 249-252.
- Duncan, A. R., Erlank, A. J. & Marsh, J. S. (1984). Regional geochemistry of the Karoo igneous province. In: Erlank, A. J. (ed.) *Petrogenesis of the Volcanic Rocks of the Karoo Province*. Special Publication of the Geological Society of South Africa 13, 355-388.
- Duncan R.A., Hooper P.R., Rehacek J., Marsh J.S. and Duncan A.R. (1997). The timing and duration of the Karoo igneous event, southern Gondwana: *Journal of Geophysical Research*, v. 102, p. 18127–18138, doi: 10.1029/97JB00972.
- Ewart, A., Milner, S.C., Armstrong, R.A. and Duncan, A.R. (1998). Etendeka volcanism of the Goboboseb Mountains and Messum Igneous Complex, Namibia. Part I: geochemical evidence of Early Cretaceous Tristan plume

-
- melts and the role of crustal contamination in the Paraná–Etendeka CFB. *Journal of Petrology*, 39: 191–225.
- Ewart, A., Marsh, J. S., Milner, S. C., Duncan, A. R., Kamber, B. S. & Armstrong, R. A. (2004). Petrology and geochemistry of Early Cretaceous bimodal continental flood volcanism of the NW Etendeka, Namibia. Part 1: Introduction, mafic lavas and re-evaluation of mantle source components. *Journal of Petrology* 45, 59-105.
- Faure, G. and Mensing, T.M. (2005). *Isotopes*. John Wiley & Sons, New Jersey, pp. 897.
- Galerie, C.Y., Neumann, E.R., Aarnes, I., Planke, S. (2010). Magmatic differentiation processes in saucer-shaped sills: evidence from the Golden Valley Sill in the Karoo Basin, South Africa. *Geosphere*, 6, pp. 163–188 (Special Issue LASI III Conference).
- Greenough, J.D., Dostal, J., & Mallory-Greenough, L.M. (2005). Igneous rock associations 5. Oceanic island volcanism II: mantle processes. *Geoscience Canada*, 32: 77-90.
- Hanski, E., Huhma, H., Rastas, P., Kamenetsky, V.S. (2001). The Palaeoproterozoic komatiite-picrite association of Finnish Lapland. *Journal of Petrology* 42, pp. 855–876.
- Hirose, K. and Kushiro, I. (1993). Partial Melting of Dry Peridotites at High Pressures: Determination of Compositions of Melts Segregated from Peridotite Using Aggregates of Diamond. *Earth and Planetary Science Letters*, 114, 477-489. [https://doi.org/10.1016/0012-821X\(93\)90077-M](https://doi.org/10.1016/0012-821X(93)90077-M)

-
- Hooper P.R. (1982). The Columbia River Basalts: Science, v. **215**, p. 1463–1468, doi:10.1126/science.215.4539.1463.
- Iacumin, M., De Min, A., Piccirillo, E. M. & Bellieni, G. (2003). Source mantle heterogeneity and its role in the genesis of Late Archaean-Proterozoic (27–10 Ga) and Mesozoic (200 and 130 Ma) tholeiitic magmatism in the South American Platform. *Earth-Science Reviews* 62, 365–397.
- Ionov, D.A., Hofmann, A.W., (1995). Nb–Ta-rich mantle amphiboles and micas: implications for subduction-related metasomatic trace element fractionations. *Earth Planet. Sci. Lett.* 131, 341–356.
- Jochum, K.P., Arndt, N.T., Hofmann, A.W. (1991). Nb–Th–La in komatiites and basalts: constraints on komatiite petrogenesis and mantle evolution. *Earth and Planetary Science Letters* 107, 272–289.
- Klemme, S., Gunther, D., Hametner, K., Prowatke, S., Zack, T., 2006. The partitioning of trace elements between ilmenite, ulvöspinel, armalcolite and silicate melts with implications for the early differentiation of the moon. *Chemical Geology*, 234, 251–263.
- Klemme, S., Prowatke, S., Hametner, K., Gunther, D. (2005). Partitioning of trace elements between rutile and silicate melts: implications for subduction zones. *Geochimica et Cosmochimica Acta* 69, 2361–2371.
- Lafleche, M.R., Dupuy, C., Dostal, J. (1992a). Tholeiitic volcanic rocks of the Late Archean Blake River Group, Southern Abitibi greenstone belt: origin and geodynamic implications. *Canadian Journal of Earth Sciences* 29, 1448–1458.

- Leeman, W.P., and Hawkesworth, C.J. (1986). Open magmatic systems - some isotopic and trace element constraints. *Journal of Geophysical Research* 91, 5901-5912.
- Leshner, C.M., Arndt, N.T. (1995). REE and Nd isotope geochemistry, petrogenesis and volcanic evolution of contaminated komatiites at Kambalda, Western Australia. *Lithos* 34, 127–157.
- Leshner, C. M., Burnham, O. M., Keays, R. R., Barnes, S. J., and Hulbert, L., (2001). Geochemical discrimination of barren and mineralized komatiites associated with magmatic Ni-Cu-(PGE) sulfide deposits: *Canadian Mineralogist*, v. 39, p. 673-696.
- Lightfoot, P.C., Hawkesworth, C.J., Hergt, J., Naldrett, A.J., Gorbachev, N.S., Fedorenko, V.A., Doherty, W. (1993). Remobilisation of the continental lithosphere by a mantle plume: major-, trace-element, and Sr-, Nd-, and Pb-isotopic evidence from picritic and tholeiitic lavas of the Noril'sk District, Siberian Trap, Russia. *Contributions to Mineralogy and Petrology*, 114: 171-188.
- Loewen, M.W., Duncan, R.A., Kent, A.J.R. & Krawl, K. (2013). Prolonged plume volcanism in the Caribbean Large Igneous Province: New insights from Curaçao and Haiti. *Geochemistry Geophysics Geosystems*. 14: 4241–4259. doi:10.1002/ggge.20273.
- Ludden, J., Gelinas, L., Trudel, P. (1982). Archean metavolcanics from the Rouyn-Noranda district, Abitibi greenstone belt, Quebec. 2. Mobility of

- trace elements and petrogenetic constraints. *Canadian Journal of Earth Sciences* 19, 2276–2287.
- Luttinen, A.V., Furnes, H. (2000). Flood basalts of Vestfjella: Jurassic magmatism across an Archaean-Proterozoic lithospheric boundary in Dronning Maud Land, Antarctica. *Journal of Petrology* **41**, 1271-1305.
- Mahoney, J.J.; Coffin, M.R., eds. (1997). Large igneous provinces: continental, oceanic, and planetary flood volcanism. Washington DC, USA, American Geophysical Union, 438pp. (AGU Geophysical Monograph, 100).
- Maier, W., Rasmussen, B., Fletcher, I., Godel, B., Barnes, S.J., Fisher, L., and Yang, S. (2015). Petrogenesis of the ~2.77 Ga Monts de Cristal complex, Gabon: Evidence for direct precipitation of Pt-arsenides from basaltic magma. *Journal of Petrology*, 56, 1285–1308.
- McCuaig, T.C., Kerrich, R., & Xie, Q. (1994). Phosphorus and high field strength element anomalies in Archean high-magnesian magmas as possible indicators of source mineralogy and depth. *Earth and planetary science letters* , 124 (1), 221-239.
- McDonough, W.F. and Sun, S.-S. (1995). Composition of the Earth. *Chemical Geology* 120: 223-253. doi: 10.1016/0009-2541(94)00140-4.
- Molzahn, M., Reisberg, L. & Woerner, G. (1996). Os, Sr, Nd, Pb, O isotope and trace element data from the Ferrar flood basalts, Antarctica: evidence for an enriched subcontinental lithospheric source. *Earth and Planetary Science Letters* 144, 529-545.
- Mungall, J.E. (2016). U-Pb geochronology documents out-of-sequence

emplacement of ultramafic layers in the Bushveld Igneous Complex of South Africa. *Nature Communications*, 7:13385 doi:10.1038/ncomms13385.

Myashiro, A. and Shido, F. (1975). Tholeiitic and calc-alkaline series in relation to the behaviours of titanium, vanadium, chromium and nickel, *American Journal of Science*, 275 (1975) 265-277.

Naldrett, A.J., Lightfoot, P.C., Fedorenko, V.A., Doherty, W. & Gorbachev, N.A. (1992). Geology and geochemistry of intrusions and flood basalts of the Noril'sk region, USSR, with implications for the origin of the Ni–Cu ores. *Economic Geology*, 87: 975–1004.

Osborn, E.F., 1962. Reaction series for sub-alkaline igneous rocks based on different oxygen pressure conditions, *American Mineralogy*, 47 (1962) 211-266.

Pearce, J.A. (2008). Geochemical fingerprinting of oceanic basalts with applications to ophiolite classification and the search for Archean oceanic crust. *Lithos*, 100, pp. 14-48, 10.1016/j.lithos.2007.06.016

Pearce, J.A. and Parkinson, I.J. (1993). Trace element models for mantle melting: application to volcanic arc petrogenesis; in Prichard, H.M., Alabaster, T., Harris, N.B.W., and Neary, C.R., eds., *Magmatic Processes and Plate Tectonics*, Geological Society Special Publications, no. 76, p. 373-403.

Pearce, J.A. (1996). A User's Guide to Basalt Discrimination Diagrams, in Wyman, D.A., ed., *Trace Element Geochemistry of Volcanic Rocks:*

Applications for Massive Sulphide Exploration: *Geological Association of Canada, Short Course Notes*, v. 12, p. 79-113.

Peate, D.W. (1997). The Parana-Etendeka Province. In: Mahoney, J. J. & Coffin, M. F. (eds) *Large Igneous Provinces: Continental, Oceanic, and Planetary Flood Volcanism*. Geophysical Monograph, American Geophysical Union 100, 247-272.

Puffer, J. H. (2001). Contrasting high field strength element contents of continental flood basalts from plume versus reactivated-arc sources. *Geology* 28, 675-678.

Schilling, J.-G. (1973). Iceland mantle plume: geochemical study of Reykjanes Ridge. *Nature*, 242:565-571.

Scoates, J. S., Wall, C. J., Friedman, R. M., VanTongeren, J. A. & Mathez, E. A. (2012). Age of the Bushveld Complex. Abstract, Goldschmidt Conference, Montreal, Canada, session 04G (2012).

Shimizu, K., Nakamura, E., Maruyama, S., Kabayashi, K. (2004). Discovery of Archean continental and mantle fragments inferred from xenocrysts in komatiites, the Belingwe greenstone belts, Zimbabwe. *Geology* 32, 285–288.

Sills, J.D., Savage, D., Watson, J.V., Windley, B.F. (1982). Layered ultramafic gabbro bodies in the Lewisian of northwest Scotland: geochemistry and petrogenesis. *Earth and Planetary Science Letters* 58, 345–360.

Sobolev S. V. Sobolev A. V. Kuzmin D. V. Krivolutskaya N. A. Petrunin A. G. Arndt N. T. Radko V. A. Vasiliev Y. R. (2011). Linking mantle plumes, large

igneous provinces and environmental catastrophes. *Nature* 477 , pp. 312 – 316 .

- Sun, S.S. and McDonough, W.F. (1989). Chemical and isotopic systematics of oceanic basalts; implications for mantle composition and processes. In: *Magmatism in the ocean basins*. Saunders, A.D. and Norry, M.J. (Editors), Geological Society of London, London. 42: 313-345.
- Sun, S.S., Nesbitt, R.W. (1978). Petrogenesis of Archean ultrabasic and basic volcanics – evidence from rare earth elements. *Contributions to Mineralogy and Petrology* 65, 301–325.
- Taylor, S. R. & McLennan, S.M. (1995). The geochemical evolution of the continental crust. *Review of Geophysics* 33, 241-265.
- Tiepolo, M., Bottazzi, P., Foley, S.F., Oberti, R., Vannucci, R., Zanetti, A. (2001). Fractionation of Nb and Ta from Zr and Hf at mantle depths: the role of titanian pargasite and kaersutite. *Journal of Petrology* 42, 221–232.
- Tiepolo, M., Vannucci, R., Oberti, R., Foley, S., Bottazzi, P., Zanetti, A. (2000). Nb and Ta incorporation and fractionation in titanian pargasite and kaersutite: crystal chemical constraints and implications for natural systems. *Earth Planetary Science Letters*, 176, 185–201.
- Thirlwall, M.F., Upton, B.G.J., and Jenkins, C. (1994). Interaction between continental lithosphere and the Iceland plume: Sr-Nd-Pb isotope geochemistry of Tertiary basalts, NE Greenland. *Journal of Petrology*, 35:839-879.

- Wall, C.J., Scoates, J.S., Friedman, R.M., Meurer, W.P., 2012, Age of the Stillwater Complex, Goldschmidt 2012 Conference Abstract, Mineralogical Magazine.
- Wang, K., T. Plank, J. D. Walker, and E. I. Smith (2002). A mantle melting profile across the Basin and Range, SW USA, *J. Geophys. Res.*, 107(B1), <http://dx.doi.org/10.1029/2001JB000209>.
- Watson, E.B. (1980). Apatite and phosphorous in mantle source regions: an experimental study of apatite/melt equilibria at pressures to 25 kbar, *Earth Planetary Science Letters* 51, 322-335.
- Wilson, M. (1989). *Igneous Petrogenesis: A Global Tectonic Approach*. Unwin Hyman, London. <http://dx.doi.org/10.1007/978-1-4020-6788-4>.
- Winchester, J.A., Floyd, P.A. (1976). Geochemical magma type discrimination application to altered and metamorphosed basic igneous rocks. *Earth and Planetary Science Letters* 28, 459–469.
- Xu, Y.-G., He, B., Chung, S.-L., Menzies, M.A., and Frey, F.A. (2004). Geologic, geochemical, and geophysical consequences of plume involvement in the Emeishan flood-basalt province. *Geology*, 32: 917–920.
- Zhai M G, Santosh M. (2011). The early Precambrian odyssey of the North China Craton: A synoptic overview. *Gondwana Research*, 20: 6–25
- Zhao, J.-X., Malcolm, M.T., & Korsch, R.J. (1994). Characterisation of a plume-related c. 800 Ma magmatic event and its implications for basin formation in central-southern Australia. *Earth and Planetary Science Letters*, 121: 349–367.

Figure 7.1: Legend of the eight geochemical groups (A1, A2, B, C1, C2, D, E, & F) recognized within the Frobisher suite sill geochemistry. Major geochemical groups (most commonly occurring) within the sill suite are bolded. Of these, Group D is considered the high-Ti group, and Groups A1 and A2 are considered low-Ti.

Figure 7.2: Assessment of element mobility using the method of [Cann \(1970\)](#), whereby elements are plotted against Zr. **a-c)** Incompatible elements Y, P & Ti form coherent trends with small degrees of scatter, and are thus considered to have been immobile during metamorphism. **d-f)** Large-ion lithophile elements (LILE) K, Ba, & Pb show significant scatter, and are thus considered to have become mobile during metamorphism. All values are in ppm.

Figure 7.3a-c: a) Chondrite-normalized REE diagrams for all geochemical groups. b) Primitive mantle-normalized spider diagrams for all geochemical groups. c) MORB-normalized spider diagrams for all geochemical groups. Chondrite and primitive mantle(PM)-normalization values from [Sun and McDonough \(1989\)](#), and MORB normalization values from [Pearce \(1993\)](#). Abbreviations: M = mafic; UM = ultramafic.

Figure 7.4: Bivariate diagram of $(Gd/Yb)_N$ and $(La/Sm)_N$ illustrating the differences in HREE and LREE slopes between the different geochemical groups in chondrite-normalized REE diagrams. Template from [Naldrett et al. \(1992\)](#).

Figure 7.5: Major oxide variations [SiO₂ (a), CaO (b), Al₂O₃ (c), Cr₂O₃ (d), P₂O₅ (e) & TiO₂ (f)] with Mg# as an index of fractionation.

Figure 7.6: Division between low-Ti and high-Ti groups, which reflects the groups' overall incompatible element or oxide level concentrations, and is primarily discriminated based on differing TiO_2 and P_2O_5 concentrations (Hooper, 1982). Group D is considered high-Ti, while the remainder of the groups are low-Ti. However, given the high abundance of sills containing geochemical signatures of Groups A1, A2 and D, with the rest of the groups as minor components, the term "low-Ti" will be applied strictly to Groups A1 and A2, and high-Ti strictly to Group D.

Figure 7.7: The variation in FeO with respect to decreasing MgO, an oxide that typically decreases as magmas evolve, illustrates different trends between certain geochemical groups.

Figure 7.8: Trace element variations with Mg# as an index of fractionation. Included are highly compatible trace elements (Ni, Co), high-field strength elements (Th, La, Sm, Y, Zr, Sc and Nb), and total REE. Incompatible trace elements which have distinctly higher concentrations in the high-Ti Group D compared to all other groups including Ce, Eu, Ga, Gd, Hf, La, Nb, Nd, Pr, Sm, Ta, Zr, and Hf (some not shown).

Figure 7.9a-b: Ternary diagrams demonstrating the geochemical uniqueness of the various geochemical groups, using the discriminating power of HFSE and REE. **a)** chondrite-normalized REEs (i.e. La, Gd, and Yb) utilized as discriminating variables, which more or less reflect the differences in the LREE (La/Sm), HREE (Gd/Yb) and overall slopes (La/Yb) of the REE. **B)** Nb, Y and La used as discriminating variables. These are the variables chosen for the [Cabanis and Lecolle \(1989\)](#) basalt-discrimination diagram.

Figure 7.10: Bivariate diagram of SiO₂ versus Nb/Y. Low degree partial melts at great depths beneath cold, old, thick lithosphere produce highly alkaline magmas (hence high Nb/Y). The opposite conditions produce tholeiitic magmas. Alkaline/tholeiitic division from [Greenough et al. \(2005\)](#).

Figure 7.11a-b: a) Bivariate diagram of ϵNd versus $^{147}\text{Sm}/^{144}\text{Nd}$. Two trends apparent amongst Groups A1, A2, and D, interpreted to reflect changes in partial melting of the source, and another that may reflect crustal contamination by the upper crustal rocks of the Meta Incognita microcontinent, as suggested by the trend of the samples. b) Primitive mantle-normalized trace element diagram displaying the trace element patterns from sill #27, demonstrating the difference in 15SAB-L057B02's pattern compared to the others. This is interpreted to reflect contamination. Fields for Lake Harbour Group and Ramsay River orthogneiss were derived from isotopic data from Theriault et al. (2001) and [R.J. Theriault \(unpublished data\)](#).

Figure 7.12: Bivariate diagram of ϵ_{Nd} versus Ti/Yb . Indicates that low ϵ_{Nd} values are a result of contamination, as opposed to source enrichment. Adapted from [Leeman and Hawkesworth \(1986\)](#).

Figure 7.13: Geochemical variation through the Killapait sill. Grey fields represent the variation in composition through the sill.

Figure 7.14: Geochemical variation through the Chidliak sill. Grey fields represent the variation in composition through the sill.

Figure 7.15a-c: **a)** Chondrite-normalized REE patterns for the metapyroxenitic and metaperidotitic units of two individual layered ultramafic sills, along with the REE patterns for two individual non-layered ultramafic sills. Two different geochemical signatures are apparent within these layered ultramafic sills and they were likely generated through at least two injections of distinct geochemically different magmas. **b)** Chondrite-normalized REE patterns for the layered mafic-ultramafic Killapait sill are shown, along with their location with respect to stratigraphic height within the sill. No systematic evolution in REE pattern is evident with increasing stratigraphic height in sill. **c)** Metaperidotitic and metagabbroic components of the Chidliak sill exhibit REE patterns that are very similar, with the gabbroic unit containing slightly higher Yb (and overall REE) contents than the metaperidotitic sample, consistent with the gabbro's origin through fractional crystallization.

Figure 7.16: Variation in Eu/Eu* values versus Mg#. Eu/Eu* values >1 indicate plagioclase accumulation, and Eu/Eu* values <1 indicate plagioclase removal from the magma.

Figure 7.17: a) Ti/Yb versus La/Nb diagram. b) ϵ_{Nd} versus La/Nb diagram. La/Nb increases with increasing contamination and Ti/Yb increases with increasing depth of melting. From Pearce (2008) and Lightfoot et al. (1993), respectively.

Figure 7.18: Bivariate diagram of $(Th/Yb)_{PM}$ versus $(Nb/Yb)_{PM}$, distinguishing enrichment due to source enrichment, as opposed to enrichment due to crustal contamination. Adapted from Pearce (2008).

Figure 7.19a-b: a) Bivariate diagram of La/Sm versus Ti-Yb. b) Bivariate diagram of La/Sm versus Nb/La. Distinguishing crustal contamination from lithospheric contamination. Crustal contamination increases La/Sm ratio with increasing Ti and Nb anomalies. Ti/Yb and Nb/La are used as proxies for these anomalies, respectively. From Pearce (2008) and Lightfoot et al. (1993), respectively.

Figure 7.20: Chondrite-normalized REE patterns of Group C2 compared to upper crustal units of the Meta Incognita Microcontinent. Fields for Lake Harbour Group and Ramsay River orthogneiss were derived from isotopic data from [Theriault et al. \(2001\)](#) and [R.J. Theriault \(unpublished data\)](#).

Figure 7.21: Bivariate diagram of $(Gd/Yb)_N$ and $(La/Sm)_N$ illustrating the difference in depth of melting between the various geochemical groups. Primary magmas of samples with $(Gd/Yb)_N > 2$ were generated by melting of garnet-peridotite. Primary magmas of samples with $(Gd/Yb)_N < 2$ were generated by melting of spinel-peridotite. From [Álvaro et al. \(2014\)](#).

Figure 7.22: Bivariate diagram of $(\text{Tb/Yb})_{\text{PM}}$ and $(\text{La/Sm})_{\text{PM}}$ illustrating the difference in depth of melting between the various geochemical groups. Primary magmas of samples with $(\text{Tb/Yb})_{\text{N}} > 1.8$ were generated by melting of garnet-peridotite at great depths. Primary magmas of samples with $(\text{Tb/Yb})_{\text{N}} < 1.8$ were generated by melting of spinel-peridotite at shallower depths. [From Wang \(2002\).](#)

Figure 7.23: Bivariate diagram of La/Sm versus Yb.

Figure 7.24: Mixing of Group D and A1 magmas to produce unique Group A2 samples. Using the mixing feature in Igpet, 70% of Group A1 (low-Ti) sample 14SAB-L003A01 was mixed with 30% of Group D (high-Ti) sample 14SAB-L009A02, which produced a chondrite-normalized REE pattern very similar to Group A2 sample 14SAB-L001A01.

Table 7.1: The occurrences of the various geochemical groups on both Meta Incognita and Hall peninsulas, as well as the type of sills they occur in. Abbreviations: MI = Meta Incognita Peninsula; H = Hall Peninsula; UM = ultramafic sill; M = mafic sill; MUM = mafic-ultramafic sill.

Group	Number of occurrences in different sills	Location	Sill Type
A1	5	4MI - 1H	3UM - 2M/UM
A2	11	5MI - 6H	7M - 1UM - 3MUM
B	3	1MI - 2H	1M - 2MUM
C1	4	3MI - 1H	1M - 1UM - 2MUM
C2	3	2MI - 1H	1M - 2MUM
D	8	8MI	1M - 4UM - 3MUM
E	2	2H	2M
F	2	2H	2M

Table 7.2: The occurrences of certain geochemical groups with other geochemical groups, as well as the type of sills they occur in. Boxes highlighted in grey show the occurrences of sills with only the one geochemical group.

Group	A1	A2	B	C1	C2	D	E	F
A1	1UM	-	-	-	-	-	-	-
A2	2MUM	1MUM - 1UM - 6M	-	-	-	-	-	-
B	0	0	1M	-	-	-	-	-
C1	1MUM	1MUM	2MUM	1UM - 1M	-	-	-	-
C2	1MUM	1MUM - 1M	1MUM	1MUM	0	-	-	-
D	2UM	1M	0	0	0	3MUM - 1M - 2UM	-	-
E	0	1MUM	0	0	0	0	2M	-
F	0	0	0	0	0	0	0	1MUM - 1UM

Chapter 8 – Discussion & Conclusion

8.1 Petrogenesis

Based on the number of occurrences of each geochemical group within sills of the Frobisher suite ([Table 7.1](#)), the groups may be divided into major and minor geochemical groups. The major components include low- and high-Ti Groups A1 and A2, and D, respectively. The minor component includes contaminated Groups B, C1 and C2, and samples from Groups E and F.

8.1.1 Petrogenesis of major geochemical groups

The major geochemical groups of the suite, the low-Ti Groups A1 & A2 and high-Ti Group D, appear to have experienced multiple, distinct differentiation processes during their petrogenesis, as highlighted by the rather diverse major and trace element compositions described in Chapter 7. These processes include, fractional crystallization, and crustal or lithospheric contamination.

8.1.1.1 High-Ti magma type (Group D)

The high-Ti magma type, Group D, occurs exclusively on Meta Incognita Peninsula, both along the coast in the west, and adjacent to the eastern boundary of the Narsajuaq terrane in the east ([Liikane et al., 2015](#)). With respect to emplacement within a reference (pre-collision and deformation) Meta Incognita microcontinent, the high-Ti magmas were emplaced in the more proximal facies Lake Harbour Group (a.k.a. inland-ward in the microcontinent). This signature most often occurs in monogenetic sills (i.e. only one geochemical signature within a single sill). However, in two layered ultramafic sills the high-Ti magma

type occurs with low-Ti Group A1, indicating that some layered sills were derived from multiple pulses of magma. High-Ti group samples vary from mafic to ultramafic in composition.

Geochemical evaluation of the high-Ti group indicates that its primary magma was derived from low degrees of partial melting within the garnet stability field (i.e. at depths >90 km), such that garnet remained in the mantle source (e.g. [Arndt et al., 1993](#); [Ewart et al., 1998](#); [2004a](#)). This produced a mildly alkalic tholeiitic parental magma and OIB-like chondrite-normalized REE patterns ([Puffer, 2001](#)). Positive Nb-Ta anomalies characteristic of this group indicate an enriched component within the source, which may have been eclogite, which is typically the first rock type to melt in ascending asthenospheric mantle, or perhaps eclogite (e.g. [Sobolev et al., 2011](#)). The presence of a thick lithospheric cap would limit melting to a short pressure interval at greater depths within the garnet stability field, and a smaller melt volume would be produced (typical of alkali basalt fields; e.g. [Hole and Saunders, 1996](#); [Niu et al., 2011](#)).

The parental magmas to the high-Ti magmas may have been vertically fed from the asthenosphere under a thicker lithospheric area, given the apparent depth of the melting and its location of emplacement relative to the reference Meta Incognita microcontinent. En route to the level of emplacement, the high-Ti magmas arrived more directly (i.e. less crustal residence time) compared to those of the low-Ti groups, as indicated by their high Mg# (generally > 56) and the lack of contamination. Dykes were likely established at the base of the lithosphere transferring magma to upper crustal levels, with no residence time in

crustal magma chambers, as is typically the case with alkaline magmas extruded from monogenetic volcanoes (Muraoka, 1997). The generally low silica, high MgO nature of the more alkaline magmas results in less dense magma that is less likely to get trapped in crustal magma chambers.

In plume-derived provinces, the more alkalic magmas (e.g. Figure 7.10), such as that of the high-Ti Group, are thought to either represent melting of the eclogitic or basaltic component of a plume head. In passive magma generation, parental magmas to high-Ti groups are interpreted to simply reflect low-degree melting of the mantle material. If the Frobisher suite was plume-derived, the later process of high-Ti melt generation is supported by the fact the epsilon-Nd values of the high- and low-Ti groups span approximately the same range, as expected for two melts of differing melting degrees from the same source. However this can also apply to differences in the partial melting of a plume head. If there were major eclogitic or basaltic contributions to the high-Ti group, this should be reflected in the Sm-Nd isotope geochemistry, with distinctly different epsilon Nd values for the high-Ti group compared to the low-Ti group. However, generation of the high-Ti melt from eclogitic streaks within the plume head could also explain the positive Nb anomaly, and perhaps, if the Sm-Nd isotopic characteristics of the eclogitic or basaltic components were similar to that of the rest of the plume head, overlapping ϵ Nd values within both high- and low-Ti groups would still be observed. The presence of eclogite increases the fertility of the source (Sobolev et al., 2011; Saunders, 2005), and the Nb-anomaly is interpreted to represent residual eclogite in the mantle source. If eclogitic streaks within the plume head

were a major contributor to the high-Ti magmas, this component must have had a similar isotopic composition as bulk of plume head material.

8.1.1.2 Low-Ti magma types (Groups A1 & A2)

The low-Ti magma types, Groups A1 and A2, occur on both Meta Incognita Peninsula and northern Hall Peninsula. With respect to emplacement within a reference Meta Incognita microcontinent, the low-Ti magmas were mostly emplaced in the distal facies Lake Harbour Group (aka the deeper water, off-shelf strata). Group A2 occurs with geochemical Groups A1, C1 and C2, while Group A1 only occurs with Groups A2 and D ([Table 7.2](#)). On northern Hall Peninsula, Group A2 occurs only in monogenetic sills. Group A2 samples are almost exclusively mafic, while Group A1 samples are exclusively ultramafic in composition.

The low-Ti Group was derived from high degrees of partial melting within the spinel stability field (i.e. at depths <90 km), resulting in a tholeiitic magma (e.g. [Arndt et al., 1993](#); [Xu et al., 2004](#); [Herzberg & Gazel, 2009](#)). However, it could have been derived from melting in the garnet stability field if garnet was fully exhausted from the source. Variable negative Nb-Ta anomalies indicate that the parental magmas of this geochemical group were variably contaminated. In route to the level of emplacement, substantial fractionation from primary melts occurred in Group A2, likely in staging chambers within the crust (e.g. [Cox, 1980, 1993](#)), as evidenced by the more evolved compositions (low MgO contents, generally between 7-9wt.%). Assimilation and resulting contamination, which is

often concurrent with fractional crystallization (i.e. assimilation fractional crystallization (AFC) processes) is apparent in some samples (Figures 7.11, 7.18 & 7.19). In contrast, Group A1 appears to have either traveled directly from the mantle or represents olivine cumulates, given its high Mg# (~70). However, all samples from this group appear to have suffered contamination, unlike Group A2 samples, of which only a select few are contaminated.

In comparison, Groups A1 and A2 display a very similar, overlapping range of Gd/Yb ratios, indicating they were derived from the same or a similar depth and degree of melting. In sills where both groups are present, Group A1 always comprises the ultramafic component, while Group A2 is the mafic component. It appears that Groups A1 and A2 were derived from the same parental magma, and Group A2 simply represents a more evolved (<Mg#) equivalent of Group A1. However, the gap in the data between the bulk of Group A2 samples and the Group A1 samples makes it difficult to establish the relationship through crystal fractionation, so geochemical modelling is required to conclusively establish this relationship.

The occurrence of consistently contaminated, less evolved Group A1 samples with largely uncontaminated, more highly evolved Group A2 samples suggests that perhaps fractionation of a parental magma occurred in a magma chamber where the host rocks were unable to generate a contamination signature within the fractionating magma during AFC processes (e.g. base of lithosphere).

Ito & Mahoney (2005) suggest OIB and MORB sources are identical and that these two end-members merely reflect different degrees of melting within a heterogeneous source of 90% depleted peridotite and variable proportions of enriched peridotite and pyroxenite. In a similar manner, I propose that the low- and high-Ti magmas are interpreted to represent the same (or similar) heterogeneous mantle source and that differences in trace element patterns represent different degrees of partial melting at different depths with contributions from eclogite in the latter. This is supported by their similar ϵNd values and similar patterns that appear to differ by only slight LREE-enrichment (La/Sm) and depth of melting (i.e. Gd/Yb ratio). If low-Ti magmas resulted from mixing of OIB source with depleted MORB source, a significantly higher ϵNd would be expected in Groups A1 and A2 compared to Group D (OIB-like source).

8.1.2 Petrogenesis of minor geochemical groups

8.1.2.1 Groups B, C1, and C2 magma types

Groups B, C1, and C2 occur on both Meta Incognita and northern Hall peninsulas. With respect to emplacement within reference Meta Incognita microcontinent, the Group B and C2 magmas were emplaced in both proximal and distal Lake Harbour Group (LHG) sequences (similar to Group A1 and A2's geographic occurrence), whereas Group C1 was emplaced exclusively in proximal facies Lake Harbour Group (similar to Group D's geographic occurrence). These groups generally occur together within the same intrusions and all show the characteristic signs of contamination (Table 7.2). Samples from

these groups are mostly mafic in composition, however ultramafic samples are also present. All three groups contain at least one sample containing both plagioclase and olivine (implications discussed in 8.1.4).

The large majority of samples from Groups B, C1 and C2 have experienced significant plagioclase accumulation, as indicated by the presence of prominent positive Eu anomalies and high plagioclase abundances observed in thin section and hand sample. On geochemical diagrams, samples show significant scatter and exhibit clear signs of contamination or derivation from metasomatized SCLM.

It is proposed that Groups B and C1 represent contaminated variants of Groups A2 and D, respectively, based on observations from trace element patterns and relationships between these two pairs of groups within 2D-space on bivariate plots. In [Figure 7.6a-b](#), a similarity is observed in both chondrite-normalized REE and primitive mantle-normalized trace element patterns, specifically in the HREE, which are the least affected by crustal contamination. The relationship between their slopes on the chondrite-normalized REE diagrams are further illustrated and compared on [Figure 7.4](#). A side-by-side comparison is illustrated in [Figure 8.1](#) for more clarity. Both Groups B and C1 show lower concentrations of REE compared to their suggested uncontaminated equivalents. Lower concentrations may be achieved by mixing with the depleted and metasomatically enriched Group C2. This notion is supported by the observation that Groups A2 (or A1) and B do not occur within the same sills, and Groups D and C1 do not occur together in sills.

Furthermore, Group C2 is proposed to have been derived primarily from metasomatized subcontinental lithospheric mantle, with a source previously depleted by a prior melting event (a restite) and later enriched in LREE and other more incompatible trace elements via a subduction-related metasomatic event, all occurring prior to partial melting to produce the Frobisher Bay event.

8.1.2.1.1 Group C2 - an in-depth look

Group C2 is particularly odd with regard to its geochemistry. Its pattern appears to represent extensive crustal contamination, which is supported by the significant negative Nb-Ta anomalies and high La/Sm ratios. However, Group C2 has similar or lower SiO₂ content compared to the rest of the samples, so simple crustal assimilation cannot explain its chemistry. The group is also quite depleted in HREE compared to upper crustal Meta Incognita microcontinent basement and supracrustal units, and no in-situ assimilation of this material was observed.

It is proposed that Group C2 parental magmas may have been generated from melting of metasomatized lithospheric mantle. Melts from descending slabs in subduction environments are initially SiO₂-rich, but become progressively less SiO₂-rich as melting progresses (i.e. SiO₂ is low in high-degree of melts of subducting slabs). The melts of the eclogitic slab (i.e. garnet+amphibole) are characteristically Sr-enriched (no plagioclase remaining in restite), Y-HREE-depleted (abundant garnet in restite), and high LREE/HREE ratios ([Drummond & Defant, 1990](#)). The MgO-rich character of the magma is attained via contamination with peridotite from the overlying mantle wedge. The low-SiO₂ and

high mafic elements can be explained by re-melting of peridotite previously modified by slab-melt (Shirley and Hanson, 1984; Martin et al., 2005). These processes mostly explain what is observed in Group C2.

Furthermore, significant HREE depletion indicates that Group C2 was generated from partial melting of a source where garnet or amphibole remained in the restite, retaining HREE (HREE compatible in amphibole and garnet). Support for residual garnet in the source is the high Gd/Yb ratio. Trace element patterns in Group C2 differ from most arc-rocks, but they do contain depleted HFSE (Nb-Ta, Ti) like arc rocks. This suggests that Nb and Ti are retained in amphibole and accessory rutile and titanite phases. The LREE-enrichment in Group C2 is easily explained by prior metasomatism of the source (i.e. lithospheric mantle).

The very low TiO_2 and P_2O_5 values in Group C2 (oxides that acts incompatibly during melting of peridotite) may have resulted from a previous melting event, removing incompatible elements from residual mantle source.

Alternatively, high-T, high-MgO primary picritic or komatiitic magmas may have interacted with the lithosphere during their ascent and became heavily contaminated. This is plausible because fusible lithosphere is generally that which has been previously re-fertilized by a metasomatic event, which allows it to melt more easily. The question is whether Group C2 represents direct melting of metasomatized SCLM, or whether it represents an asthenospheric source that interacted with and assimilated great amounts of metasomatized SCLM or Archean tonalitic basement (J.Mungall, pers.comm. 2017; e.g. Mungall, 2007;

Mungall, 2010; Mungall et al. 2016). The common thread in both scenarios is the dominating SCLM or crustal component in the melt. Regardless of origin, the volume of Group C2 magma emplaced during this event is very minor. However, if Groups B and C1 do contain a significant metasomatized-SCLM or crustal component, its overall presence within the Frobisher suite is actually quite significant.

8.1.2.2. Group E magma type

Group E was documented in one monogenetic sill and one polygenetic sill, both located in northern Hall Peninsula, and includes samples that are mafic in composition. The low Gd/Yb ratio indicates that it was derived from melting within the spinel stability field, or it may have been derived from the garnet stability field with garnet being completely exhausted from the source. The distinctly higher Yb contents compared to other groups at similar Mg# and prominent negative Ta, P, Ti, and Zr-Hf anomalies indicates that it was differentiated to a great extent following melt-generation. Based on geochemical similarities and its geographic proximity with Group F, it is proposed that Group E may potentially represent a more highly evolved magma of Group F.

8.1.2.3 Group F magma type

Group F was documented in only two monogenetic sills, located on northern Hall Peninsula, and samples range from mafic to ultramafic in composition. With respect to emplacement within a reference Meta Incognita

microcontinent, Group F magmas were mostly emplaced in the distal facies Lake Harbour Group sequence (aka deeper water, off-shelf strata). The parental magma to Group F was likely derived from a high degree of partial melting at shallow depths within the spinel stability field, consistent with its komatiite-like chondrite-normalized REE pattern, in which it is depleted in the most incompatible elements with relative HREE enrichment, its high-MgO content, and its location with respect to the reference microcontinent. Other sources attribute this pattern to melting of depleted upper mantle (e.g. [Workman and Hart, 2005](#)), or this mantle source is representing a significant component of the magma. All samples exhibit some degree of minor contamination, which is expected in high-T, MgO-rich high-degree partial melts, given their tendency to flow turbulently ([Huppert and Sparks, 1985](#)).

Sills of the Group F magma type lack a spatial association with the other geochemical groups, and their geochemical fingerprint is distinctly different from the other geochemical groups, including in terms of normalized trace element patterns, trace element ratios, and ϵNd values. The ϵNd values of Group F are significantly higher (+4.58 to +6.13) than those in the rest of the Frobisher suite (<3.69). Uncontaminated equivalents of Group F would have epsilon Nd values even higher than these. This observation supports an increased depleted upper mantle component in the magma, as opposed to this group simply reflecting a very high degree of partial melting of the same or similar source to the high-Ti and low-Ti magmas.

8.1.3 Potential parental magma compositions

Samples from Group A1 appear to represent the best candidates for parental magmas to the low-Ti groups, with Mg#~66 (up to 78), and Ni and Cr up to ~3000 ppm and ~2000 ppm, respectively. This is consistent with experimental results, which have shown that melting of lherzolite mantle will produce melts with Mg#66-75, Cr >1000 ppm, and Ni >400-500 ppm ([Winter, 2001](#)). However, this group may also represent olivine or pyroxene cumulates associated with a magma parental to both low-Ti groups. The best candidates for parental magma compositions for the high-Ti group are the large number of Group D samples with Mg#s ~64, and Ni and Cr greater than > 900 ppm and >1600 ppm.

8.1.4 Petrogenesis of the Killapait sill

The Killapait sill is a layered mafic-ultramafic layered intrusion located on the south coast of central Meta Incognita Peninsula. Its true thickness is at least ~470 m, but its upper contact was not observed due to erosion. The sill is preserved in a F₃ synformal basin, or syncline, if the mafic layers were situated on top of the ultramafic component at the time of emplacement.

A combination of bulk-rock, mineral and isotope geochemistry, petrography, and field observations has led to the conclusion that the Killapait sill reflects a complex history of formation, involving multiple injections of compositionally distinct magma batches into a single growing intrusion. The first indication of its complexity was the non-systematic variation in trace-element and isotope geochemistry from the basal contact to the uppermost portion of the

intrusion, which was further supported by hand sample and petrographic observations.

In a sill derived from in-situ magmatic differentiation, systematic variation in trace-element and isotope geochemistry would be expected (e.g. [Egorova & Latypov, 2013](#)) from the basal contact to the uppermost portion of the intrusion, mostly reflecting the depletion of MgO in the magma and increase in incompatible trace element concentrations. However, this is not observed in the Killapait sill. In fact, within a small stratigraphic interval, major changes in chemistry are observed that cannot be explained simply by crystal fractionation. For example, through the stratigraphy of the sill, there were four geochemical groups identified (A1, A2, C1, C2), with Group C2 appearing to represent the most voluminous component. Adding to the complexity, instead of the geochemical signatures occurring as four separate packages within the intrusion, the same group occurs at multiple levels within the magmatic stratigraphy, and in one instance, samples from three different geochemical groups (A2, C1, C2) were collected from a single station (14SAB-L012). Differences in Sm-Nd isotope geochemistry obtained from the various groups within the sill supports the recognition of separate magma batches. Assuming sills were continually injected into the centre of the partially crystallized intrusion, it could be proposed that at least 5 injections of four geochemically distinct magmas were injected into the sill. Each individual sill was likely emplaced after significant crystallization had occurred (e.g. 50%, as per general estimates of cooling rates by [Cawthorn, 2012](#)), such that repeated injections didn't mix and the sills retained their physical

identity and geochemical character. Truncation of layering within a sample collected within the gabbroic portion of the sill is shown in [Figure 8.2a](#), which appears to represent a gabbro with finer-grained gabbroic xenoliths. However, given that this sample is of the Group C1 geochemical group and the geochemistry represents the sample as a whole, it is possible that the sample represents mixing of Group D and Group C2 magmas, with the fine grained gabbro being one magma type and the coarse-grained gabbro being the other. Chilling was not observed in the field, which may be due to the fact that chilling may or may not have occurred depending on the temperature contrast between sills (i.e. 100 °C difference will result in chilled contact formation) or perhaps it was due to lack of exposure or coarsening through high-grade metamorphism. Another possibility is that after the initial sill injection, subsequent injections underplated the sills which preceded them (i.e. vertically stacked succession of sills). This could produce the non-systematic variation in geochemistry that is observed within the Killapait sill.

In the upper 200 m of the sill, which is comprised of Group C2 magma type, bulk-rock geochemical compositions suggest that magmas experienced FeO-saturation and began following a calc-alkaline trend of magmatic evolution (concurrent depletion in FeO and MgO; e.g. $Mg\# = 50$ and $MgO \sim 1$ wt.%), which coincides with the crystallization of Fe-, Ti- oxides. In this upper stratigraphic level within the sill, oxides appear to be absent, with the exception of dismembered pods of oxides in the unit below it ([Figure 8.3](#)). This may be explained by crystallization and removal of oxides from the magma prior to the

emplacement of this respective magma batch, or the oxides were separated via gravitational settling in-situ. A high-oxygen-fugacity environment may have been present, as low oxygen fugacities typically suppress ilmenite and magnetite fractionation (e.g. [Snyder et al., 1993](#))

Two Group C2 samples from the ultramafic component of the Killapait sill (14SAB-L013A01 and 14SAB-L013A02) have significantly higher positive Eu/Eu* than the rest of the samples in the sill, and contain large plagioclase megacrysts ([Figure 8.2b](#)) in coexistence with significant amounts of olivine. The occurrence of these two minerals together requires low pressures and very high temperatures (>800°C), so they are typically found in crustal mafic intrusions (i.e. <600 MPa) within units olivine gabbro or troctolite in composition ([Bucher and Grapes, 2011](#)). There has been no documented occurrence of prograde plagioclase formation in olivine-rich rocks, to date, but they could theoretically form in low-P granulite facies terrane ([Bucher and Grapes, 2011](#)). In light of this, the plagioclase within the ultramafic portion of the Killapait sill is interpreted to have grown prior to the emplacement of magma (cognate xenocryst) from which the sill crystallized, and is considered to be a primary magmatic relict; of course An-content has changed through progressive prograde metamorphism. A common and plausible site of crystallization of this plagioclase is within a shallow crustal magma chamber where it crystallized from its slowly cooling parental magma. Given the abundance of plagioclase grains within the basal area of the sill, the parental magma to the sill may have interacted with a crystal mush of plagioclase grains (e.g. [Sheth, 2016](#)) near the top of a magma chamber. The

lower pressures in the upper crust may have allowed plagioclase to float and accumulate toward the surface of the magma chamber (e.g. [Higgins and Chandrasekharam, 2007](#)).

In spite of having established that the Killapait sill does not represent one simple sill (i.e. one injection of magma that differentiated in-situ into mafic and ultramafic components), its precise history of formation remains elusive. I propose two models for the formation of the Killapait sill: (1) composite sill formation, through multiple injections of magma into a single intrusion; and (2) vertical stacking of discrete sills through underplating (overplating is far less likely, as per [Menand, \(2008\)](#); see also chapter 2). These models are based on the available data and assume continuity between the occurrences of specific geochemical groups. Note that in reality, this is not likely the case, based on the complexity observed in the limited samples collected. In both models, sills are injected as discrete bodies that do not mix with previous injections.

8.1.4.1 Models of Killapait sill formation

In the composite sill model ([Figure 8.4](#)), the sequence of events may have been as follows:

- (1) A batch of low-Ti group magma (Group A1) is first injected into the Lake Harbour Group, where it begins to differentiate into Group A2. This is supported by the presence of a fine-grained unit of Group A1 character at the basal intrusive contact, and the recognition of a Group A2 sample as the next

geochemical group higher in the stratigraphy. Alternatively, Group A2 may have been injected as a separate magma batch.

- (2) A minor injection of Group C1 magma was emplaced in the sill.
- (3) A significant volume of plagioclase-charged Group C2 mafic and ultramafic magma were emplaced, significantly increasing the size and thickness of the sill. Prior to this emplacement, a crystal mush of plagioclase was entrained by the magma, likely from the top of a magma chamber. Due to the viscosity of the ultramafic magma and density of the overlying magma, the plagioclase cannot escape through floatation to higher levels within the sill. Or perhaps some of it was able to escape, and its presence is masked by a combination of naturally plagioclase-abundant mafic rocks and high-grade metamorphism. Alternatively, if other magma batches were already mostly crystallized, this would also act as a barrier to plagioclase floatation.
- (4) Final injections into the sills comprised Group C1 magmas, which were able to retain their geochemical character and perhaps physical identity. Alternatively, if Group C1 represents contaminated Group D magma, then Group D magma may have been injected into the sills, mixing with Group C2 magmas, forming the Group C1 magma type. However, given the consistency and distinguishability of chondrite-normalized REE patterns, mixing would have likely occurred at depth in a separate magma chamber, prior to emplacement in the sill.

If the Killapait sill represents the emplacement of multiple sills comprised of geochemically distinct batches of magma, each underplating the previous sill (Figure 8.5), then the order of injection is as follows:

- (1) First injection of magma would have been a voluminous batch (or batches) of Group C2 mafic magma, assuming Group C2 continues to the upper contact.
- (2) Injections of mafic magma alternating between Group C1 and C2 types underplate the first sill as discrete bodies themselves.
- (3) Injections of ultramafic magma would have come last, underplating the mafic sills as discrete sills of differing geochemical character. At this stage of stacked-sill complex formation, the Group C2 magmas entrained a crystal mush of plagioclase prior to emplacement, likely from the top of a magma chamber.

There also exists the possibility the actual formation of the Killapait sill involved a combination of these models. For example, Group C1 and C2 magmas may have formed a composite sill that was later underplated by the Group A1 and A2 magmas (or magma if the latter two share a parental magma). The exact sequence of events to led to this sill's formation is not clear, and there are surely more variants of the models stated above. Additional field work, more detailed sampling (as much as allowed by the current exposure) and geochemical modelling is required to further test these models. Furthermore, this detailed study on the Killapait sill sheds light on the complex process that occurred during the formation of the Frobisher suite. There is a good possibility that a number of other sills may have been derived through similar process, to

greater or lesser extents. It should be also noted that these models assume that there is a basal feeder (dyke or sill) to the Killapait sill, but there is a possibility that this sill was fed laterally by dykes. In the latter case, other models of emplacement may be proposed.

8.1.5 Petrogenesis of the Chidliak sill

The Chidliak sill is a layered mafic-ultramafic layered intrusion located on northern Hall Peninsula, south of Chidliak Bay (and west of Ptarmigan Fjord). It is ~230 m in thickness (true thickness) and is preserved as an overturned string of boudins. It has three main components, including a basal metapyroxenite, volumetrically dominant metaperidotite, and an upper metagabbroic cap.

A combination of bulk-rock, mineral and isotope geochemistry, petrography, and field observations was used to understand the petrogenesis of the Chidliak sill, revealing a much more simpler history of formation, compared to the Killapait sill. One key difference is that the formation of the Chidliak sill involved multiple injections of a single magma type (Group F). Key evidence for multiple injections includes clasts of basal pyroxenite suspended within the peridotite unit (Figure 8.6), immediately above the pyroxenite-peridotite contact, indicating that the pyroxenite had crystallized prior to emplacement of the parental magma to the peridotite and had been entrained through erosion of the sill floor. This notion is further supported by variations in trace element geochemistry through the sill, specifically, the more primitive character of the core of the peridotite unit, and intermediate composition of the basal pyroxenite.

Further investigation reveals that the pyroxenite contains an intermediate composition (e.g. in Mg#, Cr, Ni) between the peridotite unit and gabbroic units, as would be expected for a parental magma composition to the peridotitic and gabbroic units. The REE concentrations of the pyroxenite are equal or lower than the other two lithologies. This, along with the other evidence, seems to suggest that the basal pyroxenite may represent “frozen” magma at the base of the sill, due to conductive cooling, prior to the onset of crystal fractionation and gravitational setting. Additionally, if the basal pyroxenite is of cumulate origin, Cr contents would be significantly higher in this unit compared to the peridotite or gabbro, which it is not. Cr contents in the pyroxenite are equal to the gabbroic unit, and significantly lower than the peridotite unit. Basal pyroxenites have been observed elsewhere in mafic-ultramafic intrusions, for example in the Chukotat Group of the Circum-Superior LIP, where they are interpreted to be a “chilled border phase” (e.g., [Bédard et al., 1984](#); [St-Onge et al., 1987](#); [Thibert et al., 1989](#)).

Samples from the centre of the peridotite unit (15SAB-L078A01 and 15SAB-L078A02) are distinctly more primitive (>Mg#, Cr, Ni) than peridotite located closer to the contacts with the gabbroic and pyroxenitic units. However, given the very similar Yb contents, these samples may just reflect a difference in the mineralogical nature of that horizon in the sill (i.e. greater abundance of olivine, or olivine cumulates). Furthermore, these remain geochemically distinct from the rest of the sill.

Two possible origins for the gabbroic cap are: formation through in-situ differentiation of the parental magma to the peridotite, or it may have differentiated from the magma which produced basal pyroxenite, which was later intruded by additional magma which formed the peridotite. The gabbro exhibits a more elevated chondrite-normalized REE patterns than both pyroxenite and peridotite, making it feasible that it could have formed from either unit. The concordant contact between peridotite and gabbro supports differentiation of the gabbro from the peridotite unit, as well as lack of gabbroic clasts suspended in the peridotite, as is the case with the pyroxenite. Sm-Nd isotopic data indicate that the gabbroic unit is derived from a more depleted mantle source ($\epsilon\text{Nd} = 6.13$) compared to the peridotitic unit ($\epsilon\text{Nd} = 4.58$). However, this difference may be the result of a combination of increased instrumental error due to such low concentrations of Nd within the Group F samples (range of 0.9-2.4 ppm Nd), natural variations within the magma reservoirs, and higher degree of crustal contamination in that sample (observed in primitive-mantle normalized diagram as Nb-anomalies).

The aforementioned evidence indicates that the Chidliak sill does not represent one simple sill (i.e. one injection of magma and differentiated in-situ into mafic and ultramafic components), but is best explained through the composite sill model of formation (proposed below).

8.1.5.1 Models of Chidliak sill formation

In this first model (Figure 8.7), the Chidliak sill formed through open-system in-situ differentiation, with continuous pulses of magma. The basal pyroxenite represents magma that was “frozen” at the base prior to initiation of olivine crystal fractionation and gravity-induced settling. The thin gabbroic unit that caps the rest of the sequence may be interpreted to represent a late differentiate from the parental magma to the sill. Sill formation was initiated by a first pulse of magma, from which the pyroxenite began to crystallize, due to loss of heat, through conduction, to the relatively cool wall rocks. Within komatiitic basalts, pyroxene is known to crystallize before olivine, as the dominant phenocrysts, which may accumulate at the base of the flows (e.g. Arndt et al. 1997). The timing of crystal settling may have been offset because of continual injections of magma into the sill, or perhaps the sill served as a conduit for magma through-flow. Turbulent flow resulting from the low viscosity, ultramafic nature of the magma and additional influx of magma partially eroded the continually crystallizing basal pyroxenite, as evidenced by pyroxenitic clasts suspended in the peridotite unit (Figure 8.6). These clasts serve as the only constraint on relative ages between units within the Chidliak sill. There are no clasts of gabbro within the peridotite, however, but this does not mean that the parental magma to the peridotite cannot have intruded between the partially crystallized gabbroic and pyroxenitic units of the sill (e.g. the second model). Crystal fractionation and gravity-induced settling (laying the peridotite unit) were only able to commence once turbidity in the system had ended. Through this differentiation process, the gabbroic cap of the sill was formed.

The second possible model (Figure 8.8) involves an initial injection of magma, which begins the in-situ differentiation into the pyroxenitic basal unit and the upper gabbroic unit. Before the sill was completely crystallized, vast amounts of primitive magma were injected into its core, forming the peridotite unit. The presence of clasts of pyroxenite “floating” with peridotite units supports this model, however, the absence of gabbroic clasts in the peridotite does not. In addition, the pyroxenitic sample right at the pyroxenite-peridotite contact has a similar but more evolved chondrite-normalized REE pattern compared to the gabbroic samples. This sample also contains thin section-scale discontinuous anorthositic layers, indicating this portion of the sill was quite evolved, approaching a mafic (gabbroic) composition.

This second model is also supported by the very different, but still depleted, Nd isotope signatures, the observation of rip-up clasts of pyroxenite just above the pyroxenite/peridotite contact, the smooth decrease in MgO with decreasing FeO from base of pyroxenite to top of pyroxenite and then into gabbro (essentially slicing out peridotite, which appears to be out of place, with higher MgO and FeO contents). Contents of Cr are similar between the basal pyroxenite and gabbroic unit, but Ni is significantly higher, which may suggest that as the pyroxenite was “freezing,” some fractionated olivine was allowed to settle.

An example of an intrusion, that has a similar stratigraphic lithological profile to the Chidliak sill, is the ca. 1075 Ma Wingellina Hills intrusion of the ca. 1090-1040 Giles event in the Musgrave Province of central Australia. This

intrusion contains a basal orthopyroxenite that is overlain by a clinopyroxenite, then peridotite, wehrlite and finally gabbro-norite (Ballhaus and Glikson, 1989). This intrusion and others like it in the Giles Complex, are interpreted to represent intrusions that have experienced magma replenishment, resulting in changes in the geochemistry and fractionating mineral phase(s) (Ballhaus and Glikson, 1989). This is consistent with both models proposed here.

However, the model that appears to be most appropriate is the first model, where the basal pyroxenite formed as a “freeze” layer and the gabbroic unit formed as a product of differentiation of the peridotite. But it should be noted that these models assume that there is a basal feeder (dyke or sill) to the Chidliak sill, but there is a possibility that this sill was fed laterally by dykes. In the latter case, other models of emplacement may be proposed.

8.1.6 Candidacy for a Large Igneous Province (LIP)

8.1.6.1 Areal Extent

One criteria that the Frobisher suite must meet in order to be considered the plumbing system to a LIP, is a minimum areal extent of c. 100,000 km², as per the definition of a LIP (Ernst, 2014) and other studies on intrusive provinces and plumbing systems of LIPs (e.g., Yale and Carpenter, 1998; Marzoli et al., 1999; Ernst et al., 2005). Based on a rough, very conservative estimate, the suite’s areal extent has been determined to be approximately 130,000 km². This areal extent is based on the extent of the Lake Harbour Group, which is pervasively intruded by Frobisher suite sills of varying thicknesses.

8.1.6.2 Volume of magma

The high volume of magmatism within LIPs is another critical attribute. While most sills are not very thick (averaging ~50 m thickness), there is a large areal extent of sill emplacement, and enough sill material to consider the suite to have resulted from an anomalously high-degree of mantle melt production, typical of the generation of mafic-ultramafic components of LIPs.

8.1.6.3 Duration and pulsed nature of magmatism

Criteria regarding the duration of magmatism cannot yet be considered due to the lack of geochronological data from this suite of sills. Once systematic dating is conducted on the suite of sills, it could be determined whether the magmatism was emplaced within a short duration pulse (or multiple pulses spanning several 10s of myr; [Ernst, 2014](#); [Ernst and Youbi, 2017](#)).

8.1.6.4 Intraplate character

The intraplate character of the Frobisher suite is indicated by the Group D magma type, which is mildly alkaline and exhibits OIB-source-like characteristics (e.g., positive Nb-Ta anomalies, LREE-enrichment with fractionated HREE indicating low-degree partial melting of peridotite within the garnet stability field, high-Ti character). These characteristics are suggestive of intraplate magmatism (i.e. mantle plume-derived magmatism).

High Ti/Y ratios are characteristic of intraplate setting, as Y is preferentially retained in garnet in restite. This doesn't occur at plate margin setting where melting is shallower (hence no garnet present), and also the addition of H₂O allows for higher degrees of partial melting that can consume garnet even if deep melting occurs ([Pearce, 1996](#)).

8.1.6.5 Provinciality within Frobisher suite

While not a characteristic feature of LIPs, the marked provinciality in the distribution of high- and low-Ti geochemical groups is common amongst LIPs, such as in the Paraná-Etendeka (e.g. [Peate, 1997](#)), Karoo (e.g. [Jourdan et al., 2007a](#)), and Emeishan (e.g. [Xu et al, 2001](#)) LIPs. As a specific example, within the Golden Valley Sill Complex of the Karoo LIP ([Neumann, 2011](#)), various sills within a confined area appear to be derived from different magma batches, or a combination of them. This same observation is apparent in the Frobisher suite. In many cases, individual intrusions appear to be derived from separate, distinct magma batches, and sills in relatively close proximity (i.e. kilometers apart) have different geochemical signatures. Despite having been obscured by several deformation events, a spatial relationship still appears to exist. Furthermore, the emplacement of Group D into proximal facies Lake Harbour Group strata suggests a “central” location with respect to a reference Meta Incognita microcontinent (as described above). In contrast, the low-Ti magmas, interpreted to reflect high-degree partial melting at shallower depths, primarily occur in distal

facies Lake Harbour Group strata and therefore near the margins of the reference microcontinent.

In support of the implications of this observation, the mildly alkalic tholeiitic magmas (high-Ti group) are generally produced and emplaced or extruded proximal to the plume centre. If there is mantle plume involvement, a speculative potential plume centre may have been in the vicinity of western Meta Incognita Peninsula, proximal to the current location of the Big Island suture.

8.1.6.6 Broad geochemical similarities to LIPs

Almost all continental LIPs are volumetrically basic (<54 wt.% SiO₂), which is consistent with the chemistry of the Frobisher suite sills. The majority of continental-flood basalts also have low Mg# and low Ni contents, similar to A2 magmas. This is interpreted to result from extensive fractionation of primary picritic magmas (e.g. [Cox, 1980](#); [Thompson, 1983](#); [Garfunkel, 2008](#)).

8.1.6.7 LIP versus non-LIP?

Until further geochronological data is obtained or the Frobisher suite is linked to an already-existing LIP event, the Frobisher suite can tentatively be considered to be part of its own Large Igneous Province, informally called the “Frobisher LIP.” Given the approximate age of c. 1900 Ma obtained for the Frobisher LIP, then the Frobisher suite sills should be examined for their potential genetic relationship to LIPs in the region ranging in age from c. 1.85-1.95 Ga (see section 8.2.2).

8.1.7 Evidence for mantle plume-involvement

A non-essential, but common, aspect of LIPs is the involvement of an anomalously hot thermal anomaly (i.e. mantle plume), which allows for the vast amounts of melt production necessary for the magmatic event to be considered a LIP.

It is possible that the mafic (-ultramafic) sills and pillow-basalts found within the initial rifted margin sediments ([Wodicka et al., 2010](#)) of the Schooner Harbour sequence ([Sanborn-Barrie et al., 2008](#)) represent initial plume-derived rift magmatism, which was later followed by the magmatism that formed the sills of the Frobisher suite.

8.1.7.1 Field evidence

The extensive outcropping of the Frobisher suite sills within the Lake Harbour Group across most of southern Baffin Island supports the idea of mass magma production beneath the Meta Incognita microcontinent at ca. 1900 Ma, and subsequent emplacement into crustal strata. The most well recognized mechanism for large volumes of melt production is mantle plume impingement on the base of the lithosphere.

8.1.7.2 Geochemical evidence

The formation of ocean-island basalts has long been attributed to mantle plumes, which have a characteristic geochemical signature. The chondrite-

normalized REE patterns of OIBs globally were averaged by [Sun and McDonough \(1989\)](#) and can be used for comparison to identify plume-related magmatism. However, it should be noted that plume-derived LIPs can have highly variable chemistry, depending on numerous factors, such as contamination, crystal fractionation, degree of partial melting, depth of partial melting, plume composition, and other differentiation processes. Samples from the high-Ti group (Group D) have chondrite-normalized REE patterns that closely resemble that of OIBs, but are not quite as enriched as OIB. This could perhaps be due to the fact LIPs are the product of plume heads, whereas OIBs form from plume tails (i.e. less extensive melting occurs in tails, so lower-degree melting; [Richards et al., 1989](#)). Thus, the high-Ti group may share a similar source with OIBs (i.e. mantle plumes).

In LIPs such as the Wrangellia LIP, the high-Ti signature has been interpreted to be plume-related, unaffected by lithosphere. The low-Ti lavas from Wrangellia are HFSE-depleted and formed from plume head impingement on the base of the arc lithosphere ([Greene et al., 2009](#)). The equivalent magma type in the Frobisher for the low-Ti lavas of Wrangellia could potentially be the HFSE-depleted Groups B, C1, C2.

Differences in $(La/Yb)_N$ between the high-Ti and low-Ti groups can be most easily attributed to differences in degree of partial melting of plume-influenced mantle source. Their similar Zr/Nb ratios support this suggestion, as differences in degree of partial melting barely affect it. Therefore, it reflects the

character of the mantle source and low-Ti and high-Ti appear to share one. The similar spread of ϵNd values ($\sim 1\text{-}3$) also suggests this.

It was suggested by [Ernst & Buchan \(2003\)](#) that uncontaminated, basaltic rocks generated from mantle plumes should either have flat REE patterns or OIB-enriched patterns, both with an absence of Nb-Ta and Ti anomalies. This closely describes Group A1 and Group D, respectively. The only exception being weak contamination in Group A1 ($\text{La/Nb} = \sim 1\text{-}1.5$).

Regardless of the potential Paleoproterozoic uplift history recorded in southern Baffin Island, recent work by [Burov and Guillou-Frottier \(2005\)](#) suggests that the amount of uplift above a plume may be small or absent in some circumstances. Therefore, this should not influence or be used to disprove plume presence at 1900 Ma beneath the microcontinent.

Other observations supporting plume involvement include:

- (1) Mafic to ultramafic-dominated magmatic provinces are commonly associated with mantle plumes;
- (2) The presence of a plume component is the only widely accepted way to generate E-MORB magmatism (e.g. Group A1+A2; [Whitmarsh et al., 1996](#)), which is widely recognized as having resulted from mixing of depleted (i.e. N-MORB) and enriched (i.e. OIB) sources (e.g. [Serrano, 2011](#)).

Overall, the best geochemical indication for plume-involvement is the high-Ti Group D. It exhibits LREE-enrichment and fractionated HREE, similar to OIB-sourced magmas, and it has generally high MgO (mostly $>18\text{wt.}\%$). Furthermore,

the presence of OIB-sourced magmas indicates the involvement of a source deeper than that of the MORB upper mantle source (Ernst, 2014).

Finally, it should be noted that OIB signatures can also be produced by non-plume processes, such as volatile-induced melting of ambient mantle and lithospheric mantle fracture-transport (Herzberg and Gazel, 2009). But high degrees of melting (e.g. Group A1+A2) require a thermal anomaly such as a plume (Herzberg and Gazel, 2009).

8.1.8 Sill emplacement

Sills of the Frobisher suite appear to have been emplaced as both single injections of a single magma type, and as multiple sheet-like intrusions (e.g. Duluth Complex; Miller et al., 2002), as the variation in chondrite-normalized REE patterns cannot be accounted for by closed-system in-situ differentiation. The intrusions involving multiple sill injections may have either formed composite sills (i.e. sills emplace inside one another), or a stacked sequence of sills, in which each successive sill may have underplated the last. In the latter case, it is possible that within some sills where the mafic and ultramafic components appear to represent different magma types, an ultramafic sill may have preceded the injection of the mafic component, meaning that certain sills that appear to be overturned may actually be the right way up.

Given the evidence for multiple injections of different magma types into the same sill, it is highly likely that all magma types utilized the same established melt transport network, as one of the critical elements of an established melt

transport network is that it is susceptible to reopening and exploitation by subsequent magma pulses ([Marsh, 2004](#)).

At this point in time, there is no dyke swarm that has been linked to the Frobisher suite, so the plumbing system to this suite of sills is not yet fully understood. This is one line of evidence that might be used to support the passive rift-related origin of the Frobisher suite. However, the apparent absence of dyke-like feeders to the sills could be due to a lack of preservation, and the few that might be exposed at surface may not have been identified during regional mapping. Alternatively, the absence of dykes may also indicate that the suite was not fed directly from a single central plume region, that sills were fed by other sills, such as in the Karoo basin of South Africa ([Svensen et al., 2012](#)), or that sills were fed by dykes outside the sedimentary basin that are emplaced as sills once the melt enters the sedimentary basin (e.g. Ernst, 2014).

Within individual sills, there is a general absence of observed assimilated metasedimentary material (e.g. xenoliths, granophyric margins), with the Chidliak sill as the single exception. The lack of crustal xenoliths may suggest the flow was more laminar, as opposed to turbulent. The absence of granophyric margins may indicate the host rocks were cold, therefore requiring substantial thermal energy to push them past the solidus and generate partial melting. This is consistent with the Frobisher suite's emplacement in a sedimentary basin (<~10 km thick).

Due to relatively thin nature of many sills within the Frobisher suite (e.g. <50 m), they would have cooled relatively fast (i.e 50% crystallized at ca. 10 yrs

and completely solid at ca. 20 yrs, as per general estimates of cooling rates of sills by [Cawthorn, 2012](#)), not allowing enough time for significant in situ differentiation and assimilation processes to take place. The sills also formed chilled margins at their contacts, which prevented heat transfer to the country rocks, which suggests that the variable amounts of contamination that occurs in certain groups (e.g. Group A2) occurred at depth, prior to final emplacement.

With regard to overall geometry of the Frobisher suite at emplacement, the sills were more likely emplaced as a sill province, similar to sills of the Karoo basin (of the Karoo-Ferrar LIP), as opposed to a dismembered stratiform sill, like those of the Giles Complex of central Australia (of the Warakurna LIP). This notion is supported by: (1) Sills present at different levels within the Lake Harbour Group stratigraphy, as well as difference in composition between those emplaced in proximal (furthest from edge of microcontinent) and distal facies (closest to edge of microcontinent); (2) The diversity of geochemical groups within sills; (3) The close proximity of sills derived from different magma types; (4) High variation in the mineralogy of different sills, however, this may partially be due to regional differences in metamorphic grade (variable from amphibolite to granulite facies); and (5) the absence of shear zones at the along-strike terminations of the sills.

As mentioned above, sills derived from high-Ti magmas occur only in the proximal LHG facies (i.e. in southern to western Meta Incognita Peninsula), while low-Ti magma-derived sills are concentrated in the distal LHG facies (i.e. on northern Hall Peninsula). The occurrence of the low-Ti magmas in the proximal LHG facies could occur as a result of any of the following: (1) interconnecting

lithospheric faults carrying low-Ti magmas further inland; (2) low-Ti magmas travelling great distances within dykes or sills to the more proximal facies; or (3) intense erosion of the subcontinental lithospheric root, such that melting below the Meta Incognita microcontinent occurs at increasing shallower depths, to the point where high-degrees of melting are occurring within the spinel-stability field.

8.1.9 Post-emplacement modification

Lithologies that are mafic to ultramafic in composition can be highly resistant to metamorphism if anhydrous conditions are maintained, even up to high grades. The pervasiveness of hornblende, and lesser biotite, throughout the sills indicates that these sills were thoroughly hydrated, and consequently recrystallized, during the high-grade metamorphic events of the Trans-Hudson Orogen (THO). Influx of water was likely facilitated by post-emplacement deformation within the sills and adjacent host rocks, as evidenced by sills forming F_3 and F_4 synclines and anticlines, where overturning of stratigraphy has occurred in some cases. The present lithologies are metamorphic in nature, however, relict primary features such as magmatic layering and pristine cores in phenocrysts (e.g. clinopyroxene in hornblende) are often present.

During the progressive closure of the Manikewan Ocean, the Meta Incognita microcontinent was thrust upward onto the Narsajauq arc. Through this process, the leading edge of the microcontinent was eroded away, leaving the trailing half of the microcontinent preserved, which today is bounded between the

Big Island and Baffin sutures, representing the approximate transition from the central region of the microcontinent to the edge of it, respectively.

8.1.10 Metamorphic Character of sills

Given the pervasiveness of upper amphibolite- to granulite-facies metamorphic assemblages throughout Lake Harbour Group (which host the sills) and later intruding phases of Cumberland Batholith, the mineral assemblages within the Frobisher suite sills are considered to be metamorphic. Beyond the regional context, further evidence for the metamorphic origin of most minerals in this mafic to ultramafic suite of sills includes: 1) mineral assemblages within both mafic and ultramafic units are consistent with amphibolite- to granulite-facies metamorphic conditions; 2) weakly to strongly developed foliation fabric, defined by amphibole (and biotite, when present), is present in numerous samples; 3) the presence of a granoblastic texture present in many samples, typical of non-foliated, high-grade metamorphic rocks; 4) the presence of amphibole in every sample, with the exception of one; 5) the coexistence of olivine (anhydrous) and amphibole (hydrous) in rocks where the former is only very weakly serpentinized along fractures; 6) An-rich character (>85% An) of plagioclase grains from numerous sills, indicative of high-grade metamorphism; 7) spinel, of similar chemical composition to all others analyzed, exhibiting a vermicular texture, which is characteristic of formation through metamorphic growth. However, the presence of relict primary minerals is permissible given the chemistry of select mineral phases (e.g. olivine within field of layered intrusion olivine compositions,

plagioclase grains of andesine to labradorite compositions). The most likely candidates for samples retaining primary magmatic mineral assemblages (or primary minerals) are those which lack all or most of the observations stated above.

8.2 Regional Implications

8.2.1 Regional comparison with other LIP and non-LIP events

The recent unpublished age of 1900 Ma is considered to be a minimum crystallization age for the suite (N.Rayner, unpublished data, 2017). With this age, the maximum age of the Lake Harbour Group (ca. 1934 $\pm$ 2 Ma; Scott, 1997), the age of the M1a granulite-facies metamorphic event (ca. 1845-1842 Ma; see chapter 3), and the knowledge that LIP magmatism can have a maximum duration of 10s of myr, it is possible that the Frobisher suite is related to LIP events that occurred in the timeframe between ca. 1845 Ma and ca. 1934 Ma (Figure 8.9). Currently recognized LIPs, and interpreted-LIP fragments, that were emplaced (or extruded) during the ca. 1934 - 1845 Ma time period are listed in Table 8.1 and their locations are shown in Figure 7.9.

Given the potential size of a flattened plume head (i.e. up to 2000 km across), additional magmatism from the same event may have occurred away from the Meta Incognita microcontinent and emplaced elsewhere. In modern day plate geometry, this is approximately the equivalent to the distance from Iqaluit, Baffin Island, Nunavut, to Ottawa, Ontario, which is precisely 2088 km (calculated by software on <http://www.distancefromto.net/>). With this in mind, it

does not seem unreasonable to entertain the idea of a relationship between Frobisher suite magmatism and another LIP- or non-LIP-scale event greatly displaced from the centre of magmatism beneath the Meta Incognita microcontinent at ca. 1900. Furthermore, given that the Manikewan Ocean closed within a short time interval after Frobisher suite magmatism, there is a good chance that any other mafic-ultramafic suites related to this LIP event are contained within Baffin Island, or proximal to it (e.g. northern Québec; Greenland). In addition to this, it is well-recognized that magma can travel great distances *via* dykes and sills (>2000 km) before finally being emplaced or erupted (e.g. Mackenzie dyke swarm of the Mackenzie LIP; [Baragar et al., 1996](#)). Due to a lack of available geochemical data in literature for most LIPs that occurred between 1925 and 1875, only comparative geochemistry between the c. 1880 Ma Circum-Superior LIP and Frobisher LIP is presented.

Through the use of primitive-normalized trace element patterns, the fields of Circum-Superior Province volcanic and intrusive rocks were compared with the average values from the Frobisher suite geochemical groups. While no exact matches in geochemical patterns were observed, select groups from the Frobisher suite show some similarities to certain Circum-Superior LIP groups ([Figure 8.10](#)), mainly based in their $[(La/Yb)_N]$ ratios. In all cases, the Circum-Superior LIP contains greater total REE contents than the Frobisher suite geochemical groups:

- (1) Group B (averaged) demonstrates a similar pattern to the ca.1874 Ma Gunflint Formation ([Fralik et al., 2002](#)) of the Circum-Superior LIP ([Figure](#)

- 8.10a). The primary differences are negative Zr-Hf anomalies in average Group B, as well as its much more prominent negative Nb-Ta anomalies.
- (2) Group F (averaged) demonstrates a similar pattern to the ca. 1864-1891 Ma Winnipegosis Belt (Hulbert et al, 1994; Burnham et al., 2004) of the Circum-Superior LIP (Figure 8.10b). The primary difference is that Group F of the Frobisher suite shows a negative Nb anomaly, whereas the Winnipegosis Belt exhibits positive Nb-anomalies, signifying a contaminated and non-contaminated geochemical pattern, respectively.
- (3) Group F (averaged) demonstrates a similar pattern to petrographic group 3 (i.e. plagioclase-phyric basalts) of the ca. 1882 Ma Chukotat Formation (Wodicka et al., 2002; Randall, 2005) of the Circum-Superior LIP (Figure 8.10c). The primary difference is that Group F of the Frobisher suite shows a negative Nb anomaly, whereas the Chukotat Formation exhibits positive Nb-anomalies, signifying a contaminated and non-contaminated geochemical pattern, respectively.

Given the similar age and presence of geochemical similarities amongst these two LIPs, further work should be conducted to determine if the Frobisher LIP and Circum-Superior LIP are in-fact related. In the event that the Frobisher and Circum-Superior LIPs are related, this would have implications in expanding the areal extent of the c. 1880 Ma Circum-Superior LIP. In addition, further comparative work should be conducted between the Frobisher LIP and other LIPs, pending availability of geochemical data from the latter. Additional high-precision U-Pb geochronology should be conducted to better constrain the age of

Frobisher suite magmatism, and greater narrow the number of possible related LIP events. Furthermore, the compositional ranges of the Frobisher suite can be seen alongside the geochemistry of some major LIPs in [Table 8.2](#). LIPs exhibiting significant similarities with the Frobisher LIP in each respective geochemical parameter are highlighted in bold.

8.3 Model for Frobisher Suite Formation

Based on all the evidence presented in this thesis, but primarily, earlier in this chapter, it appears that the Frobisher suite mafic, ultramafic and mafic-ultramafic sills were derived from the melting of a mantle plume head. A plan view of the tectonic setting is illustrated in [Figure 8.11](#), and a cross section is in [Figure 8.12](#). A detailed model of Frobisher suite formation is presented below, in four stages, with no specific time interval separating any of them. The model is also illustrated in [Figure 8.13](#).

8.3.1 Pre-Stage 1

Prior to the initial stage of the plume-based model, the Meta Incognita microcontinent was rifted from an established continental margin, and the Schooner Harbour sequence was deposited during this initial stage of rifting ([Wodicka, 2010](#); [N. Wodicka, pers. comm., 2017](#)). Following this rifting event, the attendant shelf succession to the microcontinent, the Lake Harbour Group, was established.

8.3.2 Stage 1

In stage 1 ([Figure 8.13](#)), a mantle plume departs from the core-mantle boundary, or somewhere within the lower mantle, and begins to ascend upward toward the base of the Meta Incognita microcontinent.

8.3.3 Stage 2

In stage 2 ([Figure 8.13](#)), the mantle plume impinges upon the base of the subcontinental lithospheric mantle (SCLM), beneath the Meta Incognita microcontinent, and flattens.

8.3.4 Stage 3

In stage 3 ([Figure 8.13](#)), low-degree partial melting occurs at the base of the lithosphere (within the garnet-stability field) as a result of a reduction in pressure (adiabatic decompression melting), which reduces the solidus of the plume-head material, allowing it to melt. This produces the primary magmas to the high-Ti group (Group D). The low-degree melt then begins to pool at the base of the lithosphere, and some high-Ti magma ascends through the thicker portion of the lithosphere, without experiencing significant fractionation or contamination, and directly into the crust where it is emplaced as sills within proximal facies strata of the Lake Harbour Group. Due to the buoyancy of the plume head, the plume material begins to traverse the base of the SCLM to shallower levels.

8.3.5 Stage 4

In stage 4 (Figure 8.13), the plume head material reaches shallower depths beneath the most distal facies (seaward) sequence of the Lake Harbour Group. It is there where a significantly higher degree of melting occurs as a result of continued adiabatic decompression melting. This results in the generation of low-Ti primary magmas (i.e. Groups A1 and A2). Low-Ti magmas pass through the lithosphere and pool at the base of the crust, where fractionation occurs, lowering the Mg# of most magmas and depleting them of Ni and Cr. After extensive fractionation and variable contamination, the low-Ti magmas ascend further to the level of emplacement within the Lake Harbour Group as sills. The high levels of thermal energy introduced to the base of the lithosphere, by the mantle plume, allow for thermal erosion of the metasomatized component of the SCLM, which has a significantly lower solidus temperature due to its hydrous mineral assemblage (e.g. amphibole, phlogopite), producing primary magmas to Group C2. After these Group C2 magmas are generated, they begin to mix with both high- and low-Ti magmas, producing Group C1 and Group B, respectively. These magmas ascend through the lithosphere and are also emplaced within the Lake Harbour Group as sills. High-degrees of melting of mixed plume head and depleted upper mantle material likely led to the production of the apparently limited Group F magmas.

8.4 Mineral Deposit Potential of Frobisher Suite

By virtue of their LIP affiliation and apparent plume-associated characteristics, there is considerable mineralization potential within the Frobisher

suite and their adjacent host rocks. Large magmatic events may cause increased heat flux into the crust, which may trigger crustal melting, devolatilization and large-scale fluid flow (Ernst and Jowitt, 2013; Maier, 2015). Deposits commonly associated with these conditions include magmatic PGE-Cr-V-Fe-P deposits within layered intrusions, magmatic Ni-Cu sulfide deposits in magma feeder conduits or at the basal portion of intrusions, and various hydrothermal deposits in the host rocks (Ernst and Jowitt, 2013; Maier, 2015).

8.4.1 Field and petrographic observations

Sulphides were observed in hand sample and in thin section, from almost every sample collected from the Frobisher suite sills. In samples with high sulphide abundances, associated gossans were noted in the field. Sulphide phases are dominantly pyrite and pyrrhotite, however, minor amounts of chalcopyrite are often intergrown with pyrrhotite grains. A number of sills also contain variable amounts of pentlandite associated with pyrrhotite and chalcopyrite.

A primary magmatic origin for these sulphides is suspected, in most cases, based on their textural relationships with the main silicate phases in the samples. In some samples sulphide grains exhibit a round to ovoid shape, and in most other cases sulphides are anhedral, forming interstitially between silicate mineral phases. The thicknesses of sills is quite variable, however, most are relatively thin. This may pose an issue, as thinner sills cool much more rapidly than their thicker equivalents (Cawthorn, 2012), meaning that there may not have

been sufficient time (and thermal energy) to assimilate crustal material in-situ, segregate immiscible sulphide blebs, and accumulate them at the base of the sill (effectively increasing the tenor of the prospective ore). However, if any sills had a high magma flux, complete crystallization would be delayed and any sulphides formed within the sill, or prior to emplacement and subsequently entrained by the injected magma, would significantly increase in tenor due to extensive interaction with large volumes of magma ([Barnes et al., 2015](#)).

8.4.1.1 14SAB-L009 gossan

The largest gossan identified (lat: 63.22945021, long:-71.55833542) is associated with high-Ti group D, and is located in western Meta Incognita, at station 14SAB-L009. Stakes observed on the outcrop indicate that the area is currently, or was previously, staked by an exploration company. The bright orange-weathering gossan ([Figure 2.6a](#)) appears mostly within the LHG psammitic unit directly adjacent to the sill, however, visible mineralization occurs both within the sill and the metasedimentary rocks. Sulphides within the sill appear in both mafic ([Figure 8.14a](#)) and ultramafic ([Figure 8.14b](#)) components and interpreted to have formed magmatically. However, the mineralized psammite ([Figure 8.14c](#)) is presumed to have formed post- or syn-magmatically, through hydrothermal processes remobilizing primary magmatic sulphide mineralization. The most mineralized samples obtained from the LHG are mostly comprised of coarse-grained quartz, with round to ovoid shaped sulphide grains. Within the sill, sulphides are quite anhedral, forming interstitially between

plagioclase and clinopyroxene grains. The presence of graphite in the gabbroic part of the sill indicates remobilization of elemental components within the psammite, and perhaps within sill lithologies as well. Sulphides within both the mafic (gabbroic) component of the sill and the psammite comprise primarily pyrrhotite, with minor amounts of chalcopyrite. Within the ultramafic (pyroxenitic) component of the sill, the same sulphide assemblage exists, but pentlandite is also present, in abundances less than both pyrrhotite and chalcopyrite. Given the sulphide assemblage within the sill, the mineralization within this sill is considered to be Cu-Ni (-PGE).

8.4.2 Geochemical perspective

It has been recognized that when multiple magma suites are contained within the same province, Ni-Cu-PGE ores are characteristically associated with the suite that was generated by the greatest degree of partial melting ([Zhou et al., 2008](#); [Godel et al., 2011](#)). A reliable proxy, for example, is low Ti content ([Zhou et al., 2008](#); [Godel et al., 2011](#)), which is characteristic of Groups A1, A2, B, C1, C2, E, and F, and suggests that Group D is the least prospective for Ni-Cu-PGE mineralization.

The tholeiitic low-Ti Groups A1 and A2 are believed to represent high-degree melts (i.e. larger volume of melt) from the plume source that rose to a thin lithospheric area, whereas the mildly alkaline high-Ti Group D is suggested to have been extracted from the source at far deeper levels and to a lower degree (i.e. smaller volume). Because of the high degree of melting experienced by the

primary magma to the low-Ti groups, they are more enriched in Ni, Cu and PGEs (Keays, 1995; Rehkämer et al., 1999) compared to Group D, and are thus more capable of producing high-tenor ores. However, some authors suggest that low-degree partial melting can produce magmas that may be prospective for Ni-Cu-PGE sulfide mineralization (e.g., Mungall et al., 2006; Naldrett, 2010). The presence of a significant gossan (L009; Figure 3.6a) in a Group D sill is attributed to highly sulphidic LHG host rocks, and the low-tenor of the sulphides within it is consistent with the low prospectivity of high-Ti Group D.

8.4.2.1 Chalcophile element geochemistry

In Figure 8.15, the Ni, Cu, Pt, and Pd compositions of samples are plotted against Mg#. Prior to assessment of the concentrations of these chalcophile elements, it must be noted that these particular chemical species can be remobilized via hydrothermal fluids, and thus, current elemental abundances of chalcophile elements may not reflect the original abundance in the unmetamorphosed, and unaltered, primary magmatic rock.

Samples from Groups A1 and D generally contain the highest Ni concentrations (1200-1500 ppm and 800-1000 ppm, respectively), mainly because they are ultramafic in nature. Groups C2 and F display a range of Ni contents spanning from high (900-1000 ppm) to low (0-200 ppm) concentrations. Groups A2 and E exhibit quite low Ni concentrations (<200 ppm), generally.

About half of all the samples have Pt concentrations below detection and the majority of samples have Pd contents above detection, but are considered

low in concentration (i.e. <30 ppb). Amongst those samples with detectable concentration of Pt and Pd, distinct differences can be seen between geochemical groups. Groups A1 and D exhibit relatively low concentrations of Pt and Pd compared to other groups above detection. All of Group A1 is above detection, while the majority of Group D samples are below detection in Pt and a few fall below detection in Pd. Pt and Pd concentrations are distinctly higher in Groups A2 and F, with the least evolved Group F samples attaining Pt and Pd concentrations of 25 and 37 ppb, respectively. Most Group C1 and C2 samples have Pt levels below detection, but Pd levels are most above it. Those samples with Pt and Pd above detection in Groups C1 and C2 show similar concentrations to those observed in Group D, with few of the least evolved Group C2 samples with concentrations of Pt and Pd up to 25 and 50 ppb, respectively. All Group B and E samples fall beneath the detection limit.

Groups B, C1, D and E exhibit a wide range of Cu values, mostly ranging between 0 and ~140 ppm, with some concentrations in the most evolved samples up to ~170, 180, 220 and 280 ppm, respectively. These Cu contents are considered generally low. Groups A1, A2, C2 and F contain relatively low concentration of Cu concentrations, generally falling between 0 and 800 ppm, with a few Group A2 samples up to ~130 ppm and Group F samples up to ~170 ppm. In Groups D and C1, Cu content increases as samples evolve (decreasing Mg#), consistent with Cu's incompatible behaviour in sulfur-undersaturated magmas. In Groups A2 and C2, Cu stays at an approximately consistent concentration through evolution.

In the $\log(\text{Cu}/\text{Pd})$ versus $\text{Mg}\#$ diagram ([Figure 8.16](#)), samples display a wide range in Cu/Pd ratios, with some Frobisher suite sills being PGE-depleted relative to primitive mantle, and others Pd-enriched. Group C1 and D samples appear to be mostly depleted in PGEs relative to primitive mantle, while Group C2 and F samples are PGE-enriched. Samples from the low-Ti groups (Groups A1+A2) show both PGE-depletion and -enrichment relative to primitive mantle, however, Group A2 samples show PGE-depletion and -enrichment to greater degrees, while Group A1 Cu/Pd ratios straddle that of the primitive mantle. The PGE-depleted nature of Group C1 and D samples could reflect a number of things, including magmatic equilibration with sulphide prior to final emplacement of their parental magmas, assimilation of Cu-rich crust, relative enrichment in Cu in the mantle source ([Maier, 2015](#)), PGE remaining in the mantle, or modification of the Cu/Pd ratio by remobilization of Cu and Pd. The PGE-enrichment in Groups C2 and F could reflect precipitation of immiscible sulphide droplets during magma mixing, sulfide liquid saturation resulting from magmatic fractionation or initiated by crustal contamination ([Maier, 2015](#)), or modification of the Cu/Pd ratio by remobilization of Cu and Pd. In the case of Group D, it is also possible that PGE-depletion in samples resulted from the small-degree of mantle melting (i.e. PGE retained in mantle source), which is evident in the generation of its primary magmas.

Based on this preliminary assessment, it appears that the tholeiitic low-Ti groups serve as the best targets for Ni-Cu-PGE exploration, and have the greatest potential to host high-tenor magmatic sulphide mineralization.

8.5 Conclusions

This study on the Frobisher suite has revealed considerable variation in size, mineralogy and geochemistry within individual sills, and has led to the following main conclusions:

1. Eight different geochemical groups were recognized within the Frobisher suite, and include Groups A1, A2, B, C1, C2, D, E and F. High-Ti (Group D) and Low-Ti (Group A1+A2) magmas appear to be derived from the same source, but reflect differences in degree of partial melting and perhaps an eclogitic or basaltic components in the former, generating its consistent positive Nb-Ta anomaly. Group D magma was generated by low-degree partial melting beneath a thick portion of lithosphere beneath the Meta Incognita microcontinent (within the garnet stability field). This resulted in LREE-enrichment and fractionated HREE. As magma ascended to shallower levels into the spinel-stability field, via sub-lithospheric channeling, a higher degree of partial melting was achieved, resulting in significantly flatter trace element contents with flat HREE. High-Ti magmas arrived more directly from the mantle compared to low-Ti magmas, except for Group A1, which appears largely undifferentiated.
2. Group C2 magmas are derived from direct melting of previously metasomatized SCLM, likely triggered by the addition of thermal energy via mantle plume, bringing it above the solidus. The source may have experienced a previous partial melting event, as per the generally depleted

HREE contents, and was subsequently enriched by slab-derived fluids, prior to the plume arriving to produce the Frobisher LIP event.

3. Parental magmas of Groups B and C1 are interpreted to reflect interaction of Group A1+A2 and Group D magmas with SCLM- or crustally-contaminated Group C2 magmas.
4. Sills of the Frobisher suite appear to have been emplaced as both single injections of a single magma type, and as multiple injections of either single or multiple magma types.
5. On the basis of sufficient geographic areal extent ($>100,000 \text{ km}^2$), high MgO content of the majority of samples, chondrite-normalized REE patterns trace element patterns, the significant ultramafic component to the suite, provinciality of magma types (i.e. high- and low-Ti), and other geochemical characteristics, it is proposed that the Frobisher suite represents the plumbing system of a newly-recognized mantle plume-derived LIP, tentatively called the “Frobisher LIP.”
6. Based on this preliminary assessment, it appears that the tholeiitic low-Ti groups serve as the best targets for Ni-Cu-PGE exploration, and have the greatest potential to host high-tenor magmatic sulphide mineralization.

References

- Arndt, N.T., Albarede, F., and Nisbet, E.G. (1997). Mafic and ultramafic magmatism, in de Wit, M.J. and Ashwal, L. D., eds., *Greenstone Belts*: Oxford University Press, Oxford, p. 233-254.
- Arndt, N.T., Czamanske, G.K., Wooden, J.L. and Fedorenko, V.A. (1993). Mantle and crustal contributions to continental flood volcanism. *Tectonophysics*, 223: 39–52.
- Ballhaus, E.G. and Glikson, A.Y. (1989). Magma mixing and intraplutonic quenching in the Wingellina Hills Intrusion, Giles Complex, central Australia. *Journal of Petrology*, 30, 1443-1469.
- Baragar, W.R.A., Ernst, R.E., Hulbert, L. and Peterson, T. (1996). Longitudinal petrochemical variation in the Mackenzie dyke swarm, northwestern Canadian Shield. *Journal of Petrology*, 37: 317–359.
- Bédard, J.H. (2001). Parental magmas of the Nain Plutonic Suite anorthosites and mafic cumulates: a trace element modelling approach. *Contributions to Mineralogy and Petrology* 141 (6), 747–771.
- Belica, M.E., E.J. Piispa , J.G. Meert , et al. (2014). Paleoproterozoic mafic dyke swarms from the Dharwar craton; paleomagnetic poles for India from 2.37 to 1.88 Ga and rethinking the Columbia supercontinent. *Precambrian Research*, 244: 100–122.
- Bleeker, W. (2005). North America: Precambrian continental nucleus. In *Encyclopedia of geology*. Edited by R.C. Selley, L.R.M. Cocks and I.R. Plimer. Elsevier, Oxford, UK., pp. 8–21.

Buchan, K.L., Ernst, R.E., Bleeker, W., Davis, W.J., Villeneuve, M., van Breemen, O., Hamilton, M., Söderlund, U., 2010. Proterozoic magmatic events of the Slave Craton, Wopmay Orogen and Environs. Geological Survey of Canada Open File 5985.

Large Igneous Provinces and supercontinents: Toward completing the plate tectonic revolution (PDF Download Available). Available from: https://www.researchgate.net/publication/256806733_Large_Igneous_Provinces_and_supercontinents_Toward_completing_the_plate_tectonic_revolution [accessed Jun 19, 2017].

Bucher, K., & Grapes, R. H. (2011). Petrogenesis of metamorphic rocks (8th ed.). Berlin ; New York: Springer.

Burnham, O.M., Halden, N.M., Layton-Matthews, D., Leshner, C.M., Liwanag, J., Heaman, L.M., Hulbert, L.J., Machado, N., Michalak, D., Pacey, M., Peck, D.C., Potrel, A., Theyer, P., Toope, K., Zwanzig, H.V. (2004). Geology, stratigraphy, petrogenesis, and metallogenesis of the Thompson Nickel Belt, Manitoba: final report for CAMIRO Project 97E-02. Mineral Exploration Research Centre, Laurentian University, Sudbury. 410pp.

Burov, E. and Guillou-Frottier, L. (2005). The plume head-continental lithosphere interaction using a tectonically realistic formulation for the lithosphere. *Geophysical Journal International*, 161: 469–490.

Condie, K.C., Des Marais, D.J., Abbott, D. (2000). Geologic evidence for a mantle superplume event at 1.9 Ga. *Geo-physics, Geochemistry, Geosystems*, Vol. 1. Paper No. 2000GC000095.

-
- Cox, K.G. (1980). A model for flood basalt volcanism. *Journal of Petrology*, 21: 629–650.
- Cox, K.G. (1993). Continental Magmatic Underplating. *Philosophical Transactions of the Royal Society, A* 1993 342 155-166; DOI: 10.1098/rsta.1993.0011.
- Drummond, M. S., and Defant, M. J. (1990). A model for trondhjemite-tonalite-dacite genesis and crustal growth via slab melting: Archean to modern comparisons: *Journal of Geophysical Research*, v. 95, no. B13, p. 21503-21521.
- Egorova, V. and Latypov, R. (2013). Mafic–ultramafic sills: New Insights from M- and S-shaped mineral and whole-rock compositional profiles. *Journal of Petrology* 54, 10, 2155–2191.
- Ernst, R.E. (2014). *Large Igneous Provinces*: Cambridge, UK, Cambridge University Press, pp. 653.
- Ernst, R.E. and Bell, K. (2010). Large Igneous Provinces (LIPs) and carbonatites. *Mineralogy and Petrology*, 98: 55–76.
- Ernst, R.E. and Bleeker, W. (2010). Large igneous provinces (LIPs), giant dyke swarms, and mantle plumes: significance for breakup events within Canada and adjacent regions from 2.5 Ga to present: *Canadian Journal of Earth Sciences*, v. 47, p. 695-739, doi: 10.1139/E10-025.
- Ernst, R.E. and Buchan, K.L. (2003). Recognizing mantle plumes in the geological record. *Annual Reviews Earth and Planetary Science*, 31: 469–523.

- Ernst, R.E. and Buchan, K.L. and Campbell, I.H. (2005). Frontiers in large igneous province research. *Lithos*, 79: 271–297.
- Ernst, R.E., and Jowitt, S.M. (2013). Large Igneous Provinces (LIPs) and metallogeny, in *Tectonics, metallogeny, and discovery: The North American Cordillera and similar accretionary settings* edited by M. Colpron, T. Bissig, B.G. Rusk and J.F.H. Thompson: Society of Economic Geologists Special Publication 17, p. 17–51.
- Ernst, R.E., Hamilton, M.A., Söderlund, U., Hanes, J.A., Gladkochub, D.P., Okrugin, A.V., Kolotilina, T., Mekhonoshin, A.S., Bleeker, W., LeCheminant, A.N., Buchan, K.L., Chamberlain, K.R., & Didenko, A.N. (2016). Long-lived connection between southern Siberia and northern Laurentia in the Proterozoic. *Nature Geoscience* (in press), doi: 10.1038/NGEO2700.
- Ernst, R.E., Srivastava, R.K. (2008). India's place in the Proterozoic world: constraints from the large igneous provinces (LIP) record. In: R.K. Srivastava, Ch. Sivaji, and N.V. Chalapathi Rao (eds.) *Indian Dykes: Geochemistry, Geophysics, and Geochronology*, Narosa Publishing House Pvt. Ltd, New Delhi, India, pp. 41-56.
- Ernst, R.E., Youbi, N. (2017). How Large Igneous Provinces affect global climate, sometimes cause mass extinctions, and represent natural markers in the geological record. *Palaeogeography, Palaeoclimatology, Palaeoecology* v. 478, p. 30-52.

-
- Ewart, A., Marsh, J.S., Milner, S.C., Duncan, A.R., Kamber, B.S. and Armstrong, R.A. (2004). Petrology and Geochemistry of Early Cretaceous Bimodal Continental Flood Volcanism of the NW Etendeka, Namibia. Part 1: Introduction, Mafic Lavas and Re-evaluation of Mantle Source Components. *J Petrology*, 45 (1): 59-105. doi: 10.1093/petrology/egg083
- Ewart, A., Milner, S.C., Armstrong, R.A, and Duncan, A.R. (1998). Etendeka volcanism of the Goboboseb Mountains and Messum Igneous Complex, Namibia. Part I: geochemical evidence of Early Cretaceous Tristan plume melts and the role of crustal contamination in the Paraná–Etendeka CFB. *Journal of Petrology*, 39: 191–225.
- Fralick, P.W., Davis, D.W., Kissin, S.A. (2002). The age of the Gunflint Formation, Ontario, Canada: single U-Pb age determinations from reworked volcanic ash, *Canadian Journal of Earth Science*, 39, pp. 1085–1091.
- French, J.E. , Heaman, L.M. , Chacko, T. , & Srivastava, R.K. (2008). 1891–1883 Ma Southern Bastar–Cuddapah mafic igneous events, India: a newly recognized large igneous province. *Precambrian Research*, 160: 308–322.
- Gladkochub, D.P, Pisarevsky, S.A, Donskaya, T.V, et al. (2010). Proterozoic mafic magmatism in Siberian craton: an overview and implications for paleocontinental reconstruction. *Precambrian Research*, 183: 660–668.
- Godel B., Barnes, S.-J. and Wolfgang M.D. (2011). Parental magma composition inferred from trace element in cumulus and intercumulus silicate minerals :

-
- an example from the Lower and Lower Critical Zones of the Bushveld Complex, South-Africa. *Lithos*, 125, (1-2), p. 537-552.
- Greene, A.R., Scoates, J.S., Weis, D., Nixon, G.T. and Kieffer, B. (2009). Melting history and magmatic evolution of basalts and picrites from the accreted Wrangellia oceanic plateau, Vancouver Island, Canada. *Journal of Petrology* 50: 467–505.
- Hanson, R.E., Crowley, J.L., Bowring, S.A., et al. (2004a). Coeval large-scale magmatism in the Kalahari and Laurentian cratons during Rodinia assembly. *Science*, 304: 1126–1129.
- Hanson, R.E., Crowley, J.L., Bowring, S.A., et al. (2004b). Coeval large-scale magmatism in the Kalahari and Laurentian cratons during Rodinia assembly. *Science*, 304: 1126–1129.
- Herzberg, C. and Gazel, E. (2009). Petrological evidence for secular cooling in mantle plumes. *Nature Publishing Group* 619–622.
- Higgins, M.D., and Chandrasekharam, D. (2007). Nature of sub-volcanic magma chambers, Deccan province, India: Evidence from quantitative textural analysis of plagioclase megacrysts in the giant plagioclase basalts: *Journal of Petrology*, v. 48, p. 885–900, doi: 10.1093/petrology/egm005.
- Hole, M.J. and Saunders, A.D. (1996). The generation of small melt-fractions in truncated melt columns: constraints from magmas erupted above slab windows and implications from MORB genesis: *Mineralogical Magazine*, v. 60, p. 173–189.

- Hulbert, L.J., Kyser, K., Carlson, R., Leshner, C.M., Joudrie, C. (1994). The Winnipegosis Komatiites; a New Komatiite Belt, Central Manitoba. Minerals Colloquium Program with Abstracts Geological Survey of Canada, Ottawa.
- Huppert H. E. and Sparks R. S. J. (1985a) Cooling and contamination of mafic and ultramafic magmas during ascent through continental crust. *Earth Planetary Science Letters* 74, 371–386.
- Ito, G. and Mahoney, J.J. (2005). Flow and melting of a heterogeneous mantle: 1. Method and importance to the geochemistry of ocean island and mid ocean ridge basalts. *Earth and Planetary Science Letters* 230, pp. 29–46.
- Jourdan, F., Bertrand, H., Sharer, U., Blitchert-Toft, J., Féraud, G., and Kampunzu, A.B. (2007). Major-trace element and Sr–Nd–Hf–Pb isotope compositions of the Karoo large igneous province in Botswana–Zimbabwe. *Journal of Petrology*, 48: 1043–1077.
- Keays, R.R. (1995). The role of komatiitic and picritic magmatism and S-saturation in the formation of ore deposits. *Lithos*, 34: 1–18.
- Le Bas, M.J. (2000). IUGS Reclassification of the high-Mg and picritic volcanic rocks: *Journal of Petrology*, v. 41, p. 1467-1470.
- Liikane, D.A., St-Onge, M.R., Kjarsgaard, B.A., Rayner, N.M., Ernst, R.E. and Kastek, N. (2015). Frobisher suite mafic, ultramafic and layered mafic-ultramafic sills, southern Baffin Island, Nunavut; in *Summary of Activities 2015*, Canada-Nunavut Geoscience Office, p. 21–32.
- Maier, W., Rasmussen, B., Fletcher, I., Godel, B., Barnes, S.J., Fisher, L., and

-
- Yang, S. (2015). Petrogenesis of the ~2.77 Ga Monts de Cristal complex, Gabon: Evidence for direct precipitation of Pt-arsenides from basaltic magma. *Journal of Petrology*, 56, 1285–1308.
- Marsh, B. (2004). A magmatic mush column Rosetta Stone: the McMurdo dry valleys of Antarctica. *EOS Transactions, American Geophysical Union* 85:497–508.
- Martin, H., Smithies, R.H., Rapp, R., Moyen, J.-F. and Campion, D. (2005). An overview of adakite, tonalite-trondhjemite-granodiorite (TTG), and sanukitoid: Relationships and some implications for crustal evolution: *Lithos*, v. 79, p. 1–24, doi: 10.1016/j.lithos.2004.04.048.
- Marzoli, A., Renne, P.R., Piccirillo, E.M., Ernesto, M., Bellieni, G., De Minet, A. (1999). Extensive 200-million-year-old continental flood basalts of the Central Atlantic Magmatic Province. *Science*, 284: 616–618.
- Menand, T. (2008). The mechanics and dynamics of sills in layered elastic rocks and their implications for the growth of laccoliths and other igneous complexes. *Earth Planetary Science Letters* 267, 93–99. doi: 10.1016/j.epsl.2007.11.043.
- McDonough, W.F. and Sun, S.-S. (1995). Composition of the Earth. *Chemical Geology* 120: 223-253. doi: 10.1016/0009-2541(94)00140-4.
- Miller, J.D., Jr., Green, J.C., Severson, M.J., Chandler, V.W., Hauck, S.A., Peterson, D.M., and Wahl, T.E. (2002). Geology and mineral potential of the Duluth Complex and related rocks of northeastern Minnesota: Minnesota Geological Survey Report of Investigations 58, 207 p.

-
- Minifie, M.J., Kerr, A.C., Ernst, R.E., et al. (2013). The northern and southern sections of the western ca. 1880 Ma Circum-Superior Large Igneous Province, North America: the Pickle Crow dyke connection? *Lithos*, 174: 217–235.
- Mungall J. E. (2007). Crustal contamination of picritic magmas during transport through dikes: the Expo Intrusive Suite, Cape Smith fold belt, New Quebec. *Journal of Petrology* 48, 1021–1039.
- Mungall, J.E. (2016). U-Pb geochronology documents out-of-sequence emplacement of ultramafic layers in the Bushveld Igneous Complex of South Africa. *Nature Communication* 7:13385 doi:10.1038/ncomms13385.
- Mungall, J.E., Hanley, J.J., Arndt, N.T., and Debecdelievre, A. (2006). Evidence from meimechites and other low-degree mantle melts for redox controls on mantle-crust fractionation of platinum-group elements: *Proceedings of the National Academy of Sciences of the United States of America*, v. 103, p. 12,695–12,700.
- Mungall, J.E., Harvey, J.D, Balch S.J., Azar, B., Atkinson, J. and Hamilton, M.A. (2010). Eagle’s Nest: 2010, A Magmatic Ni-Sulfide Deposit in the James Bay Lowlands, Ontario, Canada 2010 Society of Economic Geologists, Inc. Special Publication 15, pp. 539–557
- Muraoka, H. (1997). Conceptual model for emplacement depth of magma chambers and genesis of hydrothermal systems. *Proc. International Geological Congress Vol. 9*, pp. 143-155.

- Naldrett, A.J. (2010). From the mantle to the bank: The life of a Ni-Cu-(PGE) sulfide deposit: *South African Journal of Geology*, v. 113, p. 1–32.
- Neumann, E.-R., Svensen, H., Galerne, C.Y. and Planke, S., (2011). Multistage evolution of dolerites in the Karoo Large Igneous Province, Central South Africa. *Journal of Petrology*, 52: 959–984.
- Niu, Y., Wilson, M., Humphreys, E.R., O'Hara, M. J. (2011). The origin of intra-plate ocean island basalts (OIB): the lid effect and its geodynamic implications. *Journal of Petrology*, 52 (7-8): 1443–1468. doi:10.1093/petrology/egr030.
- Olsson, J.R., Klausen, M.B., Hamilton, M.A., Marz, N., Söderlund, U. & Roberts, J.A., (2016). Baddeleyite U–Pb ages and geochemistry of the 1875–1835 Ma Black Hills Dyke Swarm across northeastern South Africa: part of a trans-Kalahari Craton back-arc setting? *GFF*. doi: 10.1080/11035897.2015.1103781
- Pearce, J. A. (1996). A user's guide to basalt discrimination diagrams. in Wyman, D.A., ed.. *Trace element geochemistry of volcanic rocks: applications for massive sulfide exploration*: Geological Association of Canada. Short Course Notes. v. 12. p. 79- 113.
- Peate, D.W. (1997). The Parana-Etendeka Province. In: Mahoney, J. J. & Coffin, M. F. (eds) *Large Igneous Provinces: Continental, Oceanic, and Planetary Flood Volcanism*. Geophysical Monograph, American Geophysical Union 100, 247-272.

- Peng, P., 2014. Precambrian mafic dyke swarms in the North China Craton and their geological implications. *Science China-Earth Sciences*, Volume 58, Issue 5, pp. 649–675.
- Peng, P., Zhai, M.-G., Zhang, H.-F., & Guo, J.-H. (2005). Geochronological constraints on the Paleoproterozoic evolution of the North China craton: SHRIMP zircon ages of different types of mafic dykes. *International Geology Review*, 47: 492–508.
- Puffer, J.H. (2001). Contrasting HFSE contents of plume sourced and reactivated arc-sourced continental flood basalts. *Geology*, 29: 675–678.
- Randall, W. (2005). U-Pb geochronology of the Expo Igneous Suite, Cape Smith Belt, and the Kyak Bay intrusion, New Quebec Orogen: implications for the tectonic evolution of the northeastern Trans- Hudson Orogen MSc thesis University of Toronto, Toronto 51pp.
- Rehkämper, M., Halliday, A.N., Alt, J.C., Fitton, J.G., Zipfel, J., and Takazawa, E. (1999): Non-chondritic platinum-group element ratios in oceanic mantle lithosphere: petrogenetic signature of melt percolation? *Earth and Planetary Science Letters*, 172(1-2), 65-81, doi:10.1016/S0012-821X(99)00193-4
- Richards, M.A., Duncan, R.A., and Courtillot, V.E. (1989). Flood basalts and hot-spot tracks – plume heads and tails. *Science*, 246: 103–107.
- Robinson J.A., and Wood B.J. (1998). The depth of the spinel to garnet transition at the peridotite solidus. *Earth and Planetary Science Letters* 164: 277–284.

-
- Sanborn-Barrie, M., St-Onge, M.R., Young, M.D., and James, D.T. (2008). Bedrock geology of south-western Baffin Island, Nunavut: expanding the tectonostratigraphic framework with relevance to mineral resources; Geological Survey of Canada, Current Research 2008-6, 16p.
- Saunders, A.D. (2005). Large igneous provinces: origin and environmental consequences. *Elements*, 1: 259–263.
- Scott, D.J. (1997). Geology, U-Pb, and Pb-Pb geochronology of the Lake Harbour area, southern Baffin Island: implications for the Paleoproterozoic tectonic evolution of northeastern Laurentia. *Canadian Journal of Earth Sciences*, 34: 140-155.
- Serrano, L., L. Ferrari, M. L. Martinez, C. M. Petrone, and C.Jaramillo (2011). An integrative geologic, geochronologic and geochemical study of Gorgona Island, Colombia: Implications for the formation of the Caribbean Large Igneous Province, *Earth Planetary Science Letters*, 309(3–4), 324–336, doi:10.1016/j.epsl.2011.07.011.
- Sheth, H. (2016). Giant plagioclase basalts: continental flood basalt-induced remobilization of anorthositic mushes in a deep crustal sill complex. *Geological Society of America Bulletin*, v. 128, pp. 916-925.
- Shirey, S.B. and Hanson, G.N. (1984). Mantle-derived Archean monzodiorites and trachyandesites. *Nature*, v310, pp. 222-224.
- Snyder, D., Carmichael, I.S.E. and Wiebe, R.A. (1993). Experimental study of liquid evolution in an Fe-rich, layered mafic intrusion: constraints of Fe-Ti

- oxide precipitation on the T-fO₂ and T-*q* paths of tholeiitic magmas. *Contributions to Mineralogy and Petrology* (1993) 113: 73. doi:10.1007/BF00320832
- Sobolev, S.V., Sobolev, A.V., Kuzmin, D.R., Krivolutskaya, N.A., Petrunin, A.G., Arndt, N.T., Radko, V.A. & Vasiliev, Y.R. (2011). Linking mantle plumes, large igneous provinces and environmental catastrophes. *Nature*, 477: 312–316.
- Söderlund, U., Hofmann, A., Klausen, M.B, et al. (2010). Towards a complete magmatic barcode for the Zimbabwe craton: baddeleyite U–Pb dating of regional dolerite dyke swarms and sill provinces. *Precambrian Research*, 183: 388–398.
- St-Onge, M.R., Lucas, S.B., Scott, D.J., and Bégin, N.J. (1987). Tectonostratigraphy and structure of the lac Watts-lac Cross-rivière, Déception area, central Cape Smith Belt, northern Quebec; in *Current Research, Part A*; Geological Survey of Canada, Paper 87-1A, p. 619-632.
- Svensen, H., Corfu, F., Polteau, S., Hammer, Ø., Planke, S., 2012. Rapid magma emplacement in the Karoo Large Igneous Province. *Earth Planet. Sci. Lett.* 325–326, 1–9. <http://dx.doi.org/10.1016/j.epsl.2012.01.015>
- Sun, S.S. and McDonough, W.F. (1989). Chemical and isotopic systematics of oceanic basalts; implications for mantle composition and processes. In: *Magmatism in the ocean basins*. Saunders, A.D. and Norry, M.J. (Editors), Geological Society of London, London. 42: 313-345.

- Thériault R.J., St-Onge M.R., Scott D.J. (2001). Nd isotopic and geochemical signature of the Paleoproterozoic Trans-Hudson Orogen, southern Baffin Island, Canada: implications for the evolution of eastern Laurentia. *Precambrian Research*, 108: 113–138.
- Thibert, F., Picard, C., and Trzciencki, W. (1989). Pétrologie des filons-couches différenciés Roméo 1 et 2 dans la partie centrale de la bande du Cap Smith; *Geoscience Canada*, v.16, p. 140-144.
- Thompson, R.N., Morrison, M.A., Dickin, A.P. and Hendry, G.L. (1983). Continental flood basalts... arachnids rule OK? In *Continental basalts and mantle xenoliths*, C.J. Hawkesworth & M.J. Norry (eds), Cheshire, U.K., Shiva Publishing Limited, p. 158-185.
- Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.) (1996). *Proceedings of the Ocean Drilling Program, Scientific Results*, Vol. 149, pp. 506.
- Winter, J.D. (2001). *An Introduction to Igneous and Metamorphic Petrology*: Prentice Hall, Upper Saddle River, New Jersey, 697p.
- Wodicka, N., Madore, L., Larbi, Y., Vicker, P. (2002). Géochronologie U-Pb de filons-couches mafiques de la Ceinture de Cape Smith et de la Fosse du Labrador. In: *L'exploration minérale au Québec: notre savoir, vos découvertes. Séminaire d'information sur la recherche géologique. Programme et résumés 2002–10*. Ministère des Ressources Naturelles, Québec. 48pp.

-
- Wodicka, N., Whalen, J. B., St-Onge, M. R., and Corrigan, D. (2010). Meta Incognita microcontinent revisited: insights from U-Pb geochronology and Nd isotopes. In Geological Association of Canada-Mineralogical Association of Canada, Abstract (Vol. 35).
- 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. doi: 10.1016/j.epsl.2004.12.005.
- Xu, Y., Chung, S.L., Jahn, B.-M., Wu, G. (2001). Petrologic and geochemical constraints on the petrogenesis of Permian-Triassic Emeishan flood basalts in southwestern China *Lithos*, 58, pp. 145-168
- Xu, Y.-G., He, B., Chung, S.-L., Menzies, M.A., and Frey, F.A. (2004). Geologic, geochemical, and geophysical consequences of plume involvement in the Emeishan flood-basalt province. *Geology*, 32: 917–920.
- Yale, L.B. and Carpenter, S.J. (1998). Large igneous provinces and giant dike swarms: proxies for supercontinent cyclicity and mantle convection. *Earth and Planetary Science Letters*, 163: 109–122.
- Zhou, M.-F., Arndt, N.T., Malpas, J., Wang, C.Y., Kennedy, A.K. (2008). Two magma series and associated ore deposit types in the Permian Emeishan Large Igneous Province, SW China *Lithos*, 103, pp. 352-368.

Figure 8.1: Chondrite-normalized REE and primitive mantle-normalize trace element diagrams illustrating the geochemical similarity between Groups C1 and D, and A2 and B. Normalization values from [Sun and McDonough \(1989\)](#).

Figure 8.2a-b: **a)** Gabbro (14SAB-L015A01) with apparent finer-grained gabbroic xenoliths, from the Killapait sill, Meta Incognita Peninsula; **b)** Plagioclase megacrysts within the ultramafic portion of the Killapait sill.

Figure 8.3: Pods of Fe-oxide within the gabbroic portion of the Killapait sill. Photo was taken at station 14SAB-L015 (Lat: 62.280; Long: -68.068).

Figure 8.4: Model #1 for the formation of the Killapait sill of Meta Incognita Peninsula, southern Baffin Island. Stages 1-5 are described within illustration and text. The distinction between solidified and non-solidified portions of the sills are not distinguished in this diagram.

Killapaik Sill model #1

Stage 1: Initial injection of Group A1 magma

Stage 2: Injection of Group A2 magma into previous sill

Stage 3: Injection of Group C2 magma into previous sill

Stage 4: Massive injection of Group C2 magma into previous sill

Stage 5: Injection of Group C1 magma into previous sill

Figure 8.5: Model #2 for the formation of the Killapait sill of Meta Incognita Peninsula, southern Baffin Island. Stages 1-6 are described within illustration and text. The distinction between solidified and non-solidified portions of the sills are not distinguished in this diagram.

Killapait Sill model #2

Stage 1: Initial injection of Group C2 magma

Stage 2: Injection of Group C1 magma, underplating previous sill

Stage 3: Second injection of Group C2 magma, underplating previous sill

Stage 4: Second injection of Group C1 magma, underplating previous sill

Stage 5: Injection of Group A1 magma, underplating previous sill

Stage 6: Group A2 magma differentiated in-situ, producing evolved magma

Figure 8.6: Metapyroxenitic xenolith within stratigraphically overlying metaperidotitic unit of the Chidliak sill. Photo was taken at station 15SAB-L077 (Lat: 64.562; Long: -66.916).

Figure 8.7: Model #1 for the formation of the Chidliak sill of northern Hall Peninsula, southern Baffin Island. Stages 1-4 are described within illustration and text.

Chidliak sill model #1

Stage 1: Initial injection of turbulent magma

Stage 2: Continual influx of magma and "Freezing" of basal pyroxenite

Stage 3: Continuous magma flow delays differentiation and variable turbidity erodes basal pyroxenite

Stage 4: Influx of magma ceases and in-situ differentiation occurs, forming peridotite and gabbro components

Figure 8.8: Model #2 for the formation of the Chidliak sill of northern Hall Peninsula, southern Baffin Island. Stages 1-4 are described within illustration and text.

Chidliak Sill model #2

Stage 1: Initial injection of Group F magma

Stage 2: Differentiation of magma into basal pyroxenite and upper gabbro

Stage 3: Additional influx of Group F magma injected in between basal pyroxenite and gabbro units, inflating sill.

Stage 4: Influx of magma ceases and in-situ differentiation occurs, forming peridotite components

Figure 8.9: Global map showing the schematic distribution of LIPs which occurred between 1845 and 1934 Ma. Modified from [Ernst \(2014\)](#). See [Table 8.1](#) for more information on each LIP. Map is configured in the Robinson projection.

Figure 8.10: Comparison of select Frobisher suite magma types with select geochemical groups from the Circum-Superior LIP.

Figure 8.11: Map-view of the relative positioning of the main lithotectonic elements in the north-eastern segment of the Trans-Hudson orogen (THO) at ca. 1900 Ma. Possible Moho-level extensions of the Superior craton are traced by the blue coloured hashed lines. Abbreviations: Am, Amer group; B, Burwell arc; Bf, Bravo formation; FFG, Flin Flon–Glennie Complex; GFtz, Great Falls tectonic zone; Ke, Ketyet group; LL, La Ronge–Lynn Lake belts; M.I., Meta Incognita micro-continent; Pe, Penrhyn group; Pi, Piling group; STZ, Snowbird Tectonic Zone; TA, Tasiuyak domain; TNB, Thompson Nickel Belt. Modified from [Corrigan et al. \(2009\)](#).

ca. 1900 Ma

Figure 8.12: Cross-section of a transect from the Superior craton through to the Meta Incognita microcontinent. Not to scale. Based on Figure 9 from [Thériault et al. \(2001\)](#).

Figure 8.13: Model of Frobisher suite formation. At stage 1, the mantle plume is ascending towards the base of the subcontinental lithospheric mantle, below the Meta Incognita microcontinent. At stage 2, the mantle plume impinges upon the base of the lithosphere and flattens. At stage 3, low-degree partial melting occurs at the base of the lithosphere, generating high-Ti magmas (Group D), which ascend through the lithosphere directly to the Lake Harbour Group, without significant fractionation or contamination. The plume head then spreads to shallower depths where higher-degree melting occurs. At stage 4, low-Ti magmas are injected into the lithosphere and ascend to the base of the crust where extensive fractionation occurs, and are later emplaced as sills within the more distal sequence of the Lake Harbour Group. Thermal erosion of metasomatized SCLM produces Group C2 magmas. Mixing between Group C2 and Groups A2 and D produce Group B and C1 magmas. Based on Figure 9 from [Thériault et al., \(2001\)](#).

Stage 1: Mantle plume ascent **Stage 2: Lithosphere impingement and plume-head flattening** **Stage 3: Sub-lithospheric channeling and partial melting of plume head** **Stage 4: Melt ascent and sill emplacement (ca. 1900 Ma)**

Figure 8.14a-c: Mineralized hand-samples from the gossan at station 14SAB-L009. Wet hand sample slab on left and close-up of it on right **a)** Mineralized ultramafic sample (14SAB-L009A02) containing Po+Cpy+Pn; **b)** Mineralized mafic sample (14SAB-L009B02) containing Po+Cpy; **c)** Mineralized metasedimentary host-rock sample (14SAB-L009B01) containing Po+Cpy.

Figure 8.15: Chalcophile element variations within the Frobisher suite, showing Pt (a), Pd (b), Ni (c), and Cu (d) variations compared to Mg#, as an index of fractionation. Cu + Ni values are in ppm, and Pt + Pd values are in ppb. Samples with Pt or Pd concentrations below detection are plotted at zero.

Figure 8.16: Log(Cu/Pd) versus Mg# diagram illustrating PGE-enrichment and PGE-depletion in Frobisher suite samples.

Table 8.1: Details of LIPs which occurred between 1845 - 1934 Ma.

Age	LIP Name	Type	Location(s)	Area (Mkm ²)	Selected references
1.87-1.85	Kalaro-Nimnyrsky		Siberia craton	A = 0.10	Gladkochub <i>et al.</i> , 2010; Ernst <i>et al.</i> , 2016a
1.88	Circum-Superior (Pan Superior)	C	Superior craton	A = 0.60	Ernst and Bell, 2010; Minifie <i>et al.</i> , 2013
1.88	River–Morel	C	Slave craton	A = 0.06	Buchan <i>et al.</i> , 2010, 2016
1.88	Mashonaland	C	Zimbabwe craton	A = 0.16	Soderlund <i>et al.</i> , 2010
1.88-1.84	Soutpansberg-Black Hills	C	Kaapvaal craton		Hanson <i>et al.</i> , 2004b; Olsson <i>et al.</i> , 2016
1.89–1.87	Uatuma	S	Amazonian	A = 1.50	Klein <i>et al.</i> , 2012
1.89	Cuddapah-Bastar	C	Bastar	A = 0.03	2008; Belica <i>et al.</i> , 2014
1.90	Hearne-Snowbird	C	cratons		Ernst and Bleeker, 2010
1.92	Hartley (Waterberg-Olifantshoek)	C	Kaapvaal craton		Hanson <i>et al.</i> , 2004a
1.93	Xuwujia	C	North China craton	A = 0.01	Peng <i>et al.</i> , 2005; Peng, 2015

Notes: "LIP" column: bold = LIPs; not bold = interpreted LIP fragments. LIP size, mainly from Ernst and Buchan, 2001, 2004 and Ernst *et al.*, 2008, and Bryan, 2007. See also Bryan and Ferrari, 20

Type: C = continental, O = oceanic and S = silicic.

Area/Volume are minimum estimates and very approximate

In some cases no estimate was provided because the inhomogeneous distribution precluded a simple calculation.

Table 8.2: Comparison of geochemistry between Frobisher Suite and other major continental and oceanic LIPs. Values with ranges similar to that of the Frobisher Suite are in bold.

Compositional ranges of some key geochemical parameters of the large igneous provinces											
Province	Bushveld	Mid-continent	Siberia	Emei Mt	Karoo	Ferrar	Paraná	Ontong Java	Kerguelen	Deccan	Frobisher
MgO wt. %	18.0–3.5	16.8–3.8	39.8–2.9	30.1–3.6	21.5–4.0	10.7–3.1	20.0–3.2	11.2–5.4	10.2–3.0	15.3–3.1	30.04–0.96
Al ₂ O ₃ wt. %	17.5–8.4	17.9–8.3	21.1–1.7	18.1–5.6	17.2–7.5	21.9–11.4	18.3–9.2	17.1–13.4	21.0–13.2	18.8–10.0	29.39–4.41
Mg#	0.82–0.44	0.76–0.36	0.85–0.305	0.85–0.34	0.80–0.31	0.73–0.31	0.79–0.33	0.65–0.32	0.69–0.38	0.73–0.32	0.77–0.23
CaO/Al ₂ O ₃	0.81–0.50	1.69–0.27	2.95–0.28	2.12–0.25	1.22–0.44	0.87–0.49	0.89–0.46	0.90–0.64	0.81–0.24	1.31–0.24	1.85–0.31
Ni ppm	720–36	876–18	2175–6	1406–4	1140–8	167–7	893–15	227–12	220–15	573–8	1985–8
Nd ppm	38.2–3.4	71.1–4.6	113–2.5	74–1.4	97.4–6.9	41.3–7.7	73–5.0	14.1–4.3	80–3.6	49–5.9	50–0.9
Yb ppm	3.3–0.72	8.1–0.8	6.2–0.4	4.0–0.7	4.9–1.2	6.2–1.4	5.2–1.1	3.8–1.6	5.7–1.4	4.7–0.76	3.49–0.07
(La/Sm) _n	4.7–1.9	4.2–1.2	6.0–1.1	5.2–0.34	3.2–0.49	5.9–1.6	3.3–0.92	2.5–0.79	5.11–0.86	7.3–0.86	6.96–0.47
(La/Yb) _n	14.1–3.0	24.2–1.5	63–1.7	30.1–0.24	33.3–1.4	5.5–2.1	9.8–1.1	1.67–0.78	15.2–0.65	16.6–1.5	35.65–0.50
εNd(i)	–4.6–(–9.8)	3.3–(–9.0)	7.9–(–12.3)	7.8–(–7.9)	9.1–(–15.9)	–1.3–(–7.0)	9.0–(–12.1)	7.5–3.7	5.5–(–8.2)	7.4–(–19.9)	6.13–(–6.64)

Appendix A - Sample Coordinates

Sample ID	Latitude	Longitude		Sample ID	Latitude	Longitude
14SAB-H006A01	62.384508	-67.867314		15SAB-D196A01	66.268850	-67.817625
14SAB-H080A01	62.747432	-67.917699		15SAB-D196A02	66.268850	-67.817625
14SAB-H080B01	62.747432	-67.917699		15SAB-D196B01	66.268850	-67.817625
14SAB-L001A01	62.918385	-69.075670		15SAB-D237A01	66.921363	-69.181112
14SAB-L001B01	62.918385	-69.075670		15SAB-F013B01	64.812520	-68.320195
14SAB-L002A01	62.977125	-68.528075		15SAB-F034B01	65.073013	-67.816722
14SAB-L002A02	62.977125	-68.528075		15SAB-L028A01	64.738937	-67.350467
14SAB-L003A01	62.885507	-68.543792		15SAB-L056B01	64.824098	-66.920294
14SAB-L003A02	62.885507	-68.543792		15SAB-L056B02	64.824098	-66.920294
14SAB-L003B01	62.885507	-68.543792		15SAB-L057B01	64.825702	-66.918307
14SAB-L004A01	62.806867	-68.092280		15SAB-L057B02	64.825702	-66.918307
14SAB-L004A02	62.806867	-68.092280		15SAB-L059A01	64.829712	-66.918999
14SAB-L005A01	62.755075	-68.109577		15SAB-L059B01	64.829712	-66.918999
14SAB-L005A02	62.755075	-68.109577		15SAB-L066B01	66.818655	-67.353518
14SAB-L006A01	63.569594	-71.307382		15SAB-L076A01	64.562237	-66.916107
14SAB-L006A02	63.569594	-71.307382		15SAB-L076A02	64.562237	-66.916107
14SAB-L006B01	63.569594	-71.307382		15SAB-L077A01	64.562167	-66.915524
14SAB-L007A01	63.474285	-71.225592		15SAB-L077B01	64.562167	-66.915524
14SAB-L007A02	63.474285	-71.225592		15SAB-L078A01	64.562375	-66.914344
14SAB-L007A03	63.474285	-71.225592		15SAB-L078A02	64.562375	-66.914344
14SAB-L008A01	63.475609	-71.442489		15SAB-L079A01	64.562303	-66.913217
14SAB-L008A02	63.475609	-71.442489		15SAB-L080A01	64.562303	-66.913217
14SAB-L009A01	63.229450	-71.558335		15SAB-L080A02	64.564432	-66.914869
14SAB-L009A02	63.229450	-71.558335		15SAB-L080A03	64.564432	-66.914869
14SAB-L009B02	63.229450	-71.558335		15SAB-L112A01	66.267307	-67.806855
14SAB-L010A01	63.238899	-71.407389		15SAB-L113A01	66.267262	-67.807957
14SAB-L010C01	63.238890	-71.407389		15SAB-L113B01	66.267262	-67.807957
14SAB-L011A01	63.428252	-70.244096		15SAB-L119A01	65.254222	-67.080111
14SAB-L011A02	63.428252	-70.244096		15SAB-M005A01	65.253933	-69.987772
14SAB-L012A01	62.281523	-68.078272		15SAB-M008A01	65.296073	-69.983567
14SAB-L012A02	62.281523	-68.078272		15SAB-M009A01	65.314663	-69.958581
14SAB-L012A03	62.281523	-68.078272		15SAB-R174C01	65.120800	-67.661760
14SAB-L013A01	62.281397	-68.076061		15SAB-R174C02	65.120800	-67.661760
14SAB-L013A02	62.281397	-68.076061				
14SAB-L014A01	62.280957	-68.073719				
14SAB-L014A02	62.280957	-68.073719				
14SAB-L015A01	62.279553	-68.068479				
14SAB-L015A02	62.279553	-68.068479				
14SAB-L016A01	62.278300	-68.066164				
14SAB-L017A02	62.274693	-68.061174				
14SAB-R005A01	62.281342	-68.078134				
14SAB-R006A01	62.279788	-68.070389				
14SAB-R006A02	62.279788	-68.070389				
14SAB-R010A01	62.281057	-68.047466				
14SAB-S130A01	62.700590	-67.967291				
14SUB-H029 B2	64.435592	-65.245080				

Appendix B - Field & Petrographic Data

Sample ID	Group	Rock Type (general)	Rock Type (specific)	Sil Type	SIL #	Metamorphic assemblage	Sulphide Assemblage	Avg. Grain size (mm)	Plagioclase	Clinopyroxene	Orthopyroxene	Olivine	Amphibole	Biotite	Spinel	Sulphides	Ilmenite	Magnetite	Chlorite	Garnet
14 SAB-H006-A01	C1	gabro	hornblende gabro	mafic	1	Hbl+Ph+Bt	Py-Cpy	0.7	40	0	0	0	55	<1%	0	5	0	<1%	0	0
14 SAB-H080-A01	A1	hornblende	olivine orthopyroxene hornblende	ultramafic	4	Hbl+Opx+Ol	Cpy	1.2	0	0	38	11	50	0	0	trace	1	0	0	0
14 SAB-H080-B01	A2	hornblende	olivine hornblende	ultramafic	4	Hbl+Ol	Py-Cpy	1.0	0	0	0	35	65	0	0	trace	0	0	0	0
14 SAB-L001-A01	A2	hornblende	olivine orthopyroxene hornblende	mafic-ultramafic	5	Hbl+Opx+Ol	Py-Cpy	0.8	0	0	16	34	70	0	0	trace	<1%	0	0	0
14 SAB-L001-B01	A2	gabronorite	hornblende pyroxene mela-gabronorite	mafic-ultramafic	6	Hbl+Ph+Cpx+Opx	Py	0.5	25	25	10	0	30	0	0	trace	<1%	0	0	0
14 SAB-L002-A01	A1	gabronorite	orthopyroxene olivine hornblende	ultramafic	6	Hbl+Opx+Ol	Py-Cpy	1.0	0	0	10	40	50	0	0	trace	<1%	0	0	0
14 SAB-L002-A02	D	peridotite	hornblende peridotite	ultramafic	6	Hbl+Ol+/-Bt	Py-Cpy	0.8	0	0	0	40	55	5	0	0	0	0	0	0
14 SAB-L003-A01	A1	peridotite	hornblende peridotite	mafic-ultramafic	7	Ol+Hbl+Spl	Py	1.0	0	0	0	60	39	0	1	trace	0	0	0	0
14 SAB-L003-A02	A1	hornblende	orthopyroxene olivine hornblende	mafic-ultramafic	7	Hbl+Opx+Ol	Py	0.7	0	0	25	25	49	0	1	trace	trace	0	0	0
14 SAB-L003-B01	A2	gabronorite	pyroxene hornblende mela-gabronorite	mafic-ultramafic	7	Hbl+Opx+Ol+/-Spl	Py-Cpy+Py	1.0	20	10	20	0	50	0	0	trace	0	0	0	0
14 SAB-L004-A01	D	pyroxenite	olivine hornblende pyroxenite	ultramafic	8	Hbl+Opx+Ol+/-Spl	Py-Cpy+Py	0.5	0	20	29	10	39	0	0	trace	1	0	0	0
14 SAB-L004-A02	D	clinopyroxenite	hornblende olivine clinopyroxenite	ultramafic	8	Cpx+Ol+Hbl+Spl	Py-Cpy	0.8	0	50	0	25	23	0	1	trace	1	0	0	0
14 SAB-L005-A01	D	hornblende	orthopyroxene olivine hornblende	mafic-ultramafic	9	Hbl+Opx+Ol	Py-Cpy+Py+Pn	0.8	0	0	8	21	70	0	0	trace	1	0	0	0
14 SAB-L005-A02	D	gabro	dinopyroxene hornblende mela-gabro	mafic-ultramafic	9	Hbl+Cpx+Pl+/-Bt	Py-Cpy	0.7	13	28	0	0	54	4	0	trace	1	0	0	0
14 SAB-L006-A01	D	gabronorite	hornblende pyroxene mela-gabronorite	mafic-ultramafic	10	Hbl+Ph+Cpx+Opx	Py-Cpy	1.0	25	20	8	0	45	0	0	trace	2	0	0	0
14 SAB-L006-A02	D	gabronorite	hornblende pyroxene mela-gabronorite	mafic-ultramafic	10	Ph+Cpx+Hbl+Opx	Py-Cpy	1.2	34	30	5	0	30	0	0	trace	1	0	0	0
14 SAB-L006-B01	D	hornblende	olivine pyroxene hornblende	mafic-ultramafic	10	Hbl+Opx+Ol	Py-Cpy	0.7	0	15	16	30	35	0	2	trace	2	0	0	0
14 SAB-L007-A01	D	hornblende	olivine orthopyroxene hornblende	ultramafic	11	Hbl+Opx+Ol	Py-Cpy	1.2	0	0	15	5	80	0	0	trace	0	0	0	0
14 SAB-L007-A02	D	hornblende	olivine orthopyroxene hornblende	ultramafic	11	Hbl+Opx+Ol	Py-Cpy	1.0	0	0	35	5	60	0	0	trace	1	0	0	0
14 SAB-L007-A03	D	hornblende	olivine orthopyroxene hornblende	ultramafic	11	Hbl+Opx+Ol	Py-Cpy	0.8	0	0	20	30	70	0	0	trace	0	0	0	0
14 SAB-L008-A01	D	pyroxenite	plagioclase-bearing hornblende pyroxenite	mafic	12	Hbl+Opx+Ol	Py-Cpy	0.5	5	20	25	0	35	15	0	0	trace	0	0	0
14 SAB-L008-A02	D	hornblende	plagioclase-bearing orthopyroxene hornblende	mafic	12	Hbl+Cpx+Pl	Py-Cpy	0.7	4	30	0	0	65	0	0	0	1	0	0	0
14 SAB-L009-A01	D	orthopyroxenite	hornblende orthopyroxenite	mafic-ultramafic	13	Opx+Hbl+Bt	Py-Cpy	1.7	0	0	52	0	30	15	0	1	2	0	0	0
14 SAB-L009-A02	D	orthopyroxenite	hornblende orthopyroxenite	mafic-ultramafic	13	Opx+Hbl+/-Spl	Py-Cpy	1.0	0	0	49	0	45	0	0	trace	5	1	0	0
14 SAB-L009-B02	D	gabro	hornblende orthopyroxene	mafic-ultramafic	13	Ph+Cpx+Spl	Py-Cpy	0.5	48	39	0	0	0	0	3	10	0	0	0	0
14 SAB-L010-A01	B	gabronorite	hornblende pyroxene gabronorite	mafic-ultramafic	14	Ph+Cpx+Hbl+Opx	Py-Cpy+Py+Pn	0.5	45	20	15	0	20	0	0	0	0	0	0	0
14 SAB-L010-A01	C1	hornblende	olivine clinopyroxene hornblende	mafic-ultramafic	14	Hbl+Cpx+Ol	Py-Cpy+Py	0.7	0	35	0	15	50	0	0	trace	0	0	0	0
14 SAB-L011-A01	A2	gabro	hornblende pyroxene mela-gabro	mafic	15	Cpx+Ph+Hbl+Opx	Py-Cpy	0.8	30	40	5	0	25	0	0	trace	0	0	0	0
14 SAB-L011-A02	C2	gabronorite	hornblende pyroxene mela-gabronorite	mafic	15	Ph+Hbl+Cpx+Opx	Py	0.8	35	30	5	0	30	0	0	trace	trace	0	0	0
14 SAB-L012-A01	A2	peridotite	pyroxene hornblende peridotite	mafic-ultramafic	16	Ol+Hbl+Opx+Ol	Py-Cpy	1.3	0	0	16	70	20	0	4	trace	0	trace	0	0
14 SAB-L012-A02	C1	gabronorite	olivine-bearing hornblende pyroxene gabronorite	mafic-ultramafic	16	Ph+Hbl+Opx+Cpx+Ol+/-Spl	Py-Cpy+Py+Pn	2.5	53	10	15	2	20	0	0	trace	0	0	0	0
14 SAB-L012-A03	C2	peridotite	hornblende peridotite	mafic-ultramafic	16	Ol+Hbl	Py-Cpy+Py+Pn	0.7	0	0	0	65	35	0	0	0	0	0	0	0
14 SAB-L013-A01	C2	gabro	hornblende olivine gabro	mafic-ultramafic	16	Ph+Hbl+Opx+Ol	Py-Cpy	0.7	43	0	0	30	25	0	1	trace	0	0	0	0
14 SAB-L013-A02	C2	gabro	hornblende olivine mela-gabro	mafic-ultramafic	16	Ol+Hbl+Opx+Ol	Py-Cpy	0.7	13	0	0	55	30	0	2	trace	0	0	0	0
14 SAB-L014-A01	C2	peridotite	orthopyroxene hornblende peridotite	mafic-ultramafic	16	Ol+Hbl+Opx+Ol	Py	0.8	0	0	5	70	20	0	5	trace	0	0	0	0
14 SAB-L014-A02	C1	hornblende	orthopyroxene-bearing hornblende	mafic-ultramafic	16	Hbl+Spl+Opx	Py-Cpy	1.5	0	0	2	0	88	0	10	0	0	0	0	0
14 SAB-L015-A01	C1	gabro	hornblende gabro	mafic-ultramafic	16	Hbl+Pl	Py-Cpy	0.5	46	0	0	0	50	0	0	1	2	0	0	0
14 SAB-L015-A02	C1	gabro	hornblende gabro	mafic-ultramafic	16	Ph+Hbl	Py-Cpy	0.8	55	0	0	0	45	0	0	1	4	0	0	0

Sample ID	Group	Rock Type (general)	Rocky Type (specific)	Sill Type	SILL #	Metamorphic assemblage	Subhedge Assemblage	Avg. Grain size (mm)	Plagioclase	Chloropyroxene	Orthopyroxene	Olivine Amphibole	Biotite	Spinel	Ilmenite	Magnetite	Chlorite	Garnet
14 SAB-1016A01	C2	gabbro	hornblende leuco-gabbro	mafic-ultramafic	16	PH-Hbl+Bt	Py	1.2	85	0	0	0	10	5	0	0	0	0
14 SAB-1017-A02	C2	gabbro	hornblende leuco-gabbro	mafic-ultramafic	16	PH-Hbl+Bt	Py-Po-Cpy	1.7	85	0	0	0	10	5	0	0	0	0
14 SAB-1017-A02	C2	gabbro	hornblende leuco-gabbro	mafic-ultramafic	16	PH-Hbl+Opx/Spl	Py	2.2	0	0	40	58	0	2	trace	0	trace	0
14 SAB-R005 A01	C1	peridotite	hornblende peridotite	mafic-ultramafic	16	PH-Hbl+ / Bt	Py	0.8	0	0	4	trace	0	trace	0	0	0	0
14 SAB-R005 A01	C1	anorthosite	anorthosite	mafic-ultramafic	16	PH-Hbl+Bt	Py	0.8	0	0	0	30	2	0	trace	0	0	0
14 SAB-R006 A02	C2	gabbro	hornblende gabbro	mafic-ultramafic	16	PH-Hbl+Bt	Py	0.8	0	0	0	50	50	0	0	0	0	0
14 SAB-R010 A01	A1	peridotite	hornblende peridotite	mafic-ultramafic	16	Hbl+Ol	Po-Cpy	0.3	0	0	0	30	2	0	trace	0	0	0
14 SAB-S130 A01	A2	hornblende	orthopyroxene olivine hornblende	ultramafic	17	Hbl+Ol+Opx	none	0.8	0	0	5	30	65	0	0	0	0	0
14 SAB-H029 B2	F	hornblende	plagioclase-bearing olivine hornblende	ultramafic	19	Hbl+Ol+Hbl	Po-Po-Cpy	1.0	10	0	40	50	0	0	0	0	0	0
15 SAB-D196A01	C2	norite	hornblende olivine mala-norite	mafic-ultramafic	21	PH-Oil-Hbl+Opx+ / Spl	Py-Cpy	1.2	35	0	15	20	0	0	0	0	0	0
15 SAB-D196A01	C2	gabbro	hornblende pyroxene mela-gabbro	mafic-ultramafic	21	PH-Cpx+Hbl+Opx	Py-Cpy	35	25	15	0	25	0	0	0	0	0	0
15 SAB-D196A01	B	gabbro	hornblende pyroxene mela-gabbro	mafic-ultramafic	21	PH-Cpx+Opx-Hbl+Ol	Po-Py+Po-Cpy	0.7	35	25	20	5	15	0	0	0	0	0
15 SAB-D237A01	E	gabbro	pyroxene hornblende mela-gabbro	mafic	22	PH-Hbl+Cpx+Opx	Py	1.3	30	20	5	0	25	0	0	0	0	0
15 SAB-F013B01	A1	hornblende	plagioclase-bearing olivine orthopyroxene hornblende	ultramafic	23	Hbl+Opx+Ol+Ph+SpH+ / Bt	Py-Po-Cpy+Pn	0.5	5	0	28	20	45	trace	2	trace	0	0
15 SAB-F034B01	A2	gabbro	gabbro	mafic	24	Cpx+PH+Bt+Hbl	Po-Cpy	0.5	40	55	0	2	4	0	trace	1	0	0
15 SAB-L028A01	A2	gabbro	hornblende pyroxene gabbro	mafic	26	PH-Cpx+Hbl+Opx	none	0.7	45	30	8	0	15	0	0	2	0	0
15 SAB-L056B01	A2	gabbro	hornblende pyroxene gabbro	mafic	27	PH-Cpx+Hbl+Opx	Po-Cpy+Pn	0.5	40	5	0	15	0	0	0	0	0	1
15 SAB-L056B01	A2	gabbro	clinopyroxene hornblende mela-gabbro	mafic	27	Hbl+PH-Cpx	Py-Po-Cpy	1.0	35	15	0	0	0	0	0	0	0	1
15 SAB-L057B01	A2	gabbro	clinopyroxene hornblende mela-gabbro	mafic	27	Hbl+PH-Cpx	Py	1.2	20	10	0	0	70	0	0	0	0	0
15 SAB-L057B01	A2	norite	orthopyroxene hornblende mela-norite	mafic	27	Hbl+PH+Opx+Bt	Py	1.0	20	0	18	0	60	2	0	0	0	0
15 SAB-L059B01	A2	norite	orthopyroxene hornblende norite	mafic	27	Hbl+PH+Opx	Py-Cpy	0.8	40	0	10	50	0	0	0	0	0	0
15 SAB-L066B01	A2	norite	orthopyroxene hornblende mela-norite	mafic	28	Hbl+PH+Opx	Py-Cpy	2.2	35	0	25	0	40	0	0	0	0	0
15 SAB-L076A01	F	hornblende	olivine-bearing orthopyroxene hornblende	mafic-ultramafic	29	Hbl+Opx+Ol+Spl	Py-Po-Cpy	2.5	0	0	18	0	1	trace	0	0	0	0
15 SAB-L076A02	F	hornblende	olivine-bearing orthopyroxene hornblende	mafic-ultramafic	29	Hbl+Opx+Ol+Spl	Po-Pn+H-Cpy	1.5	0	0	26	2	70	0	1	trace	0	0
15 SAB-L077A01	F	hornblende	orthopyroxene olivine hornblende	mafic-ultramafic	29	Hbl+Opx+Opx+Spl	Py-Cpy	0.8	0	0	10	20	65	0	5	trace	0	0
15 SAB-L077B01	F	hornblende	plagioclase-bearing hornblende	mafic-ultramafic	29	Hbl+Pl	none	3.0	10	0	0	0	90	0	0	0	0	0
15 SAB-L078A01	F	peridotite	hornblende peridotite	mafic-ultramafic	29	Hbl+Ol+Spl	Po	1.2	0	0	46	50	0	4	trace	0	0	0
15 SAB-L078A02	F	hornblende	orthopyroxene olivine hornblende	mafic-ultramafic	29	Hbl+Opx+Opx+Spl	Po-Pn+H-Cpy	0.8	0	0	10	13	75	0	0	0	0	0
15 SAB-L079A01	F	hornblende	orthopyroxene hornblende	mafic-ultramafic	29	Hbl+Opx+Spl	Po-Cpy+Pn	0.8	0	0	40	0	53	0	7	trace	0	0
15 SAB-L080A01	F	gabbro	hornblende gabbro	mafic-ultramafic	29	Hbl+Pl	Cpy+Po	1.7	40	0	0	60	0	0	0	0	0	0
15 SAB-L080A02	F	gabbro	hornblende gabbro	mafic-ultramafic	29	Hbl+Pl	none	1.2	40	0	0	60	0	0	0	0	0	0
15 SAB-L080A03	F	gabbro	hornblende gabbro	mafic-ultramafic	29	Hbl+Pl	none	1.7	40	0	0	60	0	0	0	0	0	0
15 SAB-L080A04	F	gabbro	hornblende gabbro	mafic-ultramafic	29	Hbl+Pl+Opx	Cpy+Po-Pn	1.0	45	0	10	45	0	0	0	0	0	0
15 SAB-L112A01	C2	norite	orthopyroxene hornblende norite	mafic-ultramafic	21	Hbl+PH+Opx	Po-Cpy	1.0	35	0	20	0	45	0	0	0	0	0
15 SAB-L113A01	C2	gabbro	orthopyroxene hornblende mela-norite	mafic-ultramafic	21	Hbl+PH+Opx	Po-Cpy	1.0	35	0	20	0	45	0	0	0	0	0
15 SAB-L113A01	B	hornblende	pyroxene gabbro	mafic-ultramafic	21	PH-Hbl	Py-Cpy	1.3	50	0	0	50	0	0	0	0	0	0
15 SAB-L113B01	B	hornblende	pyroxene hornblende	mafic-ultramafic	21	PH-Hbl+Cpx+Opx	Po-Cpy+Pn	0.8	0	8	8	0	84	0	0	0	0	0
15 SAB-L119A01	A1	hornblende	olivine hornblende	ultramafic	30	Hbl+Ol+Opx	Po-Pn+H-Cpy	1.2	0	0	5	15	80	0	0	0	0	0
15 SAB-M005A01	E	gabbro	clinopyroxene hornblende gabbro	mafic	31	PH-Hbl+Cpx	Po-Cpy	0.8	49	15	0	30	0	0	2	1	3	0
15 SAB-M008A01	B	gabbro	hornblende pyroxene gabbro	mafic	32	PH-Hbl+Cpx+Opx	Py-Cpy	0.3	40	18	0	24	0	0	0	0	0	0
15 SAB-M009A01	E	gabbro	hornblende pyroxene mela-gabbro	mafic	32	Hbl+PH+Cpx+Opx	none	0.8	28	20	20	0	32	0	0	0	0	0
15 SAB-R174C01	A2	gabbro	hornblende pyroxene mela-gabbro	mafic	33	Cpx+PH+Hbl+Opx	none	0.5	35	30	20	0	22	0	0	0	3	0
15 SAB-R174C02	A2	gabbro	hornblende pyroxene mela-gabbro	mafic	33	PH-Cpx+Opx+Hbl	none	0.8	35	25	25	0	20	0	0	0	0	0
15 SAB-R174C02	A2	gabbro	hornblende pyroxene mela-gabbro	mafic	34	PH-Cpx+Opx+Hbl	none	0.3	50	25	25	0	20	0	0	0	0	0
15 SAB-W046B01	A2	gabbro	hornblende pyroxene hornblende	mafic	34	PH-Cpx+Opx+Hbl	none	0.8	50	25	25	0	20	0	0	0	0	0

Appendix C - Bulk-Rock Geochemical Data

SAMPLE	Rock Type (general)	SiO ₂	Al ₂ O ₃	FeO	MgO	CaO	TiO ₂	MnO	Cr ₂ O ₃	P ₂ O ₅	Na ₂ O	K ₂ O	SrO	BaO	LOI	Total
14 SAB-H006 A01	gabbro	43.73	17.89	11.96	8.36	12.12	1.04	0.16	0.02	0.16	1.76	0.48	0.12	0.06	0.85	100.05
14 SAB-H080 A01	hornblendite	45.52	8.77	13.25	17.90	9.52	1.76	0.20	0.14	0.16	0.90	0.20	<0.01	0.02	0.21	100.05
14 SAB-H080 B01	hornblendite	44.26	6.57	10.57	26.00	6.92	0.44	0.18	0.37	0.03	0.77	0.09	<0.01	0.02	2.34	99.74
14 SAB-L001-A01	hornblendite	46.24	7.69	11.61	22.10	8.22	0.63	0.23	0.39	0.04	0.45	0.04	<0.01	0.01	1.18	100.10
14 SAB-L001-B01	gabbro	46.41	12.64	11.64	9.59	14.79	1.34	0.19	0.13	0.10	1.13	0.35	0.05	0.02	0.34	100.00
14 SAB-L002-A01	hornblendite	46.15	6.36	10.35	25.09	7.63	0.41	0.18	0.43	0.02	0.61	0.09	<0.01	0.01	1.56	100.05
14 SAB-L002-A02	peridotite	44.44	8.64	12.65	18.60	9.14	1.47	0.19	0.20	0.14	1.20	1.26	0.02	0.07	0.47	99.90
14 SAB-L003-A01	peridotite	43.71	7.02	10.62	23.62	6.90	0.47	0.18	0.37	0.02	0.74	0.09	<0.01	0.01	4.67	99.62
14 SAB-L003-A02	hornblendite	43.42	6.80	10.09	24.14	6.46	0.44	0.17	0.39	0.04	0.60	0.09	<0.01	0.01	5.55	99.32
14 SAB-L003-B01	gabbro	45.84	12.52	11.85	11.66	11.71	1.07	0.27	0.12	0.06	1.67	0.59	0.02	0.02	0.72	99.43
14 SAB-L004-A01	pyroxenite	43.53	7.83	12.59	19.33	10.18	1.57	0.20	0.20	0.14	1.11	0.20	0.02	0.01	1.92	100.25
14 SAB-L004-A02	clinopyroxenite	42.55	6.85	13.03	19.41	12.65	1.44	0.23	0.18	0.11	0.55	0.15	0.04	0.01	1.10	99.76
14 SAB-L005-A01	hornblendite	43.49	9.22	14.19	18.53	9.63	1.89	0.21	0.15	0.17	0.65	0.18	0.02	0.01	0.38	100.30
14 SAB-L005-A02	gabbro	43.90	10.28	13.11	10.43	12.66	2.30	0.22	0.10	0.25	0.99	1.58	0.04	0.08	2.43	99.83
14 SAB-L006-A01	gabbro	43.33	12.87	16.41	4.96	11.90	3.67	0.25	0.01	0.43	2.69	0.88	0.09	0.05	5.00	99.86
14 SAB-L006-A02	gabbro	43.78	12.59	15.23	4.44	13.17	3.47	0.30	0.01	0.64	2.64	0.57	0.06	0.04	1.21	99.85
14 SAB-L006-B01	hornblendite	41.70	6.10	12.44	21.99	8.82	1.40	0.18	0.23	0.12	0.24	0.05	0.02	0.01	5.04	99.74
14 SAB-L007-A01	hornblendite	44.07	9.16	12.54	19.07	8.66	1.38	0.19	0.19	0.12	1.12	0.21	0.02	0.01	1.85	100.00
14 SAB-L007-A02	hornblendite	44.65	9.37	12.29	19.42	8.16	1.64	0.19	0.19	0.15	1.09	0.29	0.02	0.02	0.73	99.58
14 SAB-L007-A03	hornblendite	46.13	8.27	12.45	20.11	7.90	1.53	0.20	0.17	0.08	1.04	0.25	0.02	0.01	0.43	99.99
14 SAB-L008-A01	pyroxenite	45.91	12.27	15.58	7.49	11.93	2.72	0.25	0.05	0.28	1.25	0.54	0.03	0.02	-0.09	99.95
14 SAB-L008-A02	hornblendite	45.29	8.47	12.59	12.01	15.42	1.67	0.17	0.16	0.13	1.25	0.77	0.03	0.02	0.58	99.97
14 SAB-L009-A01	orthopyroxenite	44.83	8.66	12.30	22.40	5.04	1.59	0.21	0.26	0.11	0.88	1.29	0.01	0.03	0.94	99.91
14 SAB-L009-A02	orthopyroxenite	39.82	7.45	12.21	19.78	11.58	1.28	0.18	0.24	0.11	0.76	0.25	0.01	0.01	4.08	99.12
14 SAB-L009-B02	gabbro	47.96	13.14	8.72	5.62	16.12	2.58	0.15	0.42	0.22	1.77	0.27	0.02	0.02	1.99	99.96
14 SAB-L010-A01	gabbro	49.31	15.14	7.86	8.88	13.88	0.56	0.16	0.01	0.09	2.14	0.34	0.03	0.03	0.41	99.72
14 SAB-L010-C01	hornblendite	42.37	7.09	12.10	17.51	13.15	1.09	0.19	0.25	0.43	0.78	0.72	0.02	0.02	2.63	99.70
14 SAB-L011-A01	gabbro	47.82	13.91	12.02	7.49	13.56	1.29	0.22	0.04	0.10	2.00	0.25	0.02	0.02	0.12	100.20
14 SAB-L011-A02	gabbro	36.31	5.49	12.87	29.37	1.69	0.08	0.20	0.07	0.02	0.24	0.03	<0.01	0.01	11.65	99.46
14 SAB-L012-A01	peridotite	46.56	13.94	12.22	7.58	13.45	1.34	0.21	0.04	0.10	2.25	0.29	0.02	0.01	0.54	99.92
14 SAB-L012-A02	gabbro	45.96	19.25	5.63	11.37	10.69	0.25	0.13	0.09	0.02	1.00	2.02	0.15	0.12	2.54	99.86
14 SAB-L012-A03	peridotite	36.62	7.51	12.40	26.53	2.74	0.10	0.20	0.12	0.02	0.29	0.06	0.02	0.02	11.65	99.66
14 SAB-L013-A01	gabbro	41.26	17.54	9.05	17.71	8.19	0.07	0.16	0.04	0.02	0.73	0.69	0.11	0.04	3.27	99.88
14 SAB-L013-A02	gabbro	38.78	13.35	10.75	20.13	6.18	0.06	0.22	0.05	0.02	0.54	0.34	0.06	0.03	7.66	99.38
14 SAB-L014-A01	peridotite	36.74	7.41	13.62	26.51	3.57	0.11	0.18	0.08	0.05	0.41	0.09	0.03	0.02	9.66	100.00
14 SAB-L014-A02	hornblendite	33.77	22.87	11.49	16.60	8.89	0.39	0.16	0.13	0.53	1.18	0.56	0.10	0.03	1.87	99.85
14 SAB-L015-A01	gabbro	40.79	19.94	12.94	7.26	13.31	1.53	0.13	0.02	0.04	1.20	0.25	0.12	0.04	0.93	99.94
14 SAB-L015-A02	gabbro	40.44	20.77	13.34	6.41	13.22	1.52	0.12	0.02	0.04	1.24	0.31	0.13	0.04	0.88	99.96
14 SAB-L016-A01	gabbro	47.20	28.12	2.54	3.46	14.09	0.14	0.05	0.05	0.02	2.06	0.78	0.17	0.08	0.76	99.81
14 SAB-L017-A02	gabbro	49.02	27.04	2.79	3.12	12.72	0.19	0.05	0.04	0.04	2.97	0.76	0.16	0.09	0.50	99.81
14 SAB-R005 A01	peridotite	41.08	9.11	10.55	23.39	6.33	0.24	0.15	0.10	0.06	0.64	0.14	0.02	0.02	6.90	99.90
14 SAB-R006 A01	anorthosite	50.07	29.39	0.92	0.96	12.44	0.05	0.02	<0.01	0.01	3.38	1.01	0.07	0.04	1.28	99.75
14 SAB-R006 A02	gabbro	45.99	25.20	5.28	5.02	14.23	0.53	0.09	0.02	0.15	1.88	0.28	0.16	0.06	0.67	100.15
14 SAB-R010 A01	peridotite	43.05	4.41	8.67	28.36	7.46	0.34	0.20	0.34	0.02	0.37	0.08	0.02	0.02	5.10	99.39
14 SAB-S130 A01	hornblendite	38.77	10.22	12.83	23.90	5.55	0.73	0.19	0.36	0.05	0.28	0.10	0.02	0.02	5.73	100.15
14 SUB-H029 B2	hornblendite	41.15	4.94	10.28	30.04	5.64	0.22	0.16	0.34	0.01	0.22	0.07	<0.01	0.02	6.70	99.80
15SAB-D196A01	norite	41.65	12.86	12.80	20.39	8.93	0.08	0.21	0.01	0.01	0.54	0.08	0.03	0.01	0.62	99.64
15SAB-D196A02	gabbro	46.21	15.68	7.33	12.13	15.59	0.14	0.14	0.06	0.01	0.65	0.08	0.03	0.01	0.41	99.29
15SAB-D196B01	gabbro	45.04	17.26	9.15	11.74	14.40	0.15	0.15	0.04	0.01	0.79	0.09	0.04	0.02	0.36	100.25
15SAB-D237A01	gabbro	46.21	14.55	10.33	9.55	13.05	0.69	0.19	0.08	0.01	1.85	0.68	0.02	0.01	1.05	99.40
15SAB-F013B01	hornblendite	43.94	12.30	14.74	18.97	6.35	0.45	0.21	<0.01	0.06	0.88	0.43	0.04	0.02	0.11	100.15
15SAB-F034B01	gabbro	48.27	13.86	13.12	7.07	12.18	1.52	0.23	0.02	0.25	1.73	0.61	0.02	0.02	-0.18	100.20
15SAB-L028A01	gabbro	47.84	13.25	14.52	5.52	12.11	2.11	0.18	0.02	0.26	2.36	0.41	0.02	0.04	-0.13	100.20
15SAB-L056B01	gabbro	45.88	13.84	12.43	7.07	11.68	1.63	0.18	0.03	0.14	1.31	0.39	0.02	0.01	3.01	99.00
15SAB-L056B02	gabbro	47.20	14.20	12.64	7.07	10.34	1.43	0.20	0.01	0.11	1.97	0.89	0.01	0.01	1.53	99.02
15SAB-L057B01	gabbro	47.31	13.79	12.62	7.08	12.39	1.39	0.21	0.02	0.12	2.18	0.28	0.02	0.01	1.19	100.00
15SAB-L057B02	norite	50.12	11.47	10.97	13.91	8.31	0.43	0.21	0.17	0.04	2.04	0.88	0.01	0.03	0.44	100.25
15SAB-L059B01	norite	47.59	15.11	13.09	7.28	10.92	1.22	0.19	0.02	0.11	2.56	0.38	0.02	0.01	0.27	100.25
15SAB-L066B01	norite	45.79	17.55	10.91	9.58	10.63	0.80	0.17	0.03	0.07	1.57	0.80	0.01	0.02	1.09	100.25
15SAB-L076A01	hornblendite	45.21	13.41	9.16	18.77	9.61	0.19	0.14	0.12	0.01	0.89	0.16	<0.01	0.01	1.59	100.30
15SAB-L076A02	hornblendite	45.49	12.71	9.12	19.31	9.30	0.18	0.14	0.13	0.01	0.86	0.15	<0.01	0.01	1.37	99.80
15SAB-L077A01	hornblendite	42.55	12.33	9.64	21.51	7.86	0.15	0.16	0.14	0.01	0.68	0.04	<0.01	0.01	4.10	100.25
15SAB-L077B01	hornblendite	45.26	14.60	7.88	14.90	11.93	0.38	0.14	0.17	0.02	1.00	0.25	0.02	0.01	1.21	98.65
15SAB-L078A01	peridotite	39.97	8.56	9.60	25.43	5.31	0.17	0.09	0.20	0.01	0.51	0.10	<0.01	<0.01	8.95	99.98
15SAB-L078A02	hornblendite	41.50	10.95	9.56	23.80	6.53	0.17	0.14	0.20	0.01	0.57	0.10	<0.01	0.01	5.10	99.70
15SAB-L079A01	hornblendite	42.93	12.62	10.45	23.68	7.55	0.19	0.16	0.13	0.02	0.78	0.06	<0.01	0.01	0.64	100.35
15SAB-L080A01	gabbro	48.76	16.56	6.47	10.80	14.38	0.33	0.13	0.12	0.02	1.07	0.10	0.01	0.01	0.73	100.20
15SAB-L080A02	gabbro	48.23	18.02	6.47	9.92	13.31	0.34	0.13	0.13	0.01	1.36	0.10	0.01	0.01	0.87	99.63
15SAB-L080A03	norite	49.01	16.87	6.01	9.97	14.49	0.26	0.13	0.13	0.01	1.31	0.11	0.01	0.01	0.77	99.77
15SAB-L112A01	norite	43.99	21.54	6.48	10.26	12.82	0.06	0.11	0.03	0.01	1.42	0.43	0.05	0.03	0.87	98.82
15SAB-L113A01	gabbro	42.18	21.28	5.07	10.09	13.78	0.03	0.09	0.05	0.00	1.43	1.55	0.02	0.05	3.56	99.75
15SAB-L113B01	hornblendite	45.00	12.89	8.30	15.92	12.14	0.11	0.14	0.10	0.00						

SAMPLE	Ba	Ce	Cr	Cs	Dy	Er	Eu	Ga	Gd	Hf	Ho	La	Lu	Nb	Nd	Pr	Rb	Sm	Sn
14 SAB-H006 A01	418	38.1	70		1.7	0.97	1.25	19.7	2.64	1.2	0.34	17.9	0.12	0.9	19.9	4.71	3.5	3.79	1
14 SAB-H080 A01	5.4	33	930	0.37	3.17	1.62	1.24	16.6	3.88	3.5	0.59	13.8	0.18	15.4	19.5	4.33	2.6	4.29	1
14 SAB-H080 B01	6.1	3.9	2550		1.4	0.93	0.38	8.2	1.33	1.2	0.29	1.4	0.12	1	3.1	0.56	1	1.06	<1
14 SAB-L001-A01	3.1	9.1	2770	0.01	1.92	1.1	0.54	10.8	1.89	2.2	0.4	4	0.14	2	6.4	1.26	0.4	1.63	1
14 SAB-L001-B01	61.3	15.9	940		3.63	2.08	1.06	15.6	3.71	3.3	0.79	6.6	0.29	6.6	11.1	2.29	3	3.31	1
14 SAB-L002-A01	1.7	4.6	3020	0.02	1.39	0.83	0.31	7.5	1.21	1.3	0.29	1.8	0.13	1.3	3.5	0.67	1	1.08	<1
14 SAB-L002-A02	546	23.8	1400	2.99	3.11	1.48	1.2	14.1	3.63	2.8	0.57	9.5	0.17	10.4	15.5	3.34	38.6	3.85	1
14 SAB-L003-A01	7.1	4.6	2710	0.05	1.49	0.81	0.39	8.2	1.39	1	0.32	1.7	0.14	1.5	3.8	0.71	1.8	1.26	<1
14 SAB-L003-A02	26.4	4.6	2820	0.16	1.39	0.81	0.38	7.5	1.3	0.9	0.29	1.7	0.13	1.5	3.5	0.67	2.1	1.05	<1
14 SAB-L003-B01	61.3	14.4	830	0.17	3.13	1.95	0.87	13.1	2.87	2.1	0.67	6.6	0.31	3.2	8.3	1.75	5.5	2.29	2
14 SAB-L004-A01	11.1	26.7	1440	0.03	2.56	1.32	1.15	13.6	3.31	3.1	0.55	11.7	0.15	12.7	16.4	3.48	2.3	3.61	1
14 SAB-L004-A02	26.8	25.4	1310	0.04	2.57	1.44	1.31	13.3	3.37	2.7	0.54	11	0.19	11.2	15.1	3.42	2	3.57	1
14 SAB-L005-A01	20.4	31.7	1080	0.03	3.2	1.48	1.32	16.8	3.83	2.9	0.61	13.4	0.16	15.1	19.9	4.36	1.4	4.54	1
14 SAB-L005-A02	691	39.3	750	0.91	3.08	1.29	1.44	22.1	4.43	3.4	0.59	16.7	0.13	22.4	24.1	5.33	81.7	5.45	1
14 SAB-L006-A01	362	85.2	20	0.02	6.97	3.08	3.31	25.3	9.02	6.3	1.27	36.1	0.37	45.6	50	11.05	12.5	10.7	3
14 SAB-L006-A02	291	82.8	20		6.79	3.22	3.26	23	8.61	6.9	1.25	36.5	0.38	47.2	48.2	10.7	5.4	10.1	3
14 SAB-L006-B01	3.4	27.6	1620	0.01	2.35	1.11	0.92	11.6	3.07	2.5	0.45	11.6	0.14	13.7	16.7	3.56	0.5	3.67	1
14 SAB-L007-A01	26.6	22.7	1340	0.4	2.97	1.54	1.18	14.1	3.28	2.4	0.56	9.6	0.16	9.9	14.3	2.98	5	3.41	1
14 SAB-L007-A02	93.9	25.2	1410	0.28	3.41	1.7	1.28	14.6	4.34	3.1	0.67	10.1	0.21	11.1	17.3	3.64	6.9	4.45	1
14 SAB-L007-A03	27.5	14.3	1180	0.1	2.84	1.57	0.79	12.4	3.08	2.4	0.59	5.8	0.18	8.1	10.2	2.04	5.3	2.73	1
14 SAB-L008-A01	112.5	54.2	370	0.34	5.55	2.66	2.12	23.4	7.08	5.1	1.07	21.5	0.32	28	36	7.5	17	8.07	2
14 SAB-L008-A02	123	36.2	1190	1.06	3.47	1.68	1.37	14.9	4	3.4	0.64	15.7	0.19	15.1	22.4	4.84	26.9	5.16	2
14 SAB-L009-A01	176	20	1930	166	2.67	1.27	0.79	13.3	3.23	2.5	0.51	8.2	0.16	12.3	13.2	2.76	374	3.21	4
14 SAB-L009-A02	49	18.6	1760	1.15	2.66	1.4	1.04	15.6	2.95	2.2	0.54	8.3	0.16	11.1	12.4	2.52	6.4	3.01	15
14 SAB-L009-B02	33.4	51.5	3010	0.51	5.13	2.42	2.56	18.5	6.14	5.1	0.98	23.6	0.26	20.9	29.5	6.56	5	6.68	4
14 SAB-L010-A01	154	15.7	130	0.04	2.24	1.36	0.7	13.8	2.12	2.3	0.49	7	0.21	1.5	9.5	2.05	2.2	2.2	1
14 SAB-L010-C01	84.3	34.8	1710	0.19	2.3	1.12	0.75	11.2	2.71	2	0.45	13.9	0.13	9.4	17	4.2	20.9	3.46	22
14 SAB-L011-A01	32.9	13.2	290	0.02	3.5	2.14	0.96	17	3.19	2.7	0.77	5	0.3	4.5	9.9	1.92	1.6	2.89	2
14 SAB-L011-A02	13	3.1	560	0.03	0.15	0.07	0.13	3.8	0.13	0.2	0.03	1.5	0.01	0.4	1.4	0.34	0.5	0.19	<1
14 SAB-L012-A01	36.1	12.8	310	0.02	3.58	2.27	1.06	17.5	3.32	2.3	0.81	4.5	0.32	4.6	10.1	1.94	2.3	2.99	1
14 SAB-L012-A02	973	11.3	690	0.18	1.15	0.51	0.68	14.6	1.37	0.6	0.22	5	0.05	0.3	7.9	1.55	23.5	1.71	<1
14 SAB-L012-A03	68.2	5.9	950	0.05	0.24	0.16	0.25	5.5	0.42	0.3	0.06	2.9	0.02	0.4	3	0.76	2.7	0.49	<1
14 SAB-L013-A01	247	5.1	350	0.14	0.15	0.1	0.33	10.6	0.25	0.5	0.03	2.8	0.01	0.3	2.5	0.53	8	0.35	<1
14 SAB-L013-A02	229	5.2	410	0.17	0.17	0.1	0.25	8.6	0.29	0.4	0.04	2.8	0.02	0.4	2.4	0.62	5.9	0.43	<1
14 SAB-L014-A01	87.9	6.7	660	0.09	0.28	0.14	0.21	5.6	0.4	0.4	0.06	3.2	0.02	0.6	3.1	0.78	2	0.57	<1
14 SAB-L014-A02	166.5	57	910	0.2	2.09	1.24	0.98	28.6	3.5	3.3	0.41	27.2	0.12	3.2	29.4	6.93	5.6	4.49	1
14 SAB-L015-A01	299	14.9	130	0.03	1.14	0.62	0.93	29.1	1.61	1.2	0.2	7.2	0.08	1	9.2	2	3.6	2.16	<1
14 SAB-L015-A02	239	12	40	0.02	0.73	0.42	0.63	23.4	1.17	0.8	0.15	6	0.04	0.8	7	1.55	2.9	1.34	<1
14 SAB-L016-A01	581	29.5	320	0.53	0.78	0.49	0.87	18	1.18	1.4	0.15	16.9	0.06	3.5	11.2	3.13	19.7	1.82	1
14 SAB-L017-A02	674	32.8	270	0.45	0.72	0.49	0.86	19.3	1	1.9	0.14	19.2	0.06	2.7	11.9	3.44	19.6	1.78	1
14 SAB-R005 A01	41.9	11.7	680	0.03	0.85	0.47	0.5	7.7	1.15	1.2	0.15	5.5	0.06	1.6	6.5	1.52	1.6	1.46	<1
14 SAB-R006 A01	241	6.4	20	0.21	0.23	0.18	0.36	18.1	0.3	0.9	0.06	3.4	0.01	1.1	2.6	0.69	15.1	0.58	<1
14 SAB-R006 A02	400	29.3	130	0.04	1.16	0.61	1.04	18.2	1.88	1.1	0.22	14.8	0.09	1.4	14.9	3.61	2.2	2.63	<1
14 SAB-R010 A01	4.1	3.4	2340		1.15	0.75	0.37	5.5	1.05	0.9	0.26	1.4	0.07	1	2.7	0.53	1	0.97	1
14 SAB-S130 A01	10.6	8.4	2450		1.48	0.92	0.51	10.7	1.65	1.7	0.31	3.4	0.16	3.2	5.9	1.21	0.7	1.47	<1
14 SUB-H029 B2	9.3	1.6	2250	0.24	0.73	0.42	0.19	5.3	0.63	0.4	0.17	0.6	0.07	0.7	1.4	0.23	0.7	0.51	<1
15SAB-D196A01	75.9	4.7	120	0.01	0.26	0.17	0.25	8.6	0.34	0.3	0.06	2.3	0.02	0.2	2.4	0.6	1.6	0.41	<1
15SAB-D196A02	53.9	4.7	460	0.06	0.45	0.37	0.28	9.2	0.57	0.3	0.12	1.9	0.04	0.141	2.7	0.64	1.5	0.62	<1
15SAB-D196B01	85.2	5	200	0.04	0.59	0.26	0.34	12.4	0.56	0.3	0.11	2.2	0.04	0.141	3	0.67	2.6	0.6	<1
15SAB-D237A01	56.8	12	580	0.07	5.62	3.64	0.65	15.5	4.57	1	1.27	4.2	0.47	3.6	9.5	1.9	5.9	3.4	1
15SAB-F013B01	161	18.9	50	0.31	1.6	0.82	0.66	10.5	2.01	1	0.31	8.5	0.08	2.3	10.5	2.51	10.6	2.12	<1
15SAB-F034B01	178	33.4	150	0.03	5.33	3.33	1.41	18.6	5.03	2.8	1.07	13.7	0.45	9.9	20.2	4.44	4.6	4.9	1
15SAB-L028A01	342	30.1	100	0.03	5.81	3.8	1.6	19.2	5.37	3.5	1.31	12.1	0.55	17.3	17.9	4.16	1.8	4.93	<1
15SAB-L056B01	50.5	22.7	240	0.35	4.85	2.66	1.27	19.6	4.53	2.6	1.03	9.2	0.41	9.1	14.4	3.11	5.7	4.02	1
15SAB-L056B02	58.5	25.2	100	3.8	4.36	2.7	1.12	20.6	4.23	2.5	0.91	9.9	0.35	9.5	14.1	3.27	38.8	4.12	6
15SAB-L057B01	40.9	18.3	160	0.1	4.18	2.67	1.12	18.2	3.98	2.1	0.89	6.9	0.36	8	12.5	2.64	2.3	3.73	1
15SAB-L057B02	207	13.3	1220	0.19	2.76	1.82	0.6	13.3	2.48	1.2	0.61	5.5	0.26	2	7.9	1.81	19.3	2.07	1
15SAB-L059B01	38.2	16.5	160	0.04	3.91	2.67	1.14	18	3.84	2.2	0.9	6.3	0.34	6.7	11.1	2.38	2.6	3.27	1
15SAB-L066B01	120	13.9	240	0.14	2.49	1.65	0.72	17.6	2.34	1.4	0.55	5.8	0.22	2.9	8.4	1.89	23.6	2.14	4
15SAB-L076A01	15.2	1.5	870	0.03	0.75	0.52	0.23	9.5	0.62	0.2	0.18	0.6	0.08	0.3	1.2	0.23	3.9	0.45	<1
15SAB-L076A02	13.4	1.4	980	0.04	0.79	0.5	0.2	9	0.58	0.3	0.17	0.5	0.09	0.3	1	0.21	3.9	0.3	<1
15SAB-L077A01	6.4	1.1	1040	0.03	0.57	0.44	0.14	8.6	0.42	0.2	0.13	0.25	0.07	0.2	0.9	0.17	0.9	0.34	<1
15SAB-L077B01	39.8	2.4	1260	1.37	1.7	1.05	0.26	11.6	1.26	0.5	0.37	0.8	0.15	0.6	2.4	0.41	8.8	0.89	<1
15SAB-L078A01	9.1	1.6	1540	0.1	0.73	0.58	0.14	6.7	0.63	0.3	0.17	0.7	0.07	0.3	1.1	0.24	0.9	0.42	<1
15SAB-L078A02	10	1.7	1480	0.01	0.54	0.38	0.12	7.8	0.44	0.3	0.13	0.7	0.04	0.4	1	0.23	1.1	0.39	<1
15SAB-L079A01	9.3	1.7	990	0.01	0.79	0.5	0.17	9.1	0.66	0.3	0.19	0.7	0.09	0.4	1.5	0.25	1.2	0.43	<1
15SAB-L080A01	9.7	2.1	890	0.03	1.34	0.84	0.26	13	0.99	0.4	0.32	0.7	0.13	0.4	1.8	0.32	0.9	0.59	<1
15SAB-L080A02	13.7</																		

SAMPLE	Sr	Ta	Tb	Th	Tm	U	V	W	Y	Yb	Zr	Ag	As	Cd	Co	Cu	Li	Mo	Ni	Pb	Sc	Ti	
14 SAB-H006 A01	989		0.33	0.35	0.11	0.05	374	<1	8.4	0.78	50	<0.5	<5	<0.5	58	131	10	<1	41	5	28	<10	
14 SAB-H080 A01	44.2	0.8	0.57	1.21	0.19	0.3	250	2	14.7	1.23	145	<0.5	<5	0.9	88	27	10	<1	959	<2	22	<10	
14 SAB-H080 B01	47.8	0.1	0.24	0.13	0.12	0.06	157	<1	7.7	0.79	48	<0.5	<5	0.6	99	76	10	<1	1440	<2	24	<10	
14 SAB-L001-A01	12.6	0.1	0.29	0.91	0.14	0.24	184	<1	10.1	1.03	88	<0.5	<5	0.5	79	72	<10	<1	928	2	23	<10	
14 SAB-L001-B01	443	0.4	0.58	0.78	0.31	0.7	344	<1	19.1	1.89	125	<0.5	<5	<0.5	64	20	10	<1	447	4	38	<10	
14 SAB-L002-A01	34.7	0.1	0.23	0.18	0.12	0.06	151	<1	7.4	0.76	51	<0.5	<5	<0.5	89	20	10	<1	1330	<2	23	<10	
14 SAB-L002-A02	153	0.6	0.52	0.82	0.19	0.24	242	<1	14.1	1.24	104	<0.5	<5	0.7	89	97	10	<1	902	<2	24	<10	
14 SAB-L003-A01	56.9	0.1	0.24	0.22	0.13	0.08	175	<1	8	0.81	35	<0.5	<5	0.5	89	75	10	<1	1240	<2	27	<10	
14 SAB-L003-A02	59.5	0.1	0.21	0.21	0.11	0.07	167	<1	7.6	0.76	28	<0.5	<5	<0.5	82	57	10	<1	1210	<2	25	<10	
14 SAB-L003-B01	216	0.2	0.49	0.06	0.28		358	<1	17.1	1.88	67	<0.5	<5	0.6	45	45	20	<1	124	3	47	<10	
14 SAB-L004-A01	221	0.8	0.48	1.4	0.17	0.28	239	<1	12.6	1.09	113	<0.5	<5	0.6	86	59	10	<1	938	<2	23	<10	
14 SAB-L004-A02	381	0.7	0.48	1.11	0.18	0.31	209	<1	12.9	1.04	102	<0.5	<5	0.6	81	10	20	<1	909	<2	19	<10	
14 SAB-L005-A01	154.5	0.9	0.59	1.2	0.19	0.23	275	<1	14.3	1.23	102	<0.5	<5	0.7	90	118	20	<1	871	11	23	10	
14 SAB-L005-A02	325	1.1	0.58	1.3	0.16	0.37	265	<1	13.3	0.98	134	<0.5	<5	1	64	17	10	<1	499	5	20	<10	
14 SAB-L006-A01	752	2.8	1.16	3.09	0.42	0.64	409	<1	30.8	2.57	225	<0.5	<5	0.5	62	285	10				2	26	10
14 SAB-L006-A02	558	2.9	1.2	2.87	0.41	0.65	350	<1	30.3	2.57	266	<0.5	<5	0.6	51	231	10	1	48	6	24	<10	
14 SAB-L006-B01	230	0.9	0.45	1.08	0.16	0.24	203	<1	10.7	0.93	93	<0.5	306	<0.5	80	15	10	<1	1150	2	21	<10	
14 SAB-L007-A01	173.5	0.6	0.49	0.81	0.2	0.21	217	<1	12.9	1.05	81	<0.5	<5	0.5	87	112	50	<1	942	<2	22	<10	
14 SAB-L007-A02	165.5	0.7	0.61	0.81	0.22	0.19	234	<1	16	1.47	114	<0.5	<5	0.5	77	107	60	<1	837	<2	22	<10	
14 SAB-L007-A03	167.5	0.5	0.48	0.48	0.19	0.09	195	<1	13.7	1.26	81	<0.5	<5	0.6	86	24	50	<1	905	<2	24	<10	
14 SAB-L008-A01	257	1.7	1.02	1.05	0.37	0.49	436	<1	24.7	2	191	<0.5	<5	0.5	68	117	30	<1	173	4	33	10	
14 SAB-L008-A02	308	0.9	0.59	1.64	0.24	0.57	273	<1	15.5	1.38	124	<0.5	<5	<0.5	77	51	20	<1	709	<2	27	<10	
14 SAB-L009-A01	77.7	0.8	0.47	0.87	0.18	0.21	257	<1	12.7	1.11	82	<0.5	591	<0.5	86	15	90	<1	946	2	26	10	
14 SAB-L009-A02	142	0.7	0.44	0.9	0.18	0.24	212	<1	13.1	1.05	75	<0.5	61	<0.5	98	213	10	<1	976	<2	25	<10	
14 SAB-L009-B02	201	1.4	0.92	1.85	0.32	2.68	413	1	23.1	1.9	196	<0.5	27	0.9	60	138	30	1	442	4	46	10	
14 SAB-L010-A01	277	0.1	0.35	0.33	0.2	0.16	213	<1	11.7	1.32	89	<0.5	<5	<0.5	40	93	30	<1	63	<2	48	<10	
14 SAB-L010-C01	189.5	0.9	0.41	2.5	0.15	1.52	199	1	10.7	0.94	74	<0.5	565	0.7	71	14	10	<1	792	3	23	<10	
14 SAB-L011-A01	199	0.3	0.58	0.38	0.31	0.28	332	<1	19.3	1.82	93	<0.5	<5	0.5	52	83	10	<1	128	2	42	<10	
14 SAB-L011-A02	51.6		0.02	0.3	0.02	0.08	23	<1	0.7	0.09	7	<0.5	<5	<0.5	126	34	10	<1	777	<2	6	<10	
14 SAB-L012-A01	187.5	0.3	0.59	0.1	0.32	0.09	353	<1	20.2	2.21	78	<0.5	<5	<0.5	54	134	20	<1	128	2	44	10	
14 SAB-L012-A02	1275		0.16	0.16	0.07		99	<1	4.9	0.42	17	<0.5	<5	<0.5	45	28	10	<1	184	<2	27	<10	
14 SAB-L012-A03	187		0.04	0.2	0.02		41	<1	1.3	0.14	11	<0.5	<5	<0.5	106	6	20	<1	585	<2	9	<10	
14 SAB-L013-A01	927		0.04	0.17	0.01	0.06	17	<1	0.8	0.07	18	<0.5	<5	<0.5	73	10	30	<1	365	<2	3	<10	
14 SAB-L013-A02	592		0.04	0.26	0.02	0.05	24	<1	0.9	0.11	15	<0.5	<5	0.5	82	6	30	<1	412	<2	5	<10	
14 SAB-L014-A01	303		0.06	0.41	0.02	0.09	42	<1	1.4	0.17	14	<0.5	<5	<0.5	109	8	<10	<1	535	<2	7	<10	
14 SAB-L014-A02	846	0.1	0.4	2.2	0.14	0.78	70	1	10.8	0.96	146	<0.5	<5	0.7	80	1	30	<1	238	<2	9	<10	
14 SAB-L015-A01	1370	0.1	0.22	0.23	0.08	0.07	894	1	5.4	0.47	52	<0.5	<5	0.5	59	186	10	<1	168	2	26	<10	
14 SAB-L015-A02	1120		0.15	0.16	0.06		705	<1	3.7	0.3	30	<0.5	<5	0.5	59	222	20	<1	195	<2	22	<10	
14 SAB-L016-A01	1435	0.2	0.16	0.43	0.06	0.14	27	<1	4.3	0.34	60	<0.5	<5	<0.5	15	4	20	<1	66	5	3	<10	
14 SAB-L017-A02	1410	0.1	0.13	0.72	0.06	0.23	32	<1	3.9	0.41	88	<0.5	<5	<0.5	14	42	20	<1	56	10	3	<10	
14 SAB-R005 A01	195.5	0.1	0.15	0.52	0.06	0.11	84	2	4.1	0.41	48	<0.5	<5	0.5	91	54	10	<1	436	3	25	<10	
14 SAB-R006 A01	656	0.1	0.05	0.15	0.01	0.12	8	<1	1.3	0.12	39	<0.5	<5	<0.5	4	3	20	<1	17	9	1	<10	
14 SAB-R006 A02	1330	0.1	0.24	0.17	0.08		119	<1	5.9	0.48	42	<0.5	<5	<0.5	23	32	10	<1	63	2	13	<10	
14 SAB-R010 A01	221	0.1	0.2	0.11	0.08	0.05	100	1	6.1	0.59	34	<0.5	243	0.6	99	29	10	<1	1985	<2	21	<10	
14 SAB-S130 A01	148.5	0.2	0.28	0.35	0.13	0.1	250	<1	8.3	0.94	66	<0.5	17	<0.5	83	5	10	<1	974	2	40	<10	
14 SUB-H029 B2	48.7	0.7	0.12	0.05	0.09		76	<1	4.1	0.44	17	<0.5	7	0.5	105	20	<10	<1	1980	<2	19	<10	
15SAB-D196A01	262		0.05	0.09	0.02		63	<1	1.5	0.17	11	<0.5	<5	<0.5	105	48	<10	<1	155	5	17	<10	
15SAB-D196A02	266		0.1	0.89	0.06	0.06	140	<1	2.7	0.29	6	<0.5	<5	<0.5	54	83	<10	<1	87	5	47	<10	
15SAB-D196B01	334		0.09	0.1	0.05		155	<1	2.6	0.3	9	<0.5	<5	<0.5	64	189	<10	<1	86	11	34	<10	
15SAB-D237A01	195.5	0.1	0.89	0.28	0.51	0.05	299	<1	31.9	3.23	22	<0.5	<5	<0.5	50	138	30	<1	98	4	50	<10	
15SAB-F013B01	383		0.27	1.15	0.12	0.29	74	1	7.4	0.68	32	<0.5	<5	<0.5	93	84	<10	<1	185	4	13	<10	
15SAB-F034B01	206	0.4	0.87	0.67	0.48	0.31	386	1	28	3.11	98	<0.5	<5	<0.5	49	28	20	1	55	6	47	<10	
15SAB-L028A01	205	0.9	0.99	0.3	0.53	0.25	473	<1	33	3.49	119	<0.5	<5	<0.5	52	23	20	1	42	2	46	<10	
15SAB-L056B01	185.5	0.4	0.79	0.23	0.42	0.08	411	<1	25.1	2.6	93	<0.5	<5	<0.5	52	20	30	<1	111	3	43	<10	
15SAB-L056B02	138.5	0.5	0.76	0.45	0.35	0.51	382	1	22.6	2.38	75	<0.5	<5	0.5	52	23	10	1	78	7	36	<10	
15SAB-L057B01	170	0.4	0.69	0.18	0.39	0.11	384	<1	22.9	2.18	67	<0.5	<5	<0.5	56	27	20	1	88	5	41	<10	
15SAB-L057B02	71.4	0.1	0.44	0.39	0.25	0.14	164	<1	15.2	1.73	37	<0.5	<5	<0.5	68	2	10	<1	390	5	28	<10	
15SAB-L059B01	168.5	0.2	0.7	0.13	0.37	0.07	357	<1	21.9	2.24	72	<0.5	<5	<0.5	53	22	10	<1	100	4	43	<10	
15SAB-L066B01	137	0.1	0.45	0.1	0.25	0.07	181	1	14.2	1.59	43	<0.5	<5	<0.5	60	41	<10	<1	191	6	20	<10	
15SAB-L076A01	45.2	0.3	0.12	0.05	0.08		100	1	4.4	0.49	9	<0.5	<5	<0.5	66	42	10	<1	697	2	20	<10	
15SAB-L076A02	42.8	0.3	0.12	0.06	0.07		101	<1	4.2	0.44	9	<0.5	<5	<0.5	62	53	10	<1	675	3	19	<10	
15SAB-L077A01	47.5	0.2	0.11			0.06	77	<1	3.3	0.36													

SAMPLE	Zn	C	S	Au	Pt	Pd	Mg#	Eu/Eu*	(La/Sm)N	(Gd/Yb)N	(La/Yb)N	TOTAL REE
14 SAB-H006 A01	67	0.05	0.3	0.001	<0.005	0.001	41.15	1.15	3.05	2.80	16.46	92.64
14 SAB-H080 A01	118	0.01	<0.01	0.001	<0.005	0.001	57.46	0.91	2.08	2.61	8.05	87.59
14 SAB-H080 B01	69	0.06	0.05	0.001	0.007	0.007	71.09	0.98	0.85	1.39	1.27	15.62
14 SAB-L001-A01	34	0.06	<0.01	0.006	0.008	0.008	65.56	0.94	1.58	1.52	2.79	29.84
14 SAB-L001-B01	89	0.1	0.01	0.001	0.011	0.009	45.16	0.92	1.29	1.62	2.50	53.54
14 SAB-L002-A01	43	0.02	0.01	0.001	0.006	0.006	70.80	0.83	1.08	1.32	1.70	16.92
14 SAB-L002-A02	101	0.01	<0.01	0.004	<0.005	0.004	59.52	0.97	1.59	2.42	5.50	68.1
14 SAB-L003-A01	46	0.07	0.02	<0.001	0.009	0.005	68.99	0.90	0.87	1.42	1.51	17.79
14 SAB-L003-A02	32	0.02	0.02	0.001	0.008	0.008	70.53	0.99	1.05	1.42	1.60	16.9
14 SAB-L003-B01	101	0.01	0.1	0.002	0.015	0.012	49.60	1.04	1.86	1.26	2.52	45.79
14 SAB-L004-A01	45	0.53	0.03	0.001	<0.005	0.003	60.56	1.00	2.09	2.51	7.70	72.67
14 SAB-L004-A02	32	0.26	0.01	<0.001	<0.005	<0.001	59.84	1.14	1.99	2.68	7.59	69.61
14 SAB-L005-A01	81	0.07	0.02	0.003	<0.005	0.001	56.63	0.94	1.91	2.58	7.81	86.51
14 SAB-L005-A02	203	0.59	0.02	0.001	<0.005	<0.001	44.31	0.87	1.98	3.74	12.22	103.56
14 SAB-L006-A01	147	0.11	0.03	0.001	<0.005	<0.001	23.21	1.00	2.18	2.90	10.08	221.22
14 SAB-L006-A02	130	0.35	0.11	0.001	<0.005	0.002	22.57	1.04	2.33	2.77	10.19	215.99
14 SAB-L006-B01	25	0.54	0.01	0.008	<0.005	0.004	63.86	0.81	2.04	2.73	8.95	72.71
14 SAB-L007-A01	99	0.33	0.07	<0.001	<0.005	0.003	60.32	1.06	1.82	2.58	6.56	64.42
14 SAB-L007-A02	93	0.01	0.01	0.004	<0.005	0.01	61.24	0.88	1.47	2.44	4.93	74.6
14 SAB-L007-A03	102	0.02	<0.01	0.001	<0.005	0.001	61.76	0.83	1.37	2.02	3.30	46.05
14 SAB-L008-A01	133	0.03	0.09	0.003	0.009	0.008	32.47	0.84	1.72	2.93	7.71	149.46
14 SAB-L008-A02	96	0.1	0.01	0.003	0.006	0.004	48.82	0.89	1.96	2.40	8.16	97.86
14 SAB-L009-A01	73	0.02	0.05	0.017	0.006	0.008	64.55	0.74	1.65	2.41	5.30	57.76
14 SAB-L009-A02	46	0.08	1.4	0.003	0.005	0.007	61.83	1.05	1.78	2.32	5.67	55.25
14 SAB-L009-B02	223	0.13	3.27	0.005	0.006	0.005	39.19	1.20	2.28	2.67	8.91	138.47
14 SAB-L010-A01	61	0.07	0.1	<0.001	<0.005	<0.001	53.03	0.98	2.05	1.33	3.80	45.44
14 SAB-L010-C01	197	0.09	0.14	0.005	0.008	0.005	59.13	0.72	2.59	2.38	10.61	82.32
14 SAB-L011-A01	90	0.01	0.1	0.001	<0.005	0.002	38.39	0.96	1.12	1.45	1.97	46.48
14 SAB-L011-A02	97	0.06	0.03	0.001	0.025	0.051	69.54	2.39	5.10	1.19	11.95	7.18
14 SAB-L012-A01	91	0.01	0.09	0.001	<0.005	0.001	38.28	1.02	0.97	1.24	1.46	46.81
14 SAB-L012-A02	122	0.07	0.02	0.001	<0.005	0.001	66.87	1.31	1.89	2.70	8.54	32.09
14 SAB-L012-A03	38	0.11	<0.01	<0.001	<0.005	0.001	68.15	1.64	3.82	2.48	14.86	14.4
14 SAB-L013-A01	42	0.15	0.02	<0.001	<0.005	0.001	66.18	3.25	5.16	2.95	28.69	12.27
14 SAB-L013-A02	35	0.19	0.01	0.001	<0.005	<0.001	65.18	2.04	4.20	2.18	18.26	12.49
14 SAB-L014-A01	53	0.11	0.02	<0.001	<0.005	<0.001	66.06	1.28	3.62	1.95	13.50	15.71
14 SAB-L014-A02	227	0.01	<0.01	<0.001	<0.005	<0.001	59.09	0.73	3.91	3.02	20.32	134.86
14 SAB-L015-A01	46	0.14	0.29	0.001	<0.005	0.002	35.94	1.46	2.15	2.83	10.99	40.81
14 SAB-L015-A02	49	0.14	0.37	0.002	<0.005	0.004	32.46	1.50	2.89	3.23	14.35	31.54
14 SAB-L016-A01	26	0.06	<0.01	<0.001	<0.005	0.001	57.69	1.70	5.99	2.87	35.65	66.64
14 SAB-L017-A02	26	0.01	0.09	0.002	<0.005	<0.001	52.80	1.80	6.96	2.02	33.59	72.99
14 SAB-R005 A01	33	0.01	0.06	<0.001	<0.005	<0.001	68.91	1.14	2.43	2.32	9.62	30.48
14 SAB-R006 A01	14	0.06	0.01	<0.001	<0.005	<0.001	51.12	2.36	3.78	2.07	20.32	14.99
14 SAB-R006 A02	38	0.03	0.07	<0.001	<0.005	0.006	48.73	1.36	3.63	3.24	22.12	71.04
14 SAB-R010 A01	40	0.22	0.01	0.01	0.007	0.008	76.60	1.11	0.93	1.47	1.70	13.52
14 SAB-S130 A01	34	0.02	<0.01	<0.001	0.013	0.002	65.07	1.00	1.49	1.45	2.59	26.76
14 SUB-H029 B2	55	0.65	0.12	<0.001	0.005	0.006	76.46	1.02	0.76	1.18	0.98	7.2
15SAB-D196A01	75	0.04	0.04	0.001	<0.005	<0.001	61.43	1.99	3.62	1.65	9.70	11.75
15SAB-D196A02	42	0.07	0.06	0.001	<0.005	<0.001	62.32	1.41	1.98	1.63	4.70	12.84
15SAB-D196B01	49	0.04	0.15	0.001	<0.005	<0.001	56.2	1.76	2.37	1.54	5.26	13.81
15SAB-D237A01	86	0.02	0.06	0.001	<0.005	<0.001	48.04	0.50	0.80	1.17	0.93	51.85
15SAB-F013B01	50	0.07	0.16	0.003	<0.005	0.003	56.28	0.96	2.59	2.45	8.97	49.08
15SAB-F034B01	109	0.05	0.05	<0.001	<0.005	<0.001	35.02	0.86	1.80	1.34	3.16	97.72
15SAB-L028A01	123	0.06	0.09	0.001	<0.005	0.001	27.54	0.95	1.58	1.27	2.49	92.64
15SAB-L056B01	116	0.79	0.43	0.007	0.017	0.014	36.26	0.91	1.48	1.44	2.54	71.99
15SAB-L056B02	155	0.26	0.09	0.004	0.014	0.015	35.87	0.81	1.55	1.47	2.98	73.75
15SAB-L057B01	114	0.31	0.06	0.006	0.014	0.013	35.95	0.88	1.19	1.51	2.27	60.53
15SAB-L057B02	96	0.05	0.01	0.002	0.005	0.005	55.91	0.81	1.72	1.19	2.28	41.53
15SAB-L059B01	104	0.07	0.02	0.003	0.013	0.011	35.74	0.98	1.24	1.42	2.02	55.66
15SAB-L066B01	81	0.02	0.07	0.001	<0.005	0.001	46.74	0.98	1.75	1.22	2.62	42.39
15SAB-L076A01	23	0.08	0.04	0.001	0.008	0.015	67.2	1.33	0.86	1.05	0.88	7.05
15SAB-L076A02	12	0.05	0.05	0.002	0.011	0.017	67.91	1.44	1.08	1.09	0.82	6.37
15SAB-L077A01	32	0.07	0.02	0.004	0.011	0.019	69.06	1.13	0.47	0.97	0.50	5.06
15SAB-L077B01	18	0.02	0.01	0.001	<0.005	0.005	65.4	0.75	0.58	1.06	0.59	13.1
15SAB-L078A01	53	0.1	0.01	0.002	0.025	0.036	72.59	0.83	1.08	1.06	1.02	7.09
15SAB-L078A02	18	0.05	0.03	0.001	0.022	0.029	71.35	0.88	1.16	0.93	1.29	6.21
15SAB-L079A01	25	0.03	0.02	<0.001	0.013	0.025	69.39	0.97	1.05	1.09	1.00	7.69
15SAB-L080A01	43	0.04	0.04	0.002	<0.005	0.001	62.54	1.03	0.77	1.02	0.63	10.51
15SAB-L080A02	50	0.02	0.01	<0.001	<0.005	0.001	60.53	1.35	0.74	0.76	0.60	9.51
15SAB-L080A03	41	0.05	0.08	0.002	<0.005	<0.001	62.39	1.21	0.71	0.93	0.67	9.47
15SAB-L112A01	41	0.04	0.06	<0.001	<0.005	0.001	61.3	2.04	4.75	2.21	18.65	17.21
15SAB-L113A01	31	0.05	0.04	0.003	0.012	0.014	66.54	1.72	3.90	1.26	9.82	12.16
15SAB-L113B01	58	0.07	0.02	<0.001	<0.005	0.001	65.74	1.15	2.21	1.63	5.18	15.82
15SAB-L119A01	64	0.25	0.08	<0.001	0.011	0.01	68.42	0.90	1.16	1.38	1.66	19.99
15SAB-M005A01	155	0.02	0.42	<0.001	<0.005	<0.001	24.2	1.20	2.28	1.37	4.14	60.79
15SAB-M008A01	81	0.03	0.11	<0.001	<0.005	0.001	58.1	0.62	1.26	0.87	1.13	35.13
15SAB-M009A01	105	0.02	0.02	<0.001	<0.005	0.001	47.34	0.49	0.91	1.55	1.57	45.92
15SAB-R174C01	116	0.08	0.02	<0.001	0.01	0.008	39.79	0.87	1.17	1.38	1.72	60.04
15SAB-R174C02	113	0.09	0.02	<0.001	0.011	0.012	41.08	1.00	1.26	1.35	2.05	55.95
15SAB-W046B01	89	0.03	0.14	<0.001	<0.005	0.001	41.66	0.94	1.94	1.17	2.66	51.39

Appendix D - Mineral Geochemical Data

Olivine - Electron Microprobe Data																	
SAMPLE	Mineral	SiO2	Al2O3	TiO2	Cr2O3	FeO	MnO	MgO	NiO	ZnO	CaO	Na2O	K2O	BaO	F	Cl	Total
14SAB-H080A01-spot1-1	Olivine	36.74	0.01	0.02	0.00	30.44	0.38	31.54	0.29	0.00	0.00	0.00	0.00	0.00	0.00	0.00	99.43
14SAB-H080A01-spot1-2	Olivine	36.42	0.00	0.02	0.01	30.41	0.39	31.45	0.25	0.00	0.02	0.00	0.00	0.00	0.00	0.00	98.97
14SAB-H080A01-spot2-5	Olivine	37.35	0.00	0.00	0.01	31.09	0.39	31.39	0.24	0.00	0.02	0.00	0.00	0.00	0.00	0.00	100.48
14SAB-H080A01-spot2-6	Olivine	37.03	0.00	0.00	0.02	31.59	0.46	31.32	0.27	0.00	0.02	0.00	0.00	0.00	0.00	0.00	100.70
14SAB-H080B01 area 1-2	Olivine	38.88	0.00	0.00	0.01	19.30	0.31	40.91	0.31	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.74
14SAB-H080B01 area 1-3	Olivine	38.19	0.00	0.00	0.00	19.66	0.33	41.74	0.27	0.00	0.03	0.00	0.00	0.00	0.00	0.00	100.22
14SAB-H080B01 area 2-1	Olivine	38.69	0.00	0.00	0.00	19.29	0.29	41.59	0.33	0.00	0.03	0.00	0.00	0.00	0.00	0.00	100.21
14SAB-H080B01 area 2-2	Olivine	38.71	0.01	0.00	0.00	19.25	0.30	41.22	0.33	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.84
14SAB-L001A01-spot2-3	Olivine	38.01	0.00	0.00	0.00	25.10	0.42	36.28	0.30	0.00	0.00	0.00	0.00	0.00	0.00	0.00	100.11
14SAB-L001A01-spot2-4	Olivine	38.04	0.00	0.00	0.00	24.95	0.40	36.08	0.30	0.00	0.03	0.00	0.00	0.00	0.00	0.00	99.79
14SAB-L002A01-spot1-1	Olivine	38.01	0.00	0.00	0.00	19.73	0.27	40.11	0.37	0.00	0.02	0.00	0.00	0.00	0.00	0.00	98.50
14SAB-L002A01-spot1-2	Olivine	38.97	0.00	0.00	0.00	19.87	0.31	40.08	0.36	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.60
14SAB-L002A01-spot2-3	Olivine	39.19	0.00	0.00	0.01	19.91	0.27	40.10	0.29	0.00	0.01	0.00	0.00	0.00	0.00	0.00	99.79
14SAB-L002A02-spot1-1	Olivine	36.78	0.00	0.00	0.00	31.02	0.45	31.48	0.26	0.00	0.01	0.00	0.00	0.00	0.00	0.00	100.00
14SAB-L002A02-spot1-2	Olivine	36.57	0.00	0.01	0.00	31.32	0.41	31.39	0.27	0.00	0.02	0.00	0.00	0.00	0.00	0.00	100.01
14SAB-L002A02-spot2-2	Olivine	37.14	0.00	0.01	0.01	30.93	0.47	30.71	0.31	0.00	0.03	0.00	0.00	0.00	0.00	0.00	99.60
14SAB-L002A02-spot2-3	Olivine	37.04	0.00	0.00	0.01	30.92	0.46	31.03	0.28	0.00	0.04	0.00	0.00	0.00	0.00	0.00	99.79
14SAB-L003A01-spot1-2	Olivine	38.84	0.00	0.00	0.00	20.60	0.31	39.60	0.25	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.61
14SAB-L003A01-spot1-3	Olivine	38.32	0.00	0.00	0.03	20.43	0.25	39.56	0.33	0.00	0.01	0.00	0.00	0.00	0.00	0.00	98.93
14SAB-L003A01-spot2-2	Olivine	38.67	0.00	0.00	0.00	20.13	0.00	40.04	0.27	0.00	0.00	0.00	0.00	0.00	0.00	0.00	99.11
14SAB-L004A01-spot1-4	Olivine	37.53	0.00	0.00	0.00	28.76	0.00	32.90	0.30	0.00	0.01	0.00	0.00	0.00	0.00	0.00	99.51
14SAB-L004A01-spot2-6	Olivine	37.39	0.00	0.00	0.01	28.50	0.41	33.22	0.27	0.00	0.01	0.00	0.00	0.00	0.00	0.00	99.81
14SAB-L004A01-spot2-7	Olivine	37.09	0.00	0.00	0.00	28.62	0.33	32.84	0.31	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.21
14SAB-L005A01-spot1-3	Olivine	36.84	0.00	0.00	0.00	31.33	0.41	31.43	0.24	0.00	0.02	0.00	0.00	0.00	0.00	0.00	100.27
14SAB-L005A01-spot2-1	Olivine	36.75	0.00	0.00	0.00	31.31	0.45	31.74	0.26	0.00	0.01	0.00	0.00	0.00	0.00	0.00	100.51
14SAB-L006B01-spot1-5	Olivine	37.84	0.00	0.00	0.00	24.41	0.36	36.34	0.31	0.00	0.03	0.00	0.00	0.00	0.00	0.00	99.29
14SAB-L006B01-spot1-5	Olivine	37.99	0.00	0.00	0.00	24.51	0.31	36.70	0.31	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.84
14SAB-L006B01-spot2-1	Olivine	37.70	0.00	0.00	0.00	25.41	0.34	35.62	0.28	0.00	0.04	0.00	0.00	0.00	0.00	0.00	99.39
14SAB-L007A02-spot1-1	Olivine	37.94	0.00	0.00	0.00	26.22	0.32	35.51	0.28	0.00	0.03	0.00	0.00	0.00	0.00	0.00	100.30
14SAB-L007A02-spot1-2	Olivine	37.89	0.00	0.00	0.00	27.05	0.33	34.81	0.29	0.00	0.00	0.00	0.00	0.00	0.00	0.00	100.37
14SAB-L007A02-spot2-4	Olivine	37.85	0.01	0.00	0.00	27.34	0.38	34.48	0.25	0.00	0.01	0.00	0.00	0.00	0.00	0.00	100.32
14SAB-L007A02-spot2-5	Olivine	37.71	0.01	0.00	0.00	27.23	0.31	34.75	0.33	0.00	0.00	0.00	0.00	0.00	0.00	0.00	100.35
14SAB-L010C01-spot1-3	Olivine	37.13	0.00	0.00	0.02	28.55	0.41	33.58	0.17	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.88
14SAB-L010C01-spot1-4	Olivine	36.89	0.00	0.00	0.00	29.07	0.43	33.83	0.15	0.00	0.03	0.00	0.00	0.00	0.00	0.00	100.40
14SAB-L010C01-spot2-1	Olivine	37.14	0.00	0.00	0.03	28.67	0.42	33.66	0.12	0.00	0.02	0.00	0.00	0.00	0.00	0.00	100.06
14SAB-R010A01-spot1-1	Olivine	40.14	0.00	0.01	0.01	15.52	0.29	43.93	0.29	0.00	0.01	0.00	0.00	0.00	0.00	0.00	100.21
14SAB-R010A01-spot1-2	Olivine	40.26	0.00	0.01	0.02	15.63	0.34	43.88	0.35	0.00	0.03	0.00	0.00	0.00	0.00	0.00	100.52
14SAB-R010A01-spot2-2	Olivine	39.97	0.00	0.00	0.00	15.49	0.32	43.58	0.36	0.00	0.03	0.00	0.00	0.00	0.00	0.00	99.74
14SAB-R010A01-spot2-3	Olivine	40.15	0.00	0.00	0.01	15.45	0.33	43.86	0.38	0.00	0.02	0.00	0.00	0.00	0.00	0.00	100.20
14SAB-R010A01-spot2-4	Olivine	39.89	0.00	0.00	0.00	15.85	0.29	44.05	0.35	0.00	0.00	0.00	0.00	0.00	0.00	0.00	100.44
14SAB-S130A01-spot1-1	Olivine	38.86	0.00	0.00	0.00	21.37	0.37	38.92	0.20	0.00	0.01	0.00	0.00	0.00	0.00	0.00	99.73
14SAB-S130A01-spot1-2	Olivine	38.88	0.00	0.00	0.00	21.60	0.34	39.10	0.20	0.00	0.01	0.00	0.00	0.00	0.00	0.00	100.13
14SAB-S130A01-spot2-7	Olivine	38.51	0.00	0.00	0.00	21.48	0.36	39.19	0.19	0.00	0.01	0.00	0.00	0.00	0.00	0.00	99.75
15SAB-D196A01-spot2-5	Olivine	38.63	0.00	0.00	0.00	24.84	0.37	37.35	0.03	0.00	0.01	0.00	0.00	0.00	0.00	0.00	101.24
15SAB-D196A01-spot3-2	Olivine	38.30	0.00	0.02	0.00	24.24	0.39	37.03	0.02	0.00	0.01	0.00	0.00	0.00	0.00	0.00	100.01
15SAB-L076A01 area 1-5	Olivine	37.82	0.00	0.00	0.00	23.21	0.30	37.82	0.17	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.36
15SAB-L076A01 area 1-6	Olivine	38.19	0.00	0.00	0.00	23.38	0.37	37.94	0.23	0.00	0.02	0.00	0.00	0.00	0.00	0.00	100.12
15SAB-L076A01 area 2-1	Olivine	38.59	0.00	0.00	0.00	23.08	0.33	37.89	0.23	0.00	0.04	0.00	0.00	0.00	0.00	0.00	100.16
15SAB-L076A01 area 2-2	Olivine	38.14	0.00	0.00	0.00	22.93	0.31	37.73	0.24	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.36
15SAB-L112A01-spot2-5	Olivine	37.57	0.00	0.00	0.01	26.91	0.38	35.46	0.09	0.00	0.03	0.00	0.00	0.00	0.00	0.00	100.45
15SAB-L119A01-spot1-1	Olivine	39.77	0.00	0.00	0.00	20.38	0.32	40.61	0.29	0.00	0.00	0.00	0.00	0.00	0.00	0.00	101.36
15SAB-L119A01-spot1-2	Olivine	39.34	0.00	0.00	0.00	20.23	0.28	39.78	0.29	0.00	0.00	0.00	0.00	0.00	0.00	0.00	99.91
15SAB-L119A01-spot2-3	Olivine	38.67	0.03	0.01	0.03	20.54	0.30	39.63	0.23	0.00	0.03	0.00	0.00	0.00	0.00	0.00	99.47
15SAB-L119A01-spot2-4	Olivine	38.50	0.00	0.01	0.01	20.39	0.30	39.62	0.25	0.00	0.03	0.00	0.00	0.00	0.00	0.00	99.09
L012A01-circle2-1	Olivine	39.65	0.01	0.00	0.02	18.02	0.22	41.99	0.13	0.00	0.00	0.00	0.00	0.00	0.00	0.00	100.05
L012A01-circle2-2	Olivine	39.51	0.00	0.00	0.00	17.67	0.25	41.30	0.16	0.00	0.02	0.00	0.00	0.00	0.00	0.00	98.90
L014-A01-spot2-1	Olivine	39.08	0.00	0.01	0.02	20.34	0.30	40.89	0.10	0.00	0.00	0.00	0.00	0.00	0.00	0.00	100.72
L014-A01-spot2-2	Olivine	39.06	0.02	0.00	0.00	20.18	0.29	41.00	0.11	0.00	0.01	0.00	0.00	0.00	0.00	0.00	100.66
L014A01-spot1-3	Olivine	38.83	0.00	0.00	0.00	20.58	0.29	40.29	0.10	0.00	0.01	0.00	0.00	0.00	0.00	0.00	100.10
L014A01-spot1-4	Olivine	38.76	0.00	0.00	0.00	20.80	0.29	39.95	0.10	0.00	0.01	0.00	0.00	0.00	0.00	0.00	99.91
L076-A02-spot2-2	Olivine	38.61	0.00	0.02	0.01	23.21	0.34	38.49	0.22	0.00	0.00	0.00	0.00	0.00	0.00	0.00	100.90
L077-A01-spot2-2	Olivine	39.17	0														

Olivine - Atoms Per Formula Unit																	
SAMPLE	Si	Al	Ti	Cr	Fe	Mn	Mg	Ni	Zn	Ca	Na	K	Ba	Σ cations	F	Cl	Fo%
14SAB-H080A01-spot1-1	1.008	0.000	0.000	0.000	0.383	0.005	1.592	0.003	0.000	0.000	0.000	0.000	0.000	2.991	0	0	81
14SAB-H080A01-spot1-2	1.015	0.000	0.000	0.000	0.380	0.005	1.582	0.003	0.000	0.001	0.000	0.000	0.000	2.985	0	0	81
14SAB-H080A01-spot2-5	1.000	0.000	0.000	0.000	0.443	0.006	1.547	0.002	0.000	0.000	0.000	0.000	0.000	3.000	0	0	78
14SAB-H080A01-spot2-6	1.001	0.000	0.000	0.000	0.450	0.006	1.539	0.002	0.000	0.000	0.000	0.000	0.000	2.999	0	0	77
14SAB-H080B01 area 1-2	0.999	0.000	0.000	0.000	0.435	0.007	1.558	0.002	0.000	0.000	0.000	0.000	0.000	3.001	0	0	78
14SAB-H080B01 area 1-3	0.998	0.001	0.000	0.000	0.431	0.006	1.562	0.002	0.000	0.000	0.000	0.000	0.000	3.001	0	0	78
14SAB-H080B01 area 2-1	0.999	0.000	0.000	0.000	0.503	0.007	1.485	0.005	0.000	0.000	0.000	0.000	0.000	3.000	0	0	75
14SAB-H080B01 area 2-2	1.005	0.001	0.000	0.000	0.451	0.006	1.527	0.004	0.000	0.000	0.000	0.000	0.000	2.995	0	0	77
14SAB-L001A01-spot2-3	1.002	0.001	0.000	0.000	0.435	0.007	1.549	0.005	0.000	0.000	0.000	0.000	0.000	2.998	0	0	78
14SAB-L001A01-spot2-4	1.002	0.000	0.000	0.000	0.432	0.007	1.553	0.005	0.000	0.000	0.000	0.000	0.000	2.998	0	0	78
14SAB-L002A01-spot1-1	0.995	0.000	0.000	0.000	0.432	0.006	1.565	0.008	0.000	0.000	0.000	0.000	0.000	3.005	0	0	78
14SAB-L002A01-spot1-2	1.007	0.000	0.000	0.000	0.429	0.007	1.543	0.007	0.000	0.000	0.000	0.000	0.000	2.993	0	0	78
14SAB-L002A01-spot2-3	1.009	0.000	0.000	0.000	0.429	0.006	1.540	0.006	0.000	0.000	0.000	0.000	0.000	2.990	0	0	78
14SAB-L002A02-spot1-1	1.000	0.000	0.000	0.000	0.706	0.010	1.277	0.006	0.000	0.000	0.000	0.000	0.000	2.999	0	0	64
14SAB-L002A02-spot1-2	0.997	0.000	0.000	0.000	0.714	0.010	1.276	0.006	0.000	0.001	0.000	0.000	0.000	3.003	0	0	64
14SAB-L002A02-spot2-2	1.013	0.000	0.000	0.000	0.706	0.011	1.249	0.007	0.000	0.001	0.000	0.000	0.000	2.987	0	0	64
14SAB-L002A02-spot2-3	1.009	0.000	0.000	0.000	0.704	0.011	1.260	0.006	0.000	0.001	0.000	0.000	0.000	2.991	0	0	64
14SAB-L003A01-spot1-2	1.003	0.000	0.000	0.000	0.541	0.008	1.437	0.007	0.000	0.001	0.000	0.000	0.000	2.997	0	0	73
14SAB-L003A01-spot1-3	1.002	0.000	0.000	0.000	0.540	0.007	1.442	0.007	0.000	0.001	0.000	0.000	0.000	2.998	0	0	73
14SAB-L003A01-spot2-2	1.003	0.000	0.000	0.000	0.566	0.008	1.413	0.006	0.000	0.001	0.000	0.000	0.000	2.997	0	0	71
14SAB-L004A01-spot1-4	1.000	0.000	0.000	0.000	0.415	0.007	1.570	0.007	0.000	0.001	0.000	0.000	0.000	2.999	0	0	79
14SAB-L004A01-spot2-6	0.982	0.000	0.000	0.000	0.423	0.007	1.600	0.006	0.000	0.001	0.000	0.000	0.000	3.018	0	0	79
14SAB-L004A01-spot2-7	0.992	0.000	0.000	0.000	0.414	0.006	1.589	0.007	0.000	0.001	0.000	0.000	0.000	3.008	0	0	79
14SAB-L005A01-spot1-3	0.996	0.000	0.000	0.000	0.414	0.007	1.581	0.007	0.000	0.000	0.000	0.000	0.000	3.004	0	0	79
14SAB-L005A01-spot2-1	0.996	0.000	0.000	0.000	0.511	0.007	1.485	0.004	0.000	0.001	0.000	0.000	0.000	3.004	0	0	74
14SAB-L006B01-spot1-5	0.998	0.000	0.000	0.000	0.511	0.008	1.479	0.005	0.000	0.001	0.000	0.000	0.000	3.002	0	0	74
14SAB-L006B01-spot1-5	1.006	0.000	0.000	0.000	0.503	0.007	1.472	0.005	0.000	0.001	0.000	0.000	0.000	2.994	0	0	75
14SAB-L006B01-spot2-1	1.003	0.000	0.000	0.000	0.504	0.007	1.478	0.005	0.000	0.001	0.000	0.000	0.000	2.997	0	0	75
14SAB-L007A02-spot1-1	1.008	0.000	0.000	0.000	0.326	0.006	1.645	0.006	0.000	0.000	0.000	0.000	0.000	2.992	0	0	83
14SAB-L007A02-spot1-2	1.009	0.000	0.000	0.000	0.328	0.007	1.639	0.007	0.000	0.001	0.000	0.000	0.000	2.991	0	0	83
14SAB-L007A02-spot2-4	1.009	0.000	0.000	0.000	0.327	0.007	1.640	0.007	0.000	0.001	0.000	0.000	0.000	2.991	0	0	83
14SAB-L007A02-spot2-5	1.009	0.000	0.000	0.000	0.325	0.007	1.643	0.008	0.000	0.000	0.000	0.000	0.000	2.991	0	0	84
14SAB-L010C01-spot1-3	1.002	0.000	0.000	0.000	0.333	0.006	1.650	0.007	0.000	0.000	0.000	0.000	0.000	2.998	0	0	83
14SAB-L010C01-spot1-4	1.003	0.000	0.000	0.000	0.580	0.007	1.400	0.006	0.000	0.001	0.000	0.000	0.000	2.997	0	0	71
14SAB-L010C01-spot2-1	1.005	0.000	0.000	0.000	0.600	0.008	1.377	0.006	0.000	0.000	0.000	0.000	0.000	2.995	0	0	70
14SAB-R010A01-spot1-1	1.006	0.000	0.000	0.000	0.608	0.009	1.366	0.005	0.000	0.000	0.000	0.000	0.000	2.994	0	0	69
14SAB-R010A01-spot1-2	1.002	0.000	0.000	0.000	0.605	0.007	1.376	0.007	0.000	0.000	0.000	0.000	0.000	2.998	0	0	69
14SAB-R010A01-spot2-2	1.006	0.000	0.000	0.000	0.446	0.007	1.529	0.005	0.000	0.000	0.000	0.000	0.000	2.994	0	0	77
14SAB-R010A01-spot2-3	1.000	0.000	0.000	0.001	0.446	0.006	1.540	0.007	0.000	0.000	0.000	0.000	0.000	2.999	0	0	78
14SAB-R010A01-spot2-4	1.004	0.000	0.000	0.000	0.437	0.000	1.549	0.006	0.000	0.000	0.000	0.000	0.000	2.996	0	0	78
14SAB-S130A01-spot1-1	1.009	0.000	0.000	0.000	0.432	0.007	1.536	0.006	0.000	0.000	0.000	0.000	0.000	2.991	0	0	78
14SAB-S130A01-spot1-2	1.013	0.000	0.000	0.000	0.436	0.006	1.527	0.006	0.000	0.000	0.000	0.000	0.000	2.987	0	0	78
14SAB-S130A01-spot2-7	1.003	0.001	0.000	0.001	0.446	0.007	1.533	0.005	0.000	0.001	0.000	0.000	0.000	2.996	0	0	77
15SAB-D196A01-spot2-5	1.002	0.000	0.000	0.000	0.444	0.007	1.538	0.005	0.000	0.001	0.000	0.000	0.000	2.997	0	0	78
15SAB-D196A01-spot3-2	1.003	0.000	0.000	0.000	0.695	0.009	1.283	0.006	0.000	0.000	0.000	0.000	0.000	2.997	0	0	65
15SAB-L076A01 area 1-5	0.999	0.000	0.000	0.000	0.698	0.009	1.287	0.006	0.000	0.000	0.000	0.000	0.000	3.000	0	0	65
15SAB-L076A01 area 1-6	1.009	0.000	0.000	0.000	0.702	0.009	1.264	0.005	0.000	0.000	0.000	0.000	0.000	2.991	0	0	64
15SAB-L076A01 area 2-1	1.002	0.000	0.000	0.001	0.715	0.010	1.264	0.006	0.000	0.001	0.000	0.000	0.000	2.998	0	0	64
15SAB-L076A01 area 2-2	1.011	0.000	0.000	0.000	0.648	0.000	1.322	0.007	0.000	0.000	0.000	0.000	0.000	2.989	0	0	67
15SAB-L112A01-spot2-5	1.006	0.000	0.000	0.000	0.641	0.009	1.332	0.006	0.000	0.000	0.000	0.000	0.000	2.994	0	0	68
15SAB-L119A01-spot1-1	1.005	0.000	0.000	0.000	0.649	0.008	1.327	0.007	0.000	0.000	0.000	0.000	0.000	2.995	0	0	67
15SAB-L119A01-spot1-2	1.009	0.000	0.000	0.000	0.464	0.008	1.506	0.004	0.000	0.000	0.000	0.000	0.000	2.991	0	0	76
15SAB-L119A01-spot2-3	1.006	0.000	0.000	0.000	0.467	0.008	1.508	0.004	0.000	0.000	0.000	0.000	0.000	2.994	0	0	76
15SAB-L119A01-spot2-4	1.001	0.000	0.000	0.000	0.467	0.008	1.519	0.004	0.000	0.000	0.000	0.000	0.000	2.999	0	0	76
L012A01-circle2-1-oliv	1.002	0.000	0.000	0.000	0.554	0.009	1.426	0.006	0.000	0.000	0.000	0.000	0.000	2.998	0	0	72
L012A01-circle2-2-oliv	1.005	0.000	0.000	0.000	0.551	0.009	1.422	0.006	0.000	0.001	0.000	0.000	0.000	2.995	0	0	72
L014-A01-spot2-1	0.996	0.000	0.000	0.000	0.596	0.009	1.401	0.002	0.000	0.001	0.000	0.000	0.000	3.004	0	0	70
L014-A01-spot2-2	1.003	0.000	0.000	0.000	0.539	0.008	1.446	0.001	0.000	0.000	0.000	0.000	0.000	2.997	0	0	73
L014A01-spot1-3	1.005	0.000	0.000	0.000	0.532	0.009	1.448	0.001	0.000	0.000	0.000	0.000	0.000	2.995	0	0	73
L014A01-spot1-4	0.998	0.000	0.000	0.000	0.642	0.009	1.346	0.004	0.000	0.001	0.000	0.000	0.000	3.001	0	0	68
L076-A02-spot2-2	0.990	0.000	0.000	0.000	0.653	0.010	1.354	0.003	0.000	0.001	0.000	0.000	0.000	3.010	0	0	67
L077-A01-spot2-2	0.997	0.000	0.000	0.001	0.644	0.010	1.347	0.003	0.000	0.001	0.000	0.000	0.000	3.002	0	0	68
L079-A01-spot1-1	1.000	0.000	0.000	0.000	0.711	0.009	1.273	0.005	0.000	0.001	0.000	0.000	0.000	3.000	0	0	64
L079-A01-spot2-3	0.996	0.000	0.000	0.000	0.710	0.010	1.282	0.006	0.000	0.000	0.000	0.000	0.000	3.004	0	0	64

Clinopyroxene - Electron Microprobe Data																	
SAMPLE	Mineral	SiO ₂	Al ₂ O ₃	TiO ₂	Cr ₂ O ₃	FeO	MnO	MgO	NiO	ZnO	CaO	Na ₂ O	K ₂ O	BaO	F	Cl	Total
14SAB-L012A02-spot1-3	Clinopyroxene (rim)	54.16	1.59	0.12	0.16	4.82	0.13	15.27	0.00	0.00	23.28	0.22	0.01	0.00	0.00	0.00	99.77
14SAB-L012A02-spot1-4	Clinopyroxene (body)	53.21	2.19	0.21	0.27	4.99	0.14	15.33	0.00	0.00	22.58	0.33	0.00	0.00	0.08	0.01	99.33
14SAB-L012A02-spot1-5	Clinopyroxene (core)	51.91	3.83	0.37	0.56	5.48	0.09	15.04	0.00	0.00	20.55	0.51	0.10	0.00	0.03	0.00	98.46
14SAB-L013A02-spot1-2-	Clinopyroxene	54.03	1.97	0.03	0.00	3.70	0.15	16.23	0.01	0.00	24.42	0.07	0.00	0.00	0.00	0.00	100.63
14SAB-L006A01-spot1-1	Clinopyroxene	50.28	2.12	0.32	0.02	15.74	0.41	8.40	0.01	0.00	21.82	0.41	0.00	0.00	0.00	0.00	99.52
14SAB-L006A01-spot1-3	Clinopyroxene	50.68	1.73	0.28	0.00	15.62	0.40	8.55	0.00	0.00	21.88	0.42	0.00	0.00	0.00	0.00	99.57
14SAB-L006A01-spot2-3	Clinopyroxene	50.67	2.02	0.31	0.00	15.57	0.40	8.55	0.02	0.00	21.91	0.47	0.00	0.00	0.00	0.00	99.90
14SAB-L006A01-spot2-4	Clinopyroxene	51.45	0.79	0.09	0.00	15.06	0.43	9.19	0.02	0.00	22.53	0.30	0.00	0.00	0.00	0.00	99.86
14SAB-L006B01-spot1-2	Clinopyroxene	53.94	1.92	0.32	0.03	4.19	0.18	15.63	0.04	0.00	24.11	0.02	0.00	0.00	0.00	0.00	100.39
14SAB-L006B01-spot2-3	Clinopyroxene	52.07	2.91	0.57	0.09	4.24	0.11	15.25	0.04	0.00	24.01	0.03	0.00	0.00	0.00	0.00	99.32
14SAB-L008A02-spot1-3	Clinopyroxene	51.63	2.44	0.40	0.07	9.68	0.19	12.24	0.06	0.00	23.09	0.23	0.01	0.00	0.00	0.00	100.04
14SAB-L008A02-spot1-4	Clinopyroxene	51.72	2.45	0.39	0.06	9.79	0.21	12.25	0.07	0.00	22.70	0.19	0.01	0.00	0.00	0.00	99.83
14SAB-L008A02-spot2-1	Clinopyroxene	51.41	2.42	0.40	0.13	9.60	0.22	12.27	0.07	0.00	22.40	0.18	0.00	0.00	0.00	0.00	99.11
14SAB-L008A02-spot2-2	Clinopyroxene	51.66	2.53	0.45	0.12	9.81	0.18	12.34	0.05	0.00	22.38	0.22	0.00	0.00	0.00	0.00	99.73
14SAB-L010A01-spot1-1	Clinopyroxene	52.41	2.18	0.25	0.03	8.49	0.24	13.26	0.00	0.00	22.52	0.39	0.00	0.00	0.00	0.00	99.78
14SAB-L010A01-spot2-2	Clinopyroxene	52.83	2.14	0.29	0.01	8.34	0.22	13.28	0.02	0.00	22.43	0.37	0.00	0.00	0.00	0.00	99.95
14SAB-L004A01-spot2-2	Clinopyroxene	53.37	2.12	0.37	0.06	4.90	0.12	15.06	0.04	0.00	23.32	0.07	0.00	0.00	0.00	0.00	99.42
14SAB-L004A01-spot2-5	Clinopyroxene	52.31	2.68	0.46	0.10	4.96	0.14	14.97	0.06	0.00	23.43	0.08	0.00	0.00	0.00	0.00	99.18
14SAB-L001B01-spot1-1	Clinopyroxene	53.05	0.65	0.05	0.06	10.68	0.25	12.55	0.05	0.00	22.74	0.15	0.00	0.00	0.00	0.00	100.22
14SAB-L001B01-spot1-2	Clinopyroxene	53.44	1.11	0.18	0.11	10.71	0.17	12.30	0.05	0.00	22.17	0.27	0.00	0.00	0.00	0.00	100.50
14SAB-L001B01-spot2-2	Clinopyroxene	52.03	1.95	0.30	0.08	9.97	0.24	12.64	0.04	0.00	22.10	0.28	0.00	0.00	0.00	0.00	99.63
14SAB-L001B01-spot2-3	Clinopyroxene	52.96	1.36	0.22	0.06	9.31	0.22	13.25	0.08	0.00	22.19	0.26	0.00	0.00	0.00	0.00	99.89
14SAB-L003B01-spot1-3	Clinopyroxene	53.10	1.28	0.14	0.05	8.85	0.42	13.62	0.02	0.00	22.22	0.27	0.00	0.00	0.00	0.00	99.96
14SAB-L003B01-spot1-4	Clinopyroxene	53.21	1.29	0.16	0.04	9.13	0.33	13.65	0.00	0.00	22.26	0.27	0.00	0.00	0.00	0.00	100.35
14SAB-L003B01-spot2-4	Clinopyroxene	52.79	1.28	0.14	0.06	8.72	0.32	13.79	0.01	0.00	22.27	0.24	0.00	0.00	0.00	0.00	99.62
14SAB-L003B01-spot2-5	Clinopyroxene	53.29	1.32	0.14	0.08	8.47	0.35	13.53	0.01	0.00	22.81	0.30	0.00	0.00	0.00	0.00	100.30
15SAB-L056B01-spot1-1	Clinopyroxene	51.79	1.42	0.19	0.01	12.45	0.31	11.07	0.02	0.00	22.57	0.23	0.00	0.00	0.00	0.00	100.06
15SAB-L056B01-spot1-3	Clinopyroxene	52.04	1.40	0.19	0.02	12.08	0.27	11.08	0.04	0.00	22.47	0.25	0.00	0.00	0.00	0.00	99.86
15SAB-L056B01-spot2-3	Clinopyroxene	51.57	1.79	0.25	0.04	12.84	0.31	10.88	0.00	0.00	21.64	0.34	0.00	0.00	0.00	0.00	99.66
15SAB-D196A01-spot2-1	Clinopyroxene	51.06	3.95	0.11	0.01	5.51	0.22	14.57	0.00	0.00	23.23	0.17	0.00	0.00	0.00	0.00	98.84
15SAB-D196A01-spot2-2	Clinopyroxene	52.17	2.97	0.13	0.01	4.90	0.18	15.07	0.01	0.00	23.90	0.15	0.00	0.00	0.00	0.00	99.47
15SAB-M005A01-spot1-2	Clinopyroxene	51.61	1.18	0.13	0.00	12.50	0.59	11.25	0.01	0.00	22.18	0.27	0.00	0.00	0.00	0.00	99.72
15SAB-M005A01-spot2-3	Clinopyroxene	51.32	1.40	0.13	0.00	13.10	0.60	10.77	0.01	0.00	21.86	0.30	0.02	0.00	0.00	0.00	99.53
14SAB-L011A02-spot1-1	Clinopyroxene	51.30	2.28	0.30	0.04	12.57	0.38	11.28	0.00	0.00	21.66	0.36	0.00	0.00	0.00	0.00	100.18
14SAB-L011A02-spot2-2	Clinopyroxene	51.26	2.78	0.36	0.04	12.35	0.28	10.81	0.03	0.00	21.95	0.43	0.00	0.00	0.00	0.00	100.28
14SAB-L010C01-spot1-1	Clinopyroxene	52.50	3.16	0.48	0.17	5.31	0.14	14.82	0.01	0.00	23.65	0.14	0.00	0.00	0.00	0.00	100.37
14SAB-L010C01-spot2-2	Clinopyroxene	51.94	3.14	0.45	0.20	6.06	0.24	14.53	0.02	0.00	23.45	0.22	0.00	0.00	0.00	0.00	100.25
14SAB-L005A02-spot1-3	Clinopyroxene	52.71	0.75	0.11	0.05	10.28	0.23	12.52	0.00	0.00	22.57	0.20	0.00	0.00	0.00	0.00	99.43
14SAB-L005A02-spot2-4	Clinopyroxene	52.28	0.81	0.08	0.06	10.01	0.24	12.60	0.03	0.00	23.03	0.18	0.01	0.00	0.00	0.00	99.33

Clinopyroxene - Atoms Per Formula Unit																		
SAMPLE	Mineral	Si	Al	Ti	Cr	Fe	Mn	Mg	Ni	Zn	Ca	Na	K	Ba	Σ cations	F	Cl	%Di
14SAB-L012A02-spot1-3	Clinopyroxene (rim)	1.99	0.07	0.00	0.00	0.15	0.00	0.84	0.00	0.00	0.91	0.02	0.00	0.00	3.98	0.00	0.00	85
14SAB-L012A02-spot1-4	Clinopyroxene (body)	1.96	0.10	0.01	0.01	0.15	0.00	0.84	0.00	0.00	0.89	0.02	0.00	0.00	3.99	0.01	0.00	85
14SAB-L012A02-spot1-5	Clinopyroxene (core)	1.93	0.17	0.01	0.02	0.17	0.00	0.83	0.00	0.00	0.82	0.04	0.00	0.00	3.99	0.00	0.00	83
14SAB-L013A02-spot1-2-	Clinopyroxene	1.96	0.08	0.00	0.00	0.11	0.00	0.88	0.00	0.00	0.95	0.01	0.00	0.00	4.00	0.00	0.00	89
14SAB-L006A01-spot1-1	Clinopyroxene	1.95	0.10	0.01	0.00	0.51	0.01	0.49	0.00	0.00	0.91	0.03	0.00	0.00	4.01	0.00	0.00	49
14SAB-L006A01-spot1-3	Clinopyroxene	1.96	0.08	0.01	0.00	0.51	0.01	0.49	0.00	0.00	0.91	0.03	0.00	0.00	4.00	0.00	0.00	49
14SAB-L006A01-spot2-3	Clinopyroxene	1.96	0.09	0.01	0.00	0.50	0.01	0.49	0.00	0.00	0.91	0.03	0.00	0.00	4.01	0.00	0.00	49
14SAB-L006A01-spot2-4	Clinopyroxene	1.98	0.04	0.00	0.00	0.49	0.01	0.53	0.00	0.00	0.93	0.02	0.00	0.00	4.01	0.00	0.00	52
14SAB-L006B01-spot1-2	Clinopyroxene	1.97	0.08	0.01	0.00	0.13	0.01	0.85	0.00	0.00	0.94	0.00	0.00	0.00	3.98	0.00	0.00	87
14SAB-L006B01-spot2-3	Clinopyroxene	1.92	0.13	0.02	0.00	0.13	0.00	0.84	0.00	0.00	0.95	0.00	0.00	0.00	4.00	0.00	0.00	87
14SAB-L008A02-spot1-3	Clinopyroxene	1.94	0.11	0.01	0.00	0.30	0.01	0.69	0.00	0.00	0.93	0.02	0.00	0.00	4.00	0.00	0.00	69
14SAB-L008A02-spot1-4	Clinopyroxene	1.94	0.11	0.01	0.00	0.31	0.01	0.69	0.00	0.00	0.91	0.01	0.00	0.00	4.00	0.00	0.00	69
14SAB-L008A02-spot2-1	Clinopyroxene	1.94	0.11	0.01	0.00	0.30	0.01	0.69	0.00	0.00	0.91	0.01	0.00	0.00	3.99	0.00	0.00	69
14SAB-L008A02-spot2-2	Clinopyroxene	1.94	0.11	0.01	0.00	0.31	0.01	0.69	0.00	0.00	0.90	0.02	0.00	0.00	4.00	0.00	0.00	69
14SAB-L010A01-spot1-1	Clinopyroxene	1.96	0.10	0.01	0.00	0.27	0.01	0.74	0.00	0.00	0.90	0.03	0.00	0.00	4.00	0.00	0.00	74
14SAB-L010A01-spot2-2	Clinopyroxene	1.97	0.09	0.01	0.00	0.26	0.01	0.74	0.00	0.00	0.89	0.03	0.00	0.00	3.99	0.00	0.00	74
14SAB-L004A01-spot2-2	Clinopyroxene	1.97	0.09	0.01	0.00	0.15	0.00	0.83	0.00	0.00	0.92	0.01	0.00	0.00	3.98	0.00	0.00	85
14SAB-L004A01-spot2-5	Clinopyroxene	1.94	0.12	0.01	0.00	0.15	0.00	0.83	0.00	0.00	0.93	0.01	0.00	0.00	3.99	0.00	0.00	84
14SAB-L001B01-spot1-1	Clinopyroxene	1.99	0.03	0.00	0.00	0.34	0.01	0.70	0.00	0.00	0.91	0.01	0.00	0.00	4.00	0.00	0.00	68
14SAB-L001B01-spot1-2	Clinopyroxene	1.99	0.05	0.00	0.00	0.33	0.01	0.68	0.00	0.00	0.89	0.02	0.00	0.00	3.98	0.00	0.00	67
14SAB-L001B01-spot2-2	Clinopyroxene	1.96	0.09	0.01	0.00	0.31	0.01	0.71	0.00	0.00	0.89	0.02	0.00	0.00	4.00	0.00	0.00	69
14SAB-L001B01-spot2-3	Clinopyroxene	1.98	0.06	0.01	0.00	0.29	0.01	0.74	0.00	0.00	0.89	0.02	0.00	0.00	3.99	0.00	0.00	72
14SAB-L003B01-spot1-3	Clinopyroxene	1.98	0.06	0.00	0.00	0.28	0.01	0.76	0.00	0.00	0.89	0.02	0.00	0.00	4.00	0.00	0.00	73
14SAB-L003B01-spot1-4	Clinopyroxene	1.98	0.06	0.00	0.00	0.28	0.01	0.76	0.00	0.00	0.89	0.02	0.00	0.00	4.00	0.00	0.00	73
14SAB-L003B01-spot2-4	Clinopyroxene	1.98	0.06	0.00	0.00	0.27	0.01	0.77	0.00	0.00	0.89	0.02	0.00	0.00	4.00	0.00	0.00	74
14SAB-L003B01-spot2-5	Clinopyroxene	1.98	0.06	0.00	0.00	0.26	0.01	0.75	0.00	0.00	0.91	0.02	0.00	0.00	4.00	0.00	0.00	74
15SAB-L056B01-spot1-1	Clinopyroxene	1.97	0.06	0.01	0.00	0.40	0.01	0.63	0.00	0.00	0.92	0.02	0.00	0.00	4.00	0.00	0.00	61
15SAB-L056B01-spot1-3	Clinopyroxene	1.98	0.06	0.01	0.00	0.38	0.01	0.63	0.00	0.00	0.91	0.02	0.00	0.00	4.00	0.00	0.00	62
15SAB-L056B01-spot2-3	Clinopyroxene	1.96	0.08	0.01	0.00	0.41	0.01	0.62	0.00	0.00	0.88	0.03	0.00	0.00	4.00	0.00	0.00	60
15SAB-D196A01-spot2-1	Clinopyroxene	1.90	0.17	0.00	0.00	0.17	0.01	0.81	0.00	0.00	0.93	0.01	0.00	0.00	4.01	0.00	0.00	82
15SAB-D196A01-spot2-2	Clinopyroxene	1.93	0.13	0.00	0.00	0.15	0.01	0.83	0.00	0.00	0.95	0.01	0.00	0.00	4.01	0.00	0.00	85
15SAB-M005A01-spot1-2	Clinopyroxene	1.97	0.05	0.00	0.00	0.40	0.02	0.64	0.00	0.00	0.91	0.02	0.00	0.00	4.01	0.00	0.00	62
15SAB-M005A01-spot2-3	Clinopyroxene	1.97	0.06	0.00	0.00	0.42	0.02	0.62	0.00	0.00	0.90	0.02	0.00	0.00	4.01	0.00	0.00	59
14SAB-L011A02-spot1-1	Clinopyroxene	1.94	0.10	0.01	0.00	0.40	0.01	0.64	0.00	0.00	0.88	0.03	0.00	0.00	4.01	0.00	0.00	62
14SAB-L011A02-spot2-2	Clinopyroxene	1.94	0.12	0.01	0.00	0.39	0.01	0.61	0.00	0.00	0.89	0.03	0.00	0.00	4.00	0.00	0.00	61
14SAB-L010C01-spot1-1	Clinopyroxene	1.92	0.14	0.01	0.00	0.16	0.00	0.81	0.00	0.00	0.93	0.01	0.00	0.00	4.00	0.00	0.00	83
14SAB-L010C01-spot2-2	Clinopyroxene	1.92	0.14	0.01	0.01	0.19	0.01	0.80	0.00	0.00	0.93	0.02	0.00	0.00	4.01	0.00	0.00	81
14SAB-L005A02-spot1-3	Clinopyroxene	1.99	0.03	0.00	0.00	0.32	0.01	0.70	0.00	0.00	0.91	0.02	0.00	0.00	4.00	0.00	0.00	68
14SAB-L005A02-spot2-4	Clinopyroxene	1.98	0.04	0.00	0.00	0.32	0.01	0.71	0.00	0.00	0.93	0.01	0.00	0.00	4.01	0.00	0.00	69

Orthopyroxene - Electron Microprobe Data																	
SAMPLE	Mineral	SiO ₂	Al ₂ O ₃	TiO ₂	Cr ₂ O ₃	FeO	MnO	MgO	NiO	ZnO	CaO	Na ₂ O	K ₂ O	BaO	F	Cl	Total
15SAB-M005A01-spot1-1	Orthopyroxene	50.31	0.84	0.10	0.01	32.58	1.50	14.76	0.00	0.00	0.61	0.00	0.00	0.00	0.00	0.00	100.72
15SAB-L056B01-spot1-2	Orthopyroxene	50.26	0.71	0.10	0.03	31.41	0.70	14.96	0.04	0.00	0.87	0.00	0.01	0.00	0.00	0.00	99.08
14SAB-L011A02-spot2-3	Orthopyroxene	50.60	0.91	0.08	0.03	31.14	0.73	15.49	0.00	0.00	0.71	0.00	0.02	0.00	0.00	0.00	99.72
14SAB-L005A02-spot1-2	Orthopyroxene	51.53	0.67	0.09	0.02	28.78	0.61	17.43	0.00	0.00	0.57	0.00	0.00	0.00	0.00	0.00	99.71
14SAB-L001B01-spot2-5	Orthopyroxene	51.28	1.09	0.14	0.00	26.91	0.50	18.34	0.04	0.00	0.74	0.00	0.00	0.00	0.00	0.00	99.05
14SAB-L001B01-spot2-4	Orthopyroxene	52.11	0.83	0.10	0.02	27.07	0.64	18.53	0.04	0.00	0.61	0.00	0.00	0.00	0.00	0.00	99.94
14SAB-L003B01-spot2-2	Orthopyroxene	52.69	0.87	0.08	0.04	24.88	0.79	20.54	0.00	0.00	0.63	0.00	0.00	0.00	0.00	0.00	100.50
14SAB-L003B01-spot2-3	Orthopyroxene	52.71	0.76	0.05	0.03	24.78	0.87	20.64	0.02	0.00	0.53	0.00	0.00	0.00	0.00	0.00	100.38
14SAB-L010A01-spot1-2	Orthopyroxene	52.47	0.90	0.05	0.01	23.87	0.56	20.67	0.02	0.00	0.64	0.00	0.00	0.00	0.00	0.00	99.19
14SAB-L010A01-spot2-4	Orthopyroxene	52.37	1.29	0.08	0.00	23.71	0.54	20.97	0.00	0.00	0.69	0.00	0.00	0.00	0.00	0.00	99.65
14SAB-L010A01-spot2-3	Orthopyroxene	52.36	1.32	0.10	0.01	23.52	0.55	20.98	0.00	0.00	0.61	0.00	0.00	0.00	0.00	0.00	99.45
14SAB-L057B02-spot2-2	Orthopyroxene	53.70	1.04	0.03	0.08	22.12	0.56	22.07	0.02	0.00	0.53	0.02	0.00	0.00	0.00	0.00	100.18
14SAB-L057B02-spot1-2	Orthopyroxene	53.09	0.96	0.02	0.05	21.99	0.60	22.26	0.04	0.00	0.56	0.00	0.00	0.00	0.00	0.00	99.58
14SAB-L057B02-spot2-3	Orthopyroxene	53.48	0.99	0.02	0.08	21.96	0.56	22.29	0.03	0.00	0.49	0.00	0.00	0.00	0.00	0.00	99.91
14SAB-L057B02-spot1-3	Orthopyroxene	53.31	0.95	0.04	0.03	21.75	0.62	22.36	0.05	0.00	0.49	0.04	0.00	0.00	0.00	0.00	99.63
14SAB-L005A01-spot1-2	Orthopyroxene	52.56	2.80	0.15	0.06	18.32	0.46	24.70	0.07	0.00	0.52	0.00	0.01	0.00	0.00	0.00	99.64
14SAB-L005A01-spot2-2	Orthopyroxene	52.49	3.12	0.17	0.06	18.25	0.43	24.74	0.04	0.00	0.61	0.00	0.00	0.00	0.00	0.00	99.91
14SAB-H080A01-spot1-3	Orthopyroxene	53.66	2.17	0.23	0.06	18.48	0.43	25.26	0.05	0.00	0.40	0.00	0.00	0.00	0.00	0.00	100.73
14SAB-H080A01-spot2-4	Orthopyroxene	53.21	2.28	0.09	0.07	18.22	0.39	24.98	0.05	0.00	0.41	0.00	0.00	0.00	0.00	0.00	99.69
14SAB-H080A01-spot1-8	Orthopyroxene	53.83	2.18	0.11	0.06	18.09	0.40	24.80	0.05	0.00	0.57	0.00	0.00	0.00	0.00	0.00	100.11
14SAB-H080A01-spot1-4	Orthopyroxene	53.58	2.29	0.12	0.04	18.04	0.40	25.04	0.07	0.00	0.49	0.00	0.00	0.00	0.00	0.00	100.07
14SAB-L004A01-spot1-1	Orthopyroxene	53.81	2.78	0.12	0.10	16.67	0.30	25.74	0.04	0.00	0.42	0.01	0.00	0.00	0.00	0.00	99.98
14SAB-L004A01-spot1-2	Orthopyroxene	52.92	2.61	0.13	0.08	16.47	0.37	25.56	0.08	0.00	0.52	0.01	0.00	0.00	0.00	0.00	98.75
14SAB-L004A01-spot2-4	Orthopyroxene	54.66	0.90	0.03	0.05	16.95	0.39	26.34	0.07	0.00	0.24	0.00	0.00	0.00	0.00	0.00	99.63
14SAB-L004A01-spot2-3	Orthopyroxene	53.93	2.50	0.12	0.07	16.57	0.29	25.90	0.07	0.00	0.46	0.00	0.00	0.00	0.00	0.00	99.91
15SAB-L112A01-spot1-3	Orthopyroxene	52.77	3.09	0.00	0.02	16.37	0.37	26.23	0.02	0.00	0.42	0.00	0.00	0.00	0.00	0.00	99.29
15SAB-L112A01-spot2-2	Orthopyroxene	53.08	3.73	0.02	0.01	15.95	0.39	26.00	0.02	0.00	0.46	0.00	0.00	0.00	0.00	0.00	99.68
14SAB-L007A02-spot1-4	Orthopyroxene	53.93	2.54	0.13	0.06	16.26	0.26	26.71	0.05	0.00	0.34	0.00	0.00	0.00	0.00	0.00	100.28
14SAB-L007A02-spot2-2	Orthopyroxene	54.61	2.27	0.07	0.07	16.12	0.31	26.86	0.04	0.00	0.17	0.00	0.00	0.00	0.00	0.00	100.52
14SAB-L007A02-spot2-3	Orthopyroxene	54.00	2.31	0.10	0.04	16.04	0.32	26.82	0.07	0.00	0.23	0.00	0.00	0.00	0.00	0.00	99.93
L012A02-spot2-2	Orthopyroxene	54.19	2.25	0.02	0.01	16.02	0.31	26.90	0.00	0.00	0.33	0.00	0.01	0.00	0.00	0.00	100.05
14SAB-L007A02-spot1-3	Orthopyroxene	53.65	2.59	0.13	0.08	15.90	0.38	26.80	0.05	0.00	0.41	0.00	0.00	0.00	0.00	0.00	99.99
L012A02-spot2-1	Orthopyroxene	54.78	1.67	0.01	0.03	15.85	0.33	27.35	0.00	0.00	0.35	0.03	0.00	0.00	0.00	0.00	100.41
15SAB-D196A01-spot3-5	Orthopyroxene	53.50	3.28	0.03	0.01	15.10	0.39	26.90	0.02	0.00	0.50	0.01	0.00	0.00	0.00	0.00	99.74
15SAB-D196A01-spot1-1	Orthopyroxene	53.43	3.30	0.02	0.00	15.20	0.41	27.15	0.00	0.00	0.52	0.00	0.00	0.00	0.00	0.00	100.05
14SAB-L001A01-spot1-2	Orthopyroxene	53.59	2.72	0.14	0.15	15.00	0.47	26.96	0.09	0.00	0.51	0.00	0.00	0.00	0.00	0.00	99.63
14SAB-L001A01-spot1-3	Orthopyroxene	53.66	2.44	0.09	0.11	15.07	0.41	27.10	0.09	0.00	0.38	0.00	0.00	0.00	0.00	0.00	99.35
14SAB-L001A01-spot2-1	Orthopyroxene	53.89	2.72	0.13	0.10	14.76	0.41	26.86	0.07	0.00	0.50	0.00	0.00	0.00	0.00	0.00	99.43
14SAB-L009A01-area1-2	Orthopyroxene	54.15	3.01	0.19	0.10	14.78	0.32	27.24	0.04	0.00	0.40	0.01	0.00	0.00	0.00	0.00	100.23
14SAB-L009A01-area1-1	Orthopyroxene	54.56	2.97	0.14	0.09	14.81	0.31	27.49	0.02	0.00	0.31	0.02	0.00	0.00	0.00	0.00	100.70
14SAB-L009A01-area2-2	Orthopyroxene	54.09	2.88	0.13	0.14	14.72	0.32	27.46	0.04	0.00	0.29	0.00	0.00	0.00	0.00	0.00	100.06
14SAB-L006B01-spot1-2	Orthopyroxene	54.43	2.58	0.17	0.05	14.63	0.31	27.39	0.04	0.00	0.45	0.03	0.00	0.00	0.00	0.00	100.07
14SAB-L009A01-area2-3	Orthopyroxene	54.04	3.00	0.14	0.12	14.63	0.31	27.41	0.04	0.00	0.39	0.00	0.00	0.00	0.00	0.00	100.09
14SAB-L009A01-area2-4	Orthopyroxene	53.79	2.95	0.14	0.13	14.38	0.32	27.66	0.06	0.00	0.40	0.00	0.00	0.00	0.00	0.00	99.83
L076-A02-spot1-3	Orthopyroxene	54.16	2.87	0.06	0.02	14.14	0.32	27.87	0.00	0.00	0.35	0.00	0.00	0.00	0.05	0.00	99.86
15SAB-L076A01 area 2-3	Orthopyroxene	54.53	2.96	0.04	0.03	14.09	0.33	28.00	0.03	0.00	0.38	0.00	0.00	0.00	0.00	0.00	100.39
15SAB-L076A01 area 2-4	Orthopyroxene	53.57	3.06	0.02	0.00	13.87	0.32	27.83	0.03	0.00	0.37	0.00	0.00	0.00	0.00	0.00	99.06
14SAB-S130A01-spot2-5	Orthopyroxene	54.40	2.91	0.12	0.07	13.42	0.35	28.54	0.05	0.00	0.32	0.00	0.00	0.00	0.00	0.00	100.18
14SAB-S130A01-spot2-6	Orthopyroxene	54.42	2.90	0.13	0.07	13.20	0.35	28.27	0.04	0.00	0.34	0.00	0.00	0.00	0.00	0.00	99.72
14SAB-S130A01-spot2-4	Orthopyroxene	54.26	0.00	0.00	0.00	13.27	0.34	28.66	0.03	0.00	0.34	0.00	0.00	0.00	0.00	0.00	96.89
L014-A01-spot2-5	Orthopyroxene	54.27	2.75	0.03	0.07	13.19	0.32	28.65	0.00	0.00	0.32	0.00	0.00	0.00	0.00	0.00	99.59
L077-A01-spot2-1	Orthopyroxene	54.67	2.77	0.06	0.04	13.14	0.33	28.75	0.03	0.00	0.29	0.00	0.00	0.00	0.00	0.00	100.06
L077-A01-spot3-3	Orthopyroxene	55.11	3.02	0.02	0.01	13.03	0.32	28.85	0.00	0.00	0.36	0.00	0.00	0.00	0.00	0.00	100.72
L077-A01-spot3-2	Orthopyroxene	55.15	2.80	0.03	0.00	13.07	0.34	29.15	0.00	0.00	0.34	0.00	0.01	0.00	0.00	0.00	100.89
L014-A01-spot2-6	Orthopyroxene	54.99	2.43	0.00	0.04	12.98	0.32	29.17	0.00	0.00	0.20	0.00	0.00	0.00	0.01	0.00	100.14
L079-A01-spot1-3	Orthopyroxene	54.30	3.07	0.01	0.03	12.58	0.29	29.17	0.00	0.00	0.38	0.01	0.01	0.00	0.00	0.00	99.86
L079-A01-spot2-1	Orthopyroxene	54.53	3.17	0.05	0.04	12.61	0.30	29.41	0.00	0.00	0.35	0.01	0.01	0.00	0.00	0.01	100.50
14SAB-L002A01-spot2-2	Orthopyroxene	55.37	1.72	0.08	0.15	12.60	0.30	29.53	0.06	0.00	0.50	0.00	0.00	0.00	0.00	0.00	100.31
14SAB-L012A01-spot2-1	Orthopyroxene	55.44	3.16	0.03	0.04	12.31	0.25	30.17	0.03	0.00	0.24	0.00	0.00	0.00	0.00	0.00	101.67

Orthopyroxene - Atoms Per Formula Unit																		
SAMPLE	Mineral	Si	Al	Ti	Cr	Fe	Mn	Mg	Ni	Zn	Ca	Na	K	Ba	Σ cations	%En		
15SAB-M005A01-spot1-1	Orthopyroxene	1.97	0.04	0.00	0.00	1.07	0.05	0.86	0.00	0.00	0.03	0.00	0.00	0.00	4.01	45		
15SAB-L056B01-spot1-2	Orthopyroxene	1.98	0.03	0.00	0.00	1.04	0.02	0.88	0.00	0.00	0.04	0.00	0.00	0.00	4.00	46		
14SAB-L011A02-spot2-3	Orthopyroxene	1.98	0.04	0.00	0.00	1.02	0.02	0.90	0.00	0.00	0.03	0.00	0.00	0.00	4.00	47		
14SAB-L005A02-spot1-2	Orthopyroxene	1.99	0.03	0.00	0.00	0.93	0.02	1.00	0.00	0.00	0.02	0.00	0.00	0.00	3.99	52		
14SAB-L001B01-spot2-5	Orthopyroxene	1.97	0.05	0.00	0.00	0.87	0.02	1.05	0.00	0.00	0.03	0.00	0.00	0.00	4.00	55		
14SAB-L001B01-spot2-4	Orthopyroxene	1.99	0.04	0.00	0.00	0.86	0.02	1.05	0.00	0.00	0.02	0.00	0.00	0.00	3.99	55		
14SAB-L003B01-spot2-2	Orthopyroxene	1.98	0.04	0.00	0.00	0.78	0.02	1.15	0.00	0.00	0.03	0.00	0.00	0.00	4.00	60		
14SAB-L003B01-spot2-3	Orthopyroxene	1.98	0.03	0.00	0.00	0.78	0.03	1.16	0.00	0.00	0.02	0.00	0.00	0.00	4.00	60		
14SAB-L010A01-spot1-2	Orthopyroxene	1.99	0.04	0.00	0.00	0.76	0.02	1.17	0.00	0.00	0.03	0.00	0.00	0.00	3.99	61		
14SAB-L010A01-spot2-4	Orthopyroxene	1.97	0.06	0.00	0.00	0.75	0.02	1.18	0.00	0.00	0.03	0.00	0.00	0.00	4.00	61		
14SAB-L010A01-spot2-3	Orthopyroxene	1.97	0.06	0.00	0.00	0.74	0.02	1.18	0.00	0.00	0.02	0.00	0.00	0.00	4.00	61		
14SAB-L057B02-spot2-2	Orthopyroxene	1.99	0.05	0.00	0.00	0.69	0.02	1.22	0.00	0.00	0.02	0.00	0.00	0.00	3.99	64		
14SAB-L057B02-spot1-2	Orthopyroxene	1.98	0.04	0.00	0.00	0.69	0.02	1.24	0.00	0.00	0.02	0.00	0.00	0.00	4.00	64		
14SAB-L057B02-spot2-3	Orthopyroxene	1.99	0.04	0.00	0.00	0.68	0.02	1.23	0.00	0.00	0.02	0.00	0.00	0.00	3.99	64		
14SAB-L057B02-spot1-3	Orthopyroxene	1.99	0.04	0.00	0.00	0.68	0.02	1.24	0.00	0.00	0.02	0.00	0.00	0.00	3.99	65		
14SAB-L005A01-spot1-2	Orthopyroxene	1.93	0.12	0.00	0.00	0.56	0.01	1.35	0.00	0.00	0.02	0.00	0.00	0.00	4.01	71		
14SAB-L005A01-spot2-2	Orthopyroxene	1.92	0.13	0.00	0.00	0.56	0.01	1.35	0.00	0.00	0.02	0.00	0.00	0.00	4.01	71		
14SAB-H080A01-spot1-3	Orthopyroxene	1.95	0.09	0.01	0.00	0.56	0.01	1.37	0.00	0.00	0.02	0.00	0.00	0.00	4.00	71		
14SAB-H080A01-spot2-4	Orthopyroxene	1.95	0.10	0.00	0.00	0.56	0.01	1.36	0.00	0.00	0.02	0.00	0.00	0.00	4.00	71		
14SAB-H080A01-spot1-8	Orthopyroxene	1.96	0.09	0.00	0.00	0.55	0.01	1.35	0.00	0.00	0.02	0.00	0.00	0.00	3.99	71		
14SAB-H080A01-spot1-4	Orthopyroxene	1.95	0.10	0.00	0.00	0.55	0.01	1.36	0.00	0.00	0.02	0.00	0.00	0.00	4.00	71		
14SAB-L004A01-spot1-1	Orthopyroxene	1.95	0.12	0.00	0.00	0.50	0.01	1.39	0.00	0.00	0.02	0.00	0.00	0.00	3.99	73		
14SAB-L004A01-spot1-2	Orthopyroxene	1.94	0.11	0.00	0.00	0.51	0.01	1.40	0.00	0.00	0.02	0.00	0.00	0.00	4.00	73		
14SAB-L004A01-spot2-4	Orthopyroxene	1.99	0.04	0.00	0.00	0.52	0.01	1.43	0.00	0.00	0.01	0.00	0.00	0.00	3.99	73		
14SAB-L004A01-spot2-3	Orthopyroxene	1.95	0.11	0.00	0.00	0.50	0.01	1.40	0.00	0.00	0.02	0.00	0.00	0.00	3.99	74		
15SAB-L112A01-spot1-3	Orthopyroxene	1.92	0.13	0.00	0.00	0.50	0.01	1.43	0.00	0.00	0.02	0.00	0.00	0.00	4.01	74		
15SAB-L112A01-spot2-2	Orthopyroxene	1.92	0.16	0.00	0.00	0.48	0.01	1.40	0.00	0.00	0.02	0.00	0.00	0.00	4.00	74		
14SAB-L007A02-spot1-4	Orthopyroxene	1.94	0.11	0.00	0.00	0.49	0.01	1.43	0.00	0.00	0.01	0.00	0.00	0.00	4.00	75		
14SAB-L007A02-spot2-2	Orthopyroxene	1.96	0.10	0.00	0.00	0.48	0.01	1.43	0.00	0.00	0.01	0.00	0.00	0.00	3.99	75		
14SAB-L007A02-spot2-3	Orthopyroxene	1.95	0.10	0.00	0.00	0.48	0.01	1.44	0.00	0.00	0.01	0.00	0.00	0.00	4.00	75		
L012A02-spot2-2	Orthopyroxene	1.95	0.10	0.00	0.00	0.48	0.01	1.45	0.00	0.00	0.01	0.00	0.00	0.00	4.00	75		
14SAB-L007A02-spot1-3	Orthopyroxene	1.94	0.11	0.00	0.00	0.48	0.01	1.44	0.00	0.00	0.02	0.00	0.00	0.00	4.00	75		
L012A02-spot2-1	Orthopyroxene	1.96	0.07	0.00	0.00	0.48	0.01	1.46	0.00	0.00	0.01	0.00	0.00	0.00	4.00	75		
15SAB-D196A01-spot3-5	Orthopyroxene	1.93	0.14	0.00	0.00	0.46	0.01	1.45	0.00	0.00	0.02	0.00	0.00	0.00	4.00	76		
15SAB-D196A01-spot1-1	Orthopyroxene	1.92	0.14	0.00	0.00	0.46	0.01	1.46	0.00	0.00	0.02	0.00	0.00	0.00	4.01	76		
14SAB-L001A01-spot1-2	Orthopyroxene	1.94	0.12	0.00	0.00	0.45	0.01	1.45	0.00	0.00	0.02	0.00	0.00	0.00	4.00	76		
14SAB-L001A01-spot1-3	Orthopyroxene	1.94	0.10	0.00	0.00	0.46	0.01	1.46	0.00	0.00	0.01	0.00	0.00	0.00	4.00	76		
14SAB-L001A01-spot2-1	Orthopyroxene	1.95	0.12	0.00	0.00	0.45	0.01	1.45	0.00	0.00	0.02	0.00	0.00	0.00	3.99	76		
14SAB-L009A01-area1-2	Orthopyroxene	1.94	0.13	0.00	0.00	0.44	0.01	1.45	0.00	0.00	0.02	0.00	0.00	0.00	3.99	77		
14SAB-L009A01-area1-1	Orthopyroxene	1.94	0.12	0.00	0.00	0.44	0.01	1.46	0.00	0.00	0.01	0.00	0.00	0.00	3.99	77		
14SAB-L009A01-area2-2	Orthopyroxene	1.94	0.12	0.00	0.00	0.44	0.01	1.47	0.00	0.00	0.01	0.00	0.00	0.00	4.00	77		
14SAB-L006B01-spot1-2	Orthopyroxene	1.95	0.11	0.00	0.00	0.44	0.01	1.46	0.00	0.00	0.02	0.00	0.00	0.00	3.99	77		
14SAB-L009A01-area2-3	Orthopyroxene	1.94	0.13	0.00	0.00	0.44	0.01	1.46	0.00	0.00	0.02	0.00	0.00	0.00	4.00	77		
14SAB-L009A01-area2-4	Orthopyroxene	1.93	0.12	0.00	0.00	0.43	0.01	1.48	0.00	0.00	0.02	0.00	0.00	0.00	4.00	77		
L076-A02-spot1-3	Orthopyroxene	1.94	0.12	0.00	0.00	0.42	0.01	1.49	0.00	0.00	0.01	0.00	0.00	0.00	4.00	78		
15SAB-L076A01 area 2-3	Orthopyroxene	1.94	0.12	0.00	0.00	0.42	0.01	1.49	0.00	0.00	0.01	0.00	0.00	0.00	4.00	78		
15SAB-L076A01 area 2-4	Orthopyroxene	1.93	0.13	0.00	0.00	0.42	0.01	1.50	0.00	0.00	0.01	0.00	0.00	0.00	4.00	78		
14SAB-S130A01-spot2-5	Orthopyroxene	1.94	0.12	0.00	0.00	0.40	0.01	1.51	0.00	0.00	0.01	0.00	0.00	0.00	4.00	79		
14SAB-S130A01-spot2-6	Orthopyroxene	1.94	0.12	0.00	0.00	0.39	0.01	1.50	0.00	0.00	0.01	0.00	0.00	0.00	3.99	79		
14SAB-S130A01-spot2-4	Orthopyroxene	2.00	0.00	0.00	0.00	0.41	0.01	1.57	0.00	0.00	0.01	0.00	0.00	0.00	4.00	79		
L014-A01-spot2-5	Orthopyroxene	1.94	0.12	0.00	0.00	0.39	0.01	1.53	0.00	0.00	0.01	0.00	0.00	0.00	4.00	79		
L077-A01-spot2-1	Orthopyroxene	1.94	0.12	0.00	0.00	0.39	0.01	1.52	0.00	0.00	0.01	0.00	0.00	0.00	4.00	80		
L077-A01-spot3-3	Orthopyroxene	1.94	0.13	0.00	0.00	0.38	0.01	1.52	0.00	0.00	0.01	0.00	0.00	0.00	3.99	80		
L077-A01-spot3-2	Orthopyroxene	1.94	0.12	0.00	0.00	0.39	0.01	1.53	0.00	0.00	0.01	0.00	0.00	0.00	4.00	80		
L014-A01-spot2-6	Orthopyroxene	1.95	0.10	0.00	0.00	0.39	0.01	1.54	0.00	0.00	0.01	0.00	0.00	0.00	4.00	80		
L079-A01-spot1-3	Orthopyroxene	1.93	0.13	0.00	0.00	0.37	0.01	1.55	0.00	0.00	0.01	0.00	0.00	0.00	4.01	81		
L079-A01-spot2-1	Orthopyroxene	1.93	0.13	0.00	0.00	0.37	0.01	1.55	0.00	0.00	0.01	0.00	0.00	0.00	4.01	81		
14SAB-L002A01-spot2-2	Orthopyroxene	1.96	0.07	0.00	0.00	0.37	0.01	1.56	0.00	0.00	0.02	0.00	0.00	0.00	4.00	81		
14SAB-L012A01-spot2-1	Orthopyroxene	1.93	0.13	0.00	0.00	0.36	0.01	1.57	0.00	0.00	0.01	0.00	0.00	0.00	4.00	81		

Amphibole - Electron Microprobe Data																	
SAMPLE	Mineral	SiO2	Al2O3	TiO2	Cr2O3	FeO	MnO	MgO	NiO	ZnO	CaO	Na2O	K2O	BaO	F	Cl	Total
14SAB-L012A01-spot2-2	Amphibole	47.08	12.14	0.25	0.10	7.21	0.14	17.48	0.03	0.00	11.79	1.44	0.00	0.00	0.00	0.00	97.66
14SAB-L012A01-spot2-rep	Amphibole	47.06	12.20	0.23	0.07	7.01	0.14	17.38	0.04	0.00	11.95	1.41	0.21	0.00	0.00	0.00	97.70
L012A01-circle2-4-hb	Amphibole	46.52	12.14	0.25	0.07	7.03	0.12	17.12	0.04	0.01	11.82	1.52	0.22	0.00	0.00	0.01	96.86
L013A02-spot1-1-hb	Amphibole	45.67	14.72	0.27	0.14	6.48	0.12	16.32	0.10	0.09	12.53	1.96	0.41	0.00	0.05	0.06	98.93
L015A01-spot2-3	Amphbole	45.00	10.90	1.25	0.02	13.19	0.24	13.29	0.00	0.00	11.19	1.56	0.34	0.01	0.01	0.02	97.02
L015A01-spot1-1	Amphibole	45.34	10.91	1.18	0.02	13.46	0.20	13.14	0.00	0.00	11.12	1.65	0.30	0.00	0.11	0.02	97.45
L017A02-spot1-2	Amphibole	46.90	8.93	0.56	0.22	11.57	0.27	14.37	0.00	0.00	11.43	1.06	0.82	0.00	0.00	0.00	96.11
L017A02-spot2-2	Amphibole	46.62	9.12	0.68	0.13	11.72	0.21	14.53	0.00	0.00	11.23	1.14	0.76	0.00	0.09	0.06	96.30
L017A02-spot2-3	Amphibole	47.05	9.16	0.55	0.16	11.83	0.26	14.15	0.01	0.00	11.38	1.02	0.81	0.00	0.00	0.00	96.37
L014-A01-spot2-4	Amphibole	45.70	12.65	0.47	0.12	7.25	0.14	16.38	0.00	0.00	12.41	1.27	0.41	0.00	0.03	0.04	96.86
L076-A02-spot1-1	Amphibole	48.09	12.44	0.27	0.04	6.58	0.06	16.84	0.00	0.00	12.18	1.13	0.21	0.00	0.00	0.01	97.83
L076-A02-spot1-2	Amphibole	47.62	12.46	0.31	0.05	6.65	0.09	16.62	0.00	0.00	12.17	1.13	0.20	0.00	0.01	0.01	97.33
L077-A01-spot3-1	Amphibole	48.86	11.27	0.22	0.07	6.25	0.11	17.39	0.00	0.00	11.48	1.03	0.05	0.00	0.01	0.00	96.75
L077-A01-spot3-2	Amphibole	48.57	11.62	0.20	0.08	5.97	0.11	17.40	0.00	0.00	11.64	1.12	0.06	0.00	0.00	0.00	96.77
L079-A01-spot1-2	Amphibole	48.76	11.82	0.32	0.08	5.90	0.10	17.76	0.00	0.00	11.83	1.33	0.10	0.00	0.06	0.01	98.07
L079-A01-spot2-2	Amphibole	48.14	11.59	0.31	0.11	5.85	0.09	17.90	0.00	0.00	11.81	1.21	0.10	0.00	0.05	0.00	97.16
L080-A02-spot1-1	Amphibole	49.96	7.98	0.41	0.31	9.66	0.19	15.95	0.00	0.00	11.54	1.05	0.09	0.00	0.06	0.02	97.23
L080-A02-spot1-2	Amphibole	51.57	7.38	0.47	0.30	11.40	0.21	16.61	0.00	0.00	10.27	0.92	0.11	0.00	0.08	0.03	99.35
L080-A02-spot1-3	Amphibole	50.06	7.60	0.43	0.26	9.95	0.24	16.14	0.00	0.00	11.20	0.98	0.11	0.00	0.00	0.03	96.99
L080-A02-spot2-3	Amphibole	50.66	7.83	0.38	0.19	9.85	0.21	16.13	0.00	0.00	11.32	0.95	0.10	0.00	0.07	0.02	97.71
L080-A02-spot2-4	Amphibole	51.09	7.55	0.39	0.18	10.37	0.23	16.02	0.00	0.00	10.74	1.00	0.09	0.00	0.06	0.02	97.75
L078A01-spot1-1	Amphibole	47.64	11.82	0.33	0.25	5.47	0.11	17.34	0.00	0.00	12.04	1.32	0.22	0.00	0.02	0.02	96.58
L078A01-spot1-2	Amphibole	48.28	11.61	0.34	0.24	5.37	0.10	17.47	0.00	0.00	12.01	1.23	0.23	0.00	0.00	0.01	96.89
L012A02-spot2-3	Amphibole	43.23	15.09	0.44	0.15	7.96	0.10	14.92	0.00	0.00	11.76	2.64	0.35	0.00	0.11	0.01	96.76
L012A02-spot2-4	Amphibole	44.25	14.23	0.48	0.24	7.53	0.10	15.44	0.00	0.00	11.89	2.54	0.34	0.00	0.04	0.02	97.11
L012A02-spot1-2	Amphibole	43.97	13.71	1.09	0.26	8.66	0.07	14.47	0.00	0.00	11.79	2.41	0.53	0.00	0.07	0.02	97.06
14SAB-L002A01-spot1-3	Amphibole	48.28	9.74	0.73	0.70	5.76	0.12	17.73	0.00	0.00	12.44	1.01	0.21	0.00	0.00	0.02	96.75
14SAB-L002A01-spot1-4	Amphibole	49.04	9.85	0.69	0.78	5.69	0.11	17.46	0.00	0.00	12.46	0.95	0.16	0.00	0.00	0.00	97.19
14SAB-L002A01-spot2-1	Amphibole	53.03	5.99	0.40	0.25	5.08	0.11	19.53	0.00	0.00	12.37	0.61	0.07	0.00	0.00	0.00	97.43
14SAB-L002A02-spot1-3	Amphibole	45.48	10.82	1.53	0.33	9.04	0.10	15.40	0.00	0.00	11.96	1.80	0.97	0.00	0.00	0.02	97.46
14SAB-L002A02-spot2-1	Amphibole	44.70	11.03	1.54	0.33	9.70	0.14	15.23	0.00	0.00	11.80	1.76	0.97	0.00	0.00	0.01	97.21
14SAB-L006A01-spot1-2	Amphibole	40.16	11.95	2.78	0.01	21.74	0.25	6.07	0.00	0.00	11.21	1.99	1.60	0.00	0.00	0.00	97.77
14SAB-L006A01-spot2-4	Amphibole	40.05	11.97	2.85	0.00	22.10	0.26	6.15	0.01	0.00	11.20	1.91	1.56	0.00	0.11	0.01	98.18
14SAB-L006B01-spot1-1	Amphibole	46.98	11.33	1.13	0.24	6.46	0.09	15.82	0.06	0.00	12.52	0.77	0.20	0.00	0.12	0.00	95.73
14SAB-L006B01-spot2-4	Amphibole	55.08	4.05	0.41	0.10	4.87	0.08	19.49	0.00	0.00	12.89	0.12	0.00	0.00	0.00	0.00	97.10
14SAB-L010A01-spot1-3	Amphibole	42.52	11.98	2.30	0.04	12.84	0.14	11.83	0.00	0.00	11.50	1.98	1.34	0.00	0.44	0.00	96.91
14SAB-L010A01-spot2-5	Amphibole	43.13	11.91	1.99	0.01	12.98	0.14	12.19	0.00	0.00	11.78	1.91	1.39	0.00	0.65	0.00	98.08
15SAB-L077B01- area 1-1	Amphibole	44.93	14.26	0.33	0.19	8.42	0.18	14.52	0.00	0.00	12.39	1.26	0.22	0.00	0.00	0.00	96.72
15SAB-L077B01-area 1-2	Amphibole	45.01	15.05	0.31	0.15	8.32	0.18	14.29	0.00	0.00	12.54	1.30	0.26	0.00	0.00	0.00	97.40
15SAB-L077B01-area2-2	Amphibole	46.60	12.70	0.39	0.19	8.79	0.13	14.78	0.00	0.00	12.50	1.11	0.19	0.00	0.01	0.01	97.40
15SAB-L080A01-area1-1	Amphibole	49.91	7.87	0.85	0.11	9.69	0.22	15.87	0.00	0.00	11.76	1.00	0.13	0.00	0.01	0.05	97.47
15SAB-L080A01-area1-2	Amphibole	49.70	7.76	0.90	0.15	9.81	0.21	15.82	0.00	0.00	11.57	1.09	0.13	0.00	0.00	0.02	97.17
15SAB-L080A01-area2-2	Amphibole	50.12	7.74	0.78	0.12	9.66	0.22	15.83	0.00	0.00	11.60	1.00	0.12	0.00	0.06	0.03	97.29
15SAB-L080A01-area2-3	Amphibole	50.28	7.70	0.78	0.11	9.29	0.20	16.10	0.00	0.00	11.59	0.89	0.14	0.00	0.02	0.04	97.15
14SAB-H080B01-area1-1	Amphibole	47.54	11.06	0.83	0.65	5.67	0.10	17.34	0.00	0.00	12.57	1.30	0.16	0.00	0.02	0.02	97.26
14SAB-H080B01 area 2-3	Amphibole	48.05	10.58	0.77	0.57	5.52	0.10	17.65	0.00	0.00	12.78	1.27	0.16	0.00	0.05	0.00	97.49
14SAB-H080B01 area 2-4	Amphibole	47.70	10.73	0.77	0.61	5.58	0.07	17.50	0.00	0.00	12.77	1.19	0.16	0.00	0.03	0.00	97.10
15SAB-L076A01 area 1-1	Amphibole	47.57	12.06	0.22	0.08	6.62	0.12	16.59	0.00	0.00	12.55	1.10	0.21	0.00	0.01	0.00	97.13

Amphibole - Electron Microprobe Data																	
SAMPLE	Mineral	SiO2	Al2O3	TiO2	Cr2O3	FeO	MnO	MgO	NiO	ZnO	CaO	Na2O	K2O	BaO	F	Cl	Total
15SAB-L076A01 area 1-2	Amphibole	47.94	11.96	0.21	0.10	6.57	0.13	16.51	0.00	0.00	12.56	1.11	0.20	0.00	0.02	0.00	97.31
15SAB-L076A01 area 2-5	Amphibole	47.80	12.12	0.24	0.07	6.51	0.08	16.89	0.00	0.00	12.52	1.09	0.21	0.00	0.00	0.00	97.53
14SAB-R010A01-spot1-4	Amphibole	49.81	9.86	0.91	0.68	4.15	0.14	18.60	0.00	0.00	12.89	0.97	0.21	0.00	0.00	0.00	98.22
14SAB-R010A01-spot2-1	Amphibole	49.10	10.16	0.97	0.78	4.15	0.11	18.13	0.00	0.00	12.87	0.96	0.26	0.00	0.13	0.00	97.62
14SAB-L007A02-spot1-5	Amphibole	44.86	13.56	1.17	0.33	7.88	0.10	15.31	0.00	0.00	12.06	1.69	0.49	0.00	0.04	0.00	97.50
14SAB-L007A02-spot1-1	Amphibole	44.65	13.40	1.14	0.32	7.79	0.10	15.19	0.00	0.00	11.90	1.64	0.49	0.00	0.18	0.00	96.80
14SAB-L016A01-spot1-1	Amphibole	47.09	9.21	0.51	0.24	10.94	0.23	14.52	0.00	0.00	11.53	1.17	0.65	0.00	0.07	0.00	96.17
14SAB-L016A01-spot2-3	Amphibole	47.41	9.15	0.44	0.19	11.16	0.26	14.63	0.00	0.00	11.73	1.07	0.66	0.00	0.17	0.00	96.86
14SAB-L003A01-spot1-1	Amphibole	48.12	10.75	0.84	0.52	6.18	0.11	17.38	0.00	0.00	12.21	1.24	0.18	0.00	0.03	0.00	97.57
14SAB-L003A01-spot2-3	Amphibole	47.14	11.14	0.80	0.61	6.06	0.10	17.33	0.00	0.00	12.16	1.18	0.16	0.00	0.09	0.00	96.76
14SAB-L008A02-spot1-1	Amphibole	41.99	12.99	2.03	0.25	13.73	0.10	11.10	0.00	0.00	11.83	2.21	0.77	0.00	0.68	0.00	97.67
14SAB-L008A02-spot1-2	Amphibole	42.31	12.68	1.97	0.24	13.59	0.11	11.38	0.00	0.00	11.84	2.09	0.74	0.00	0.76	0.00	97.69
14SAB-L008A02-spot2-3	Amphibole	42.21	12.69	1.66	0.27	13.62	0.13	11.32	0.00	0.00	11.66	2.21	0.72	0.00	0.76	0.00	97.25
14SAB-L057B02-spot1-1	Amphibole	48.92	8.74	0.57	0.26	9.75	0.17	15.38	0.00	0.00	11.44	1.19	0.53	0.00	0.15	0.00	97.11
14SAB-L057B02-spot2-4	Amphibole	47.71	9.72	0.52	0.40	10.24	0.18	14.87	0.00	0.00	11.20	1.31	0.54	0.00	0.15	0.00	96.84
15SAB-L119A01-spot1-5	Amphibole	45.26	13.18	0.90	0.60	6.72	0.07	16.04	0.00	0.00	12.40	1.70	0.22	0.00	0.00	0.00	97.09
15SAB-L119A01-spot1-4	Amphibole	44.84	13.31	0.90	0.54	6.86	0.08	16.03	0.00	0.00	12.42	1.70	0.22	0.00	0.07	0.00	96.95
15SAB-L119A01-spot2-1	Amphibole	45.94	12.53	0.81	0.64	6.31	0.06	16.27	0.00	0.00	12.21	1.63	0.20	0.00	0.01	0.00	96.61
14SAB-H080A01-spot1-5	Amphibole	45.84	11.94	0.95	0.24	8.89	0.11	15.13	0.00	0.00	12.16	1.17	0.26	0.00	0.09	0.00	96.78
14SAB-H080A01-spot1-6	Amphibole	46.72	11.17	0.86	0.16	8.84	0.12	15.14	0.00	0.00	12.27	1.23	0.23	0.00	0.21	0.00	96.95
14SAB-H080A01-spot1-7	Amphibole	46.48	12.04	0.92	0.22	8.72	0.11	14.88	0.00	0.00	12.30	1.24	0.27	0.00	0.08	0.00	97.25
14SAB-H080A01-spot1-7	Amphibole	45.98	11.91	1.10	0.25	8.94	0.12	14.88	0.00	0.00	12.27	1.23	0.30	0.00	0.10	0.00	97.08
15SAB-L077B01-area1- 1	Amphibole	44.92	15.80	0.23	0.08	8.12	0.16	14.30	0.00	0.00	11.98	1.43	0.25	0.00	0.05	0.00	97.31
15SAB-L077B01-area1- 2	Amphibole	45.39	14.82	0.24	0.12	8.10	0.16	14.30	0.00	0.00	11.93	1.30	0.25	0.00	0.02	0.00	96.61
15SAB-L077B01-area2-3	Amphibole	46.80	12.34	0.38	0.19	8.66	0.12	14.72	0.00	0.00	12.25	1.10	0.17	0.00	0.05	0.00	96.77
14SAB-L009A01-area1-3	Amphibole	44.35	13.41	1.23	0.45	7.33	0.11	15.35	0.00	0.00	11.89	2.04	0.63	0.00	0.11	0.00	96.90
14SAB-L009A01-area1-4	Amphibole	44.28	13.95	1.25	0.51	7.07	0.07	15.16	0.00	0.00	11.95	1.98	0.64	0.00	0.01	0.00	96.88
14SAB-L009A01-area2-1	Amphibole	44.18	13.82	1.18	0.50	7.19	0.14	15.41	0.00	0.00	11.49	1.96	0.62	0.00	0.08	0.00	96.57
14SAB-L004A01-spot1-5	Amphibole	42.37	14.54	1.96	0.30	8.35	0.10	14.06	0.00	0.00	11.96	2.52	0.41	0.00	0.27	0.00	96.84
14SAB-L004A01-spot1-6	Amphibole	42.75	14.37	2.00	0.30	8.15	0.08	13.96	0.00	0.00	11.79	2.62	0.42	0.00	0.16	0.00	96.61
14SAB-L004A01-spot2-1	Amphibole	42.06	14.33	2.09	0.30	8.22	0.10	14.20	0.00	0.00	11.79	2.49	0.39	0.00	0.17	0.00	96.12
14SAB-S130A01-spot1-4	Amphibole	47.69	11.09	1.21	0.26	6.32	0.10	16.62	0.00	0.00	12.59	0.62	0.27	0.00	0.08	0.00	96.85
14SAB-S130A01-spot2-1	Amphibole	48.07	10.50	1.09	0.21	6.27	0.10	16.72	0.00	0.00	12.64	0.57	0.23	0.00	0.00	0.00	96.41
14SAB-S130A01-spot2-2	Amphibole	47.82	11.04	1.25	0.18	6.40	0.11	16.64	0.00	0.00	12.44	0.60	0.22	0.00	0.00	0.00	96.71
14SAB-L001B01-spot1-5	Amphibole	42.64	11.83	2.37	0.42	15.32	0.15	10.33	0.00	0.00	11.46	1.88	1.03	0.00	0.42	0.00	97.44
14SAB-L001B01-spot2-1	Amphibole	43.00	11.89	2.43	0.24	13.94	0.11	11.43	0.00	0.00	11.52	2.02	0.89	0.00	0.46	0.00	97.92
14SAB-L003B01-spot1-5	Amphibole	45.48	10.37	1.81	0.24	12.73	0.16	12.76	0.00	0.00	11.32	1.60	0.97	0.00	0.03	0.00	97.48
14SAB-L003B01-spot2-1	Amphibole	45.23	10.47	1.87	0.22	12.08	0.14	12.79	0.00	0.00	11.40	1.55	0.95	0.00	0.17	0.00	96.86
15SAB-L056B01-spot1-4	Amphibole	43.25	11.08	1.99	0.14	17.32	0.21	9.34	0.00	0.00	11.46	1.38	1.32	0.00	0.21	0.02	97.71
15SAB-L056B01-spot3	Amphibole	43.08	11.09	1.92	0.10	17.17	0.18	9.22	0.00	0.00	11.37	1.29	1.29	0.00	0.15	0.00	96.86
14SAB-L001A01-spot1-1	Amphibole	48.74	9.80	0.98	0.39	6.40	0.07	16.57	0.00	0.00	12.61	0.69	0.07	0.00	0.00	0.00	96.31
14SAB-L001A01-spot2-2	Amphibole	46.45	11.34	1.16	0.46	7.20	0.14	15.92	0.00	0.00	12.50	0.86	0.08	0.00	0.00	0.00	96.11
15SAB-L112A01-spot1-4	Amphibole	41.86	15.43	0.23	0.06	9.23	0.08	14.47	0.00	0.00	11.97	2.85	0.35	0.00	0.31	0.00	96.83
15SAB-L112A01-spot2-1	Amphibole	43.35	14.73	0.07	0.02	8.40	0.12	15.20	0.00	0.00	11.97	2.80	0.27	0.00	0.35	0.00	97.26
15SAB-D196A01-spot2-4rep	Amphibole	42.69	14.64	0.47	0.05	8.82	0.11	14.57	0.00	0.00	12.25	2.43	0.47	0.00	0.08	0.30	96.90
15SAB-D196A01-spot3-3	Amphibole	42.43	14.68	0.46	0.06	8.90	0.08	14.80	0.00	0.00	11.98	2.50	0.46	0.00	0.00	0.28	96.63
15SAB-M005A01-spot1-3	Amphibole	41.25	11.15	2.17	0.00	19.20	0.38	8.58	0.00	0.00	11.27	1.72	1.21	0.00	0.13	0.09	97.15
15SAB-M005A01-spot2-2	Amphibole	41.91	11.08	2.06	0.00	19.69	0.35	8.57	0.00	0.00	11.50	1.67	1.17	0.00	0.23	0.09	98.31
14SAB-L011A02-spot1-2	Amphibole	41.46	11.98	2.80	0.09	18.07	0.20	9.04	0.00	0.00	11.23	2.39	0.67	0.00	0.26	0.00	98.18
14SAB-L011A02-spot1-3	Amphibole	41.66	11.58	2.38	0.09	18.21	0.19	9.04	0.00	0.00	11.14	2.34	0.63	0.00	0.16	0.00	97.43
14SAB-L010C01-spot1-2	Amphibole	40.71	14.83	2.30	0.50	9.42	0.11	12.91	0.03	0.00	12.29	1.72	2.05	0.00	0.42	0.69	97.98
14SAB-L010C01-spot2-3	Amphibole	40.59	14.64	2.29	0.47	9.37	0.13	12.87	0.00	0.00	12.34	1.90	1.88	0.00	0.54	0.70	97.72
14SAB-L005A01-spot1-1	Amphibole	45.71	12.32	1.21	0.20	9.03	0.09	14.76	0.00	0.00	12.36	0.96	0.27	0.00	0.05	0.01	96.99
14SAB-L005A01-spot2-3	Amphibole	44.53	13.05	1.41	0.22	8.93	0.08	14.36	0.00	0.00	12.55	0.95	0.30	0.00	0.00	0.00	96.38
14SAB-L005A02-spot1-1	Amphibole	44.80	9.99	1.27	0.34	14.57	0.13	11.58	0.00	0.00	11.91	1.48	0.89	0.00	0.18	0.08	97.21
14SAB-L005A02-spot2-3	Amphibole	42.02	11.90	2.44	0.17	15.66	0.16	10.17	0.00	0.00	11.37	1.88	1.29	0.00	0.20	0.08	97.33

Amphibole - Atoms Per Formula Unit																
SAMPLE	Mineral	Si	Al	Ti	Cr	Fe	Mn	Mg	Ni	Zn	Ca	Na	K	Ba	Σ cations	Mg/(Mg+Fe)
14SAB-L012A01-spot2-2	Amphibole	6.67	2.03	0.03	0.01	0.85	0.02	3.69	0.00	0.00	1.79	0.39	0.00	0.00	15.48	0.81
14SAB-L012A01-spot2-rep	Amphibole	6.67	2.04	0.02	0.01	0.83	0.02	3.67	0.01	0.00	1.81	0.39	0.04	0.00	15.50	0.82
L012A01-circle2-4-hb	Amphibole	6.65	2.05	0.03	0.01	0.84	0.01	3.65	0.00	0.00	1.81	0.42	0.04	0.00	15.52	0.81
L013A02-spot1-1-hb	Amphibole	6.42	2.44	0.03	0.02	0.76	0.01	3.42	0.01	0.01	1.89	0.53	0.07	0.00	15.62	0.82
L015A01-spot2-3	Amphibole	6.64	1.90	0.14	0.00	1.63	0.03	2.92	0.00	0.00	1.77	0.45	0.06	0.00	15.53	0.64
L015A01-spot1-1	Amphibole	6.66	1.89	0.13	0.00	1.65	0.03	2.88	0.00	0.00	1.75	0.47	0.06	0.00	15.52	0.64
L017A02-spot1-2	Amphibole	6.92	1.55	0.06	0.03	1.43	0.03	3.16	0.00	0.00	1.81	0.30	0.15	0.00	15.45	0.69
L017A02-spot2-2	Amphibole	6.88	1.59	0.08	0.02	1.45	0.03	3.20	0.00	0.00	1.78	0.33	0.14	0.00	15.48	0.69
L017A02-spot2-3	Amphibole	6.93	1.59	0.06	0.02	1.46	0.03	3.11	0.00	0.00	1.80	0.29	0.15	0.00	15.43	0.68
L014-A01-spot2-4	Amphibole	6.57	2.14	0.05	0.01	0.87	0.02	3.51	0.00	0.00	1.91	0.35	0.08	0.00	15.52	0.80
L076-A02-spot1-1	Amphibole	6.76	2.06	0.03	0.00	0.77	0.01	3.53	0.00	0.00	1.83	0.31	0.04	0.00	15.35	0.82
L076-A02-spot1-2	Amphibole	6.74	2.08	0.03	0.01	0.79	0.01	3.51	0.00	0.00	1.85	0.31	0.04	0.00	15.36	0.82
L077-A01-spot3-1	Amphibole	6.91	1.88	0.02	0.01	0.74	0.01	3.67	0.00	0.00	1.74	0.28	0.01	0.00	15.27	0.83
L077-A01-spot3-2	Amphibole	6.87	1.94	0.02	0.01	0.71	0.01	3.67	0.00	0.00	1.76	0.31	0.01	0.00	15.30	0.84
L079-A01-spot1-2	Amphibole	6.82	1.95	0.03	0.01	0.69	0.01	3.70	0.00	0.00	1.77	0.36	0.02	0.00	15.36	0.84
L079-A01-spot2-2	Amphibole	6.80	1.93	0.03	0.01	0.69	0.01	3.77	0.00	0.00	1.79	0.33	0.02	0.00	15.37	0.85
L080-A02-spot1-1	Amphibole	7.16	1.35	0.04	0.04	1.16	0.02	3.41	0.00	0.00	1.77	0.29	0.02	0.00	15.26	0.75
L080-A02-spot1-2	Amphibole	7.25	1.22	0.05	0.03	1.34	0.03	3.48	0.00	0.00	1.55	0.25	0.02	0.00	15.21	0.72
L080-A02-spot1-3	Amphibole	7.19	1.29	0.05	0.03	1.20	0.03	3.46	0.00	0.00	1.72	0.27	0.02	0.00	15.25	0.74
L080-A02-spot2-3	Amphibole	7.21	1.31	0.04	0.02	1.17	0.02	3.42	0.00	0.00	1.73	0.26	0.02	0.00	15.22	0.74
L080-A02-spot2-4	Amphibole	7.27	1.27	0.04	0.02	1.23	0.03	3.40	0.00	0.00	1.64	0.28	0.02	0.00	15.19	0.73
L078A01-spot1-1	Amphibole	6.77	1.98	0.03	0.03	0.65	0.01	3.67	0.00	0.00	1.83	0.36	0.04	0.00	15.39	0.85
L078A01-spot1-2	Amphibole	6.82	1.93	0.04	0.03	0.63	0.01	3.68	0.00	0.00	1.82	0.34	0.04	0.00	15.35	0.85
L012A02-spot2-3	Amphibole	6.28	2.58	0.05	0.02	0.97	0.01	3.23	0.00	0.00	1.83	0.74	0.06	0.00	15.78	0.77
L012A02-spot2-4	Amphibole	6.38	2.42	0.05	0.03	0.91	0.01	3.32	0.00	0.00	1.84	0.71	0.06	0.00	15.73	0.79
L012A02-spot1-2	Amphibole	6.39	2.35	0.12	0.03	1.05	0.01	3.13	0.00	0.00	1.84	0.68	0.10	0.00	15.69	0.75
14SAB-L002A01-spot1-3	Amphibole	6.88	1.63	0.08	0.08	0.69	0.01	3.76	0.00	0.00	1.90	0.28	0.04	0.00	15.35	0.85
14SAB-L002A01-spot1-4	Amphibole	6.93	1.64	0.07	0.09	0.67	0.01	3.68	0.00	0.00	1.89	0.26	0.03	0.00	15.27	0.85
14SAB-L002A01-spot2-1	Amphibole	7.40	0.98	0.04	0.03	0.59	0.01	4.06	0.00	0.00	1.85	0.17	0.01	0.00	15.14	0.87
14SAB-L002A02-spot1-3	Amphibole	6.60	1.85	0.17	0.04	1.10	0.01	3.33	0.00	0.00	1.86	0.51	0.18	0.00	15.64	0.75
14SAB-L002A02-spot2-1	Amphibole	6.53	1.90	0.17	0.04	1.18	0.02	3.32	0.00	0.00	1.85	0.50	0.18	0.00	15.68	0.74
14SAB-L006A01-spot1-2	Amphibole	6.24	2.19	0.33	0.00	2.82	0.03	1.41	0.00	0.00	1.87	0.60	0.32	0.00	15.80	0.33
14SAB-L006A01-spot2-4	Amphibole	6.21	2.19	0.33	0.00	2.87	0.03	1.42	0.00	0.00	1.86	0.58	0.31	0.00	15.80	0.33
14SAB-L006B01-spot1-1	Amphibole	6.78	1.93	0.12	0.03	0.78	0.01	3.40	0.01	0.00	1.94	0.22	0.04	0.00	15.25	0.81
14SAB-L006B01-spot2-4	Amphibole	7.67	0.66	0.04	0.01	0.57	0.01	4.05	0.00	0.00	1.92	0.03	0.00	0.00	14.97	0.88
14SAB-L010A01-spot1-3	Amphibole	6.38	2.12	0.26	0.00	1.61	0.02	2.65	0.00	0.00	1.85	0.57	0.26	0.00	15.72	0.62
14SAB-L010A01-spot2-5	Amphibole	6.41	2.08	0.22	0.00	1.61	0.02	2.70	0.00	0.00	1.88	0.55	0.26	0.00	15.73	0.63
15SAB-L077B01- area 1-1	Amphibole	6.49	2.43	0.04	0.02	1.02	0.02	3.13	0.00	0.00	1.92	0.35	0.04	0.00	15.45	0.75
15SAB-L077B01-area 1-2	Amphibole	6.45	2.54	0.03	0.02	1.00	0.02	3.05	0.00	0.00	1.92	0.36	0.05	0.00	15.44	0.75
15SAB-L077B01-area2-2	Amphibole	6.68	2.14	0.04	0.02	1.05	0.02	3.16	0.00	0.00	1.92	0.31	0.04	0.00	15.37	0.75
15SAB-L080A01-area1-1	Amphibole	7.14	1.33	0.09	0.01	1.16	0.03	3.38	0.00	0.00	1.80	0.28	0.02	0.00	15.25	0.74
15SAB-L080A01-area1-2	Amphibole	7.14	1.31	0.10	0.02	1.18	0.03	3.39	0.00	0.00	1.78	0.30	0.02	0.00	15.26	0.74
15SAB-L080A01-area2-2	Amphibole	7.18	1.31	0.08	0.01	1.16	0.03	3.38	0.00	0.00	1.78	0.28	0.02	0.00	15.23	0.74
15SAB-L080A01-area2-3	Amphibole	7.19	1.30	0.08	0.01	1.11	0.02	3.43	0.00	0.00	1.78	0.25	0.03	0.00	15.20	0.76
14SAB-H080B01-area1-1	Amphibole	6.74	1.85	0.09	0.07	0.67	0.01	3.67	0.00	0.00	1.91	0.36	0.03	0.00	15.40	0.84
14SAB-H080B01 area 2-3	Amphibole	6.79	1.76	0.08	0.06	0.65	0.01	3.72	0.00	0.00	1.94	0.35	0.03	0.00	15.40	0.85
14SAB-H080B01 area 2-4	Amphibole	6.77	1.80	0.08	0.07	0.66	0.01	3.70	0.00	0.00	1.94	0.33	0.03	0.00	15.39	0.85
15SAB-L076A01 area 1-1	Amphibole	6.76	2.02	0.02	0.01	0.79	0.01	3.51	0.00	0.00	1.91	0.30	0.04	0.00	15.38	0.82

Amphibole - Atoms Per Formula Unit																
SAMPLE	Mineral	Si	Al	Ti	Cr	Fe	Mn	Mg	Ni	Zn	Ca	Na	K	Ba	Σ cations	Mg/(Mg+Fe)
15SAB-L076A01 area 1-2	Amphibole	6.79	2.00	0.02	0.01	0.78	0.02	3.49	0.00	0.00	1.91	0.30	0.04	0.00	15.35	0.82
15SAB-L076A01 area 2-5	Amphibole	6.75	2.02	0.03	0.01	0.77	0.01	3.56	0.00	0.00	1.90	0.30	0.04	0.00	15.37	0.82
14SAB-R010A01-spot1-4	Amphibole	6.93	1.62	0.10	0.07	0.48	0.02	3.85	0.00	0.00	1.92	0.26	0.04	0.00	15.28	0.89
14SAB-R010A01-spot2-1	Amphibole	6.88	1.68	0.10	0.09	0.49	0.01	3.79	0.00	0.00	1.93	0.26	0.05	0.00	15.28	0.89
14SAB-L007A02-spot1-5	Amphibole	6.44	2.29	0.13	0.04	0.95	0.01	3.28	0.00	0.00	1.85	0.47	0.09	0.00	15.55	0.78
14SAB-L007A02-spot1-1	Amphibole	6.46	2.28	0.12	0.04	0.94	0.01	3.28	0.00	0.00	1.84	0.46	0.09	0.00	15.53	0.78
14SAB-L016A01-spot1-1	Amphibole	6.93	1.60	0.06	0.03	1.35	0.03	3.18	0.00	0.00	1.82	0.33	0.12	0.00	15.44	0.70
14SAB-L016A01-spot2-3	Amphibole	6.93	1.58	0.05	0.02	1.37	0.03	3.19	0.00	0.00	1.84	0.30	0.12	0.00	15.43	0.70
14SAB-L003A01-spot1-1	Amphibole	6.80	1.79	0.09	0.06	0.73	0.01	3.66	0.00	0.00	1.85	0.34	0.03	0.00	15.37	0.83
14SAB-L003A01-spot2-3	Amphibole	6.73	1.87	0.09	0.07	0.72	0.01	3.69	0.00	0.00	1.86	0.33	0.03	0.00	15.39	0.84
14SAB-L008A02-spot1-1	Amphibole	6.28	2.29	0.23	0.03	1.72	0.01	2.48	0.00	0.00	1.90	0.64	0.15	0.00	15.72	0.59
14SAB-L008A02-spot1-2	Amphibole	6.32	2.23	0.22	0.03	1.70	0.01	2.54	0.00	0.00	1.90	0.60	0.14	0.00	15.70	0.60
14SAB-L008A02-spot2-3	Amphibole	6.34	2.25	0.19	0.03	1.71	0.02	2.53	0.00	0.00	1.88	0.64	0.14	0.00	15.73	0.60
14SAB-L057B02-spot1-1	Amphibole	7.06	1.49	0.06	0.03	1.18	0.02	3.31	0.00	0.00	1.77	0.33	0.10	0.00	15.34	0.74
14SAB-L057B02-spot2-4	Amphibole	6.93	1.66	0.06	0.05	1.24	0.02	3.22	0.00	0.00	1.74	0.37	0.10	0.00	15.39	0.72
15SAB-L119A01-spot1-5	Amphibole	6.48	2.23	0.10	0.07	0.80	0.01	3.43	0.00	0.00	1.90	0.47	0.04	0.00	15.53	0.81
15SAB-L119A01-spot1-4	Amphibole	6.45	2.26	0.10	0.06	0.82	0.01	3.44	0.00	0.00	1.91	0.47	0.04	0.00	15.56	0.81
15SAB-L119A01-spot2-1	Amphibole	6.59	2.12	0.09	0.07	0.76	0.01	3.48	0.00	0.00	1.88	0.45	0.04	0.00	15.47	0.82
14SAB-H080A01-spot1-5	Amphibole	6.63	2.04	0.10	0.03	1.08	0.01	3.26	0.00	0.00	1.89	0.33	0.05	0.00	15.42	0.75
14SAB-H080A01-spot1-6	Amphibole	6.75	1.90	0.09	0.02	1.07	0.02	3.26	0.00	0.00	1.90	0.35	0.04	0.00	15.39	0.75
14SAB-H080A01-spot1-7	Amphibole	6.68	2.04	0.10	0.02	1.05	0.01	3.19	0.00	0.00	1.89	0.35	0.05	0.00	15.39	0.75
14SAB-H080A01-spot1-7	Amphibole	6.64	2.03	0.12	0.03	1.08	0.02	3.20	0.00	0.00	1.90	0.34	0.05	0.00	15.41	0.75
15SAB-L077B01-area1-1	Amphibole	6.42	2.66	0.02	0.01	0.97	0.02	3.05	0.00	0.00	1.83	0.40	0.05	0.00	15.44	0.76
15SAB-L077B01-area1-2	Amphibole	6.53	2.51	0.03	0.01	0.97	0.02	3.07	0.00	0.00	1.84	0.36	0.05	0.00	15.39	0.76
15SAB-L077B01-area2-3	Amphibole	6.74	2.09	0.04	0.02	1.04	0.01	3.16	0.00	0.00	1.89	0.31	0.03	0.00	15.34	0.75
14SAB-L009A01-area1-3	Amphibole	6.41	2.29	0.13	0.05	0.89	0.01	3.31	0.00	0.00	1.84	0.57	0.12	0.00	15.63	0.79
14SAB-L009A01-area1-4	Amphibole	6.39	2.37	0.14	0.06	0.85	0.01	3.26	0.00	0.00	1.85	0.55	0.12	0.00	15.60	0.79
14SAB-L009A01-area2-1	Amphibole	6.40	2.36	0.13	0.06	0.87	0.02	3.32	0.00	0.00	1.78	0.55	0.12	0.00	15.60	0.79
14SAB-L004A01-spot1-5	Amphibole	6.20	2.51	0.22	0.03	1.02	0.01	3.06	0.00	0.00	1.87	0.71	0.08	0.00	15.71	0.75
14SAB-L004A01-spot1-6	Amphibole	6.25	2.47	0.22	0.03	1.00	0.01	3.04	0.00	0.00	1.85	0.74	0.08	0.00	15.69	0.75
14SAB-L004A01-spot2-1	Amphibole	6.19	2.48	0.23	0.03	1.01	0.01	3.11	0.00	0.00	1.86	0.71	0.07	0.00	15.71	0.75
14SAB-S130A01-spot1-4	Amphibole	6.79	1.86	0.13	0.03	0.75	0.01	3.53	0.00	0.00	1.92	0.17	0.05	0.00	15.24	0.82
14SAB-S130A01-spot2-1	Amphibole	6.86	1.77	0.12	0.02	0.75	0.01	3.56	0.00	0.00	1.93	0.16	0.04	0.00	15.22	0.83
14SAB-S130A01-spot2-2	Amphibole	6.81	1.85	0.13	0.02	0.76	0.01	3.53	0.00	0.00	1.90	0.17	0.04	0.00	15.23	0.82
14SAB-L001B01-spot1-5	Amphibole	6.39	2.09	0.27	0.05	1.92	0.02	2.31	0.00	0.00	1.84	0.55	0.20	0.00	15.64	0.55
14SAB-L001B01-spot2-1	Amphibole	6.40	2.08	0.27	0.03	1.73	0.01	2.53	0.00	0.00	1.84	0.58	0.17	0.00	15.65	0.59
14SAB-L003B01-spot1-5	Amphibole	6.69	1.80	0.20	0.03	1.57	0.02	2.80	0.00	0.00	1.78	0.46	0.18	0.00	15.52	0.64
14SAB-L003B01-spot2-1	Amphibole	6.68	1.82	0.21	0.03	1.49	0.02	2.82	0.00	0.00	1.80	0.45	0.18	0.00	15.50	0.65
15SAB-L056B01-spot1-4	Amphibole	6.53	1.97	0.23	0.02	2.19	0.03	2.10	0.00	0.00	1.85	0.40	0.25	0.00	15.58	0.49
15SAB-L056B01-spot3	Amphibole	6.55	1.99	0.22	0.01	2.18	0.02	2.09	0.00	0.00	1.85	0.38	0.25	0.00	15.55	0.49
14SAB-L001A01-spot1-1	Amphibole	6.96	1.65	0.11	0.04	0.76	0.01	3.53	0.00	0.00	1.93	0.19	0.01	0.00	15.19	0.82
14SAB-L001A01-spot2-2	Amphibole	6.70	1.93	0.13	0.05	0.87	0.02	3.43	0.00	0.00	1.93	0.24	0.01	0.00	15.31	0.80
15SAB-L112A01-spot1-4	Amphibole	6.15	2.67	0.02	0.01	1.13	0.01	3.17	0.00	0.00	1.88	0.81	0.06	0.00	15.93	0.74
15SAB-L112A01-spot2-1	Amphibole	6.30	2.52	0.01	0.00	1.02	0.01	3.29	0.00	0.00	1.86	0.79	0.05	0.00	15.85	0.76
15SAB-D196A01-spot2-4rep	Amphibole	6.25	2.53	0.05	0.01	1.08	0.01	3.18	0.00	0.00	1.92	0.69	0.09	0.00	15.82	0.75
15SAB-D196A01-spot3-3	Amphibole	6.23	2.54	0.05	0.01	1.09	0.01	3.24	0.00	0.00	1.88	0.71	0.09	0.00	15.85	0.75
15SAB-M005A01-spot1-3	Amphibole	6.36	2.03	0.25	0.00	2.48	0.05	1.97	0.00	0.00	1.86	0.52	0.24	0.00	15.75	0.44
15SAB-M005A01-spot2-2	Amphibole	6.40	1.99	0.24	0.00	2.51	0.04	1.95	0.00	0.00	1.88	0.50	0.23	0.00	15.73	0.44
14SAB-L011A02-spot1-2	Amphibole	6.27	2.14	0.32	0.01	2.29	0.03	2.04	0.00	0.00	1.82	0.70	0.13	0.00	15.75	0.47
14SAB-L011A02-spot1-3	Amphibole	6.35	2.08	0.27	0.01	2.32	0.02	2.05	0.00	0.00	1.82	0.69	0.12	0.00	15.74	0.47
14SAB-L010C01-spot1-2	Amphibole	6.04	2.59	0.26	0.06	1.17	0.01	2.85	0.00	0.00	1.95	0.50	0.39	0.00	15.82	0.71
14SAB-L010C01-spot2-3	Amphibole	6.04	2.57	0.26	0.06	1.17	0.02	2.86	0.00	0.00	1.97	0.55	0.36	0.00	15.84	0.71
14SAB-L005A01-spot1-1	Amphibole	6.60	2.10	0.13	0.02	1.09	0.01	3.18	0.00	0.00	1.91	0.27	0.05	0.00	15.37	0.74
14SAB-L005A01-spot2-3	Amphibole	6.48	2.24	0.15	0.03	1.09	0.01	3.12	0.00	0.00	1.96	0.27	0.06	0.00	15.39	0.74
14SAB-L005A02-spot1-1	Amphibole	6.70	1.76	0.14	0.04	1.82	0.02	2.58	0.00	0.00	1.91	0.43	0.17	0.00	15.56	0.59
14SAB-L005A02-spot2-3	Amphibole	6.35	2.12	0.28	0.02	1.98	0.02	2.29	0.00	0.00	1.84	0.55	0.25	0.00	15.70	0.54

Plagioclase - Electron Microprobe Data																	
SAMPLE	Mineral	SiO2	Al2O3	TiO2	Cr2O3	FeO	MnO	MgO	NiO	ZnO	CaO	Na2O	K2O	BaO	F	Cl	Total
14SAB-L001B01-spot1-3	Plagioclase	49.63	32.80	0.00	0.00	0.09	0.00	0.00	0.00	0.00	16.12	2.45	0.05	0.00	0.00	0.00	101.13
14SAB-L001B01-spot1-4	Plagioclase	49.14	32.82	0.00	0.00	0.09	0.00	0.00	0.00	0.00	16.49	2.25	0.03	0.00	0.00	0.00	100.82
14SAB-L001B01-spot2-6	Plagioclase	46.72	33.94	0.00	0.00	0.06	0.00	0.00	0.00	0.00	17.21	1.65	0.03	0.00	0.00	0.00	99.60
14SAB-L001B01-spot2-7	Plagioclase	46.62	34.05	0.00	0.00	0.11	0.00	0.00	0.00	0.00	17.46	1.51	0.01	0.00	0.00	0.00	99.76
14SAB-L003B01-spot1-1	Plagioclase	51.75	31.11	0.00	0.00	0.02	0.00	0.00	0.00	0.00	14.20	3.43	0.09	0.00	0.00	0.00	100.61
14SAB-L003B01-spot1-2	Plagioclase	51.49	31.23	0.00	0.00	0.06	0.00	0.00	0.00	0.00	13.91	3.60	0.09	0.00	0.00	0.00	100.39
14SAB-L003B01-spot2-6	Plagioclase	51.39	31.42	0.00	0.00	0.10	0.00	0.00	0.00	0.00	14.31	3.44	0.10	0.00	0.00	0.00	100.76
14SAB-L003B01-spot2-7	Plagioclase	51.58	31.47	0.00	0.00	0.08	0.00	0.00	0.00	0.00	14.10	3.63	0.11	0.00	0.00	0.00	100.97
14SAB-L005A02-spot1-4	Plagioclase	50.10	32.07	0.00	0.00	0.42	0.00	0.00	0.00	0.00	15.25	3.03	0.04	0.00	0.00	0.00	100.92
14SAB-L005A02-spot2-1	Plagioclase	46.06	34.88	0.00	0.00	0.15	0.00	0.00	0.00	0.00	18.24	1.07	0.06	0.00	0.00	0.00	100.46
14SAB-L005A02-spot2-2	Plagioclase	45.47	35.24	0.00	0.00	0.12	0.00	0.00	0.00	0.00	18.95	0.87	0.01	0.00	0.00	0.00	100.66
14SAB-L006A01-spot1-4	Plagioclase	58.97	26.25	0.00	0.00	0.07	0.00	0.00	0.00	0.00	8.16	6.79	0.33	0.00	0.00	0.00	100.57
14SAB-L006A01-spot1-5	Plagioclase	58.31	25.62	0.00	0.00	0.98	0.00	0.00	0.00	0.00	7.67	6.69	0.23	0.00	0.00	0.00	99.51
14SAB-L006A01-spot2-1	Plagioclase	58.84	26.39	0.00	0.00	0.08	0.00	0.00	0.00	0.00	8.04	6.99	0.27	0.00	0.00	0.00	100.60
14SAB-L006A01-spot2-2	Plagioclase	58.74	26.43	0.00	0.00	0.08	0.00	0.00	0.00	0.00	8.04	6.89	0.24	0.00	0.00	0.00	100.42
14SAB-L010A01-spot1-4	Plagioclase	53.61	29.49	0.00	0.00	0.09	0.00	0.00	0.00	0.00	12.30	4.49	0.13	0.00	0.00	0.00	100.10
14SAB-L010A01-spot2-1	Plagioclase	52.88	29.68	0.00	0.00	0.30	0.00	0.00	0.00	0.00	12.49	4.25	0.15	0.00	0.00	0.00	99.75
14SAB-L011A02-spot1-4	Plagioclase	53.66	30.03	0.00	0.00	0.14	0.00	0.00	0.00	0.00	12.51	4.59	0.09	0.00	0.00	0.00	101.02
14SAB-L011A02-spot2-1	Plagioclase	52.38	30.82	0.00	0.00	0.15	0.00	0.00	0.00	0.00	13.74	4.02	0.08	0.00	0.00	0.00	101.19
14SAB-L012A02-spot1-1	Plagioclase	44.96	35.08	0.00	0.00	0.06	0.00	0.00	0.00	0.00	18.83	0.73	0.00	0.00	0.00	0.00	99.65
14SAB-L012A02-spot2-5	Plagioclase	44.66	35.44	0.00	0.00	0.08	0.00	0.00	0.00	0.00	19.22	0.39	0.01	0.00	0.00	0.00	99.80
14SAB-L013A02-spot1-7	Plagioclase	44.10	35.85	0.00	0.00	0.07	0.00	0.00	0.00	0.00	19.59	0.19	0.00	0.00	0.00	0.00	99.81
14SAB-L013A02-spot1-8	Plagioclase	43.96	36.07	0.00	0.00	0.11	0.00	0.00	0.00	0.00	19.70	0.15	0.00	0.00	0.00	0.00	99.99
14SAB-L015A01-spot1-2	Plagioclase	46.27	34.60	0.00	0.00	0.09	0.00	0.00	0.00	0.00	18.02	1.16	0.01	0.05	0.00	0.00	100.19
14SAB-L015A01-spot2-1	Plagioclase	46.20	34.81	0.00	0.00	0.19	0.00	0.00	0.00	0.00	18.20	1.08	0.01	0.00	0.00	0.00	100.49
14SAB-L015A01-spot2-2	Plagioclase	46.64	34.86	0.00	0.00	0.19	0.00	0.00	0.00	0.00	18.20	0.99	0.01	0.00	0.00	0.00	100.89
14SAB-L016A01-spot1-3	Plagioclase	48.31	32.68	0.00	0.00	0.14	0.00	0.00	0.00	0.00	15.77	2.46	0.07	0.00	0.00	0.00	99.42
14SAB-L016A01-spot1-4	Plagioclase	48.84	33.01	0.00	0.00	0.13	0.00	0.00	0.00	0.00	15.78	2.48	0.07	0.00	0.00	0.00	100.31
14SAB-L016A01-spot2-1	Plagioclase	49.06	33.36	0.00	0.00	0.16	0.00	0.00	0.00	0.00	16.02	2.41	0.06	0.00	0.00	0.00	101.07
14SAB-L016A01-spot2-2	Plagioclase	49.65	32.78	0.00	0.00	0.21	0.00	0.00	0.00	0.00	15.61	2.59	0.08	0.00	0.00	0.00	100.92
14SAB-L017A02-spot1-1	Plagioclase	51.77	30.65	0.00	0.00	0.11	0.00	0.00	0.00	0.00	13.48	3.67	0.16	0.08	0.00	0.00	99.92
14SAB-L017A02-spot2-4	Plagioclase	55.24	28.65	0.00	0.00	0.08	0.00	0.00	0.00	0.00	10.63	5.04	0.25	0.00	0.00	0.00	99.90
14SAB-L017A02-spot2-5	Plagioclase	55.14	29.27	0.00	0.00	0.14	0.00	0.00	0.00	0.00	11.30	4.73	0.27	0.07	0.00	0.00	100.91
14SAB-L057B02-spot1-4	Plagioclase	58.61	26.55	0.00	0.00	0.03	0.00	0.00	0.00	0.00	8.25	7.03	0.10	0.00	0.00	0.00	100.57
14SAB-L057B02-spot1-5	Plagioclase	58.46	26.76	0.00	0.00	0.18	0.00	0.00	0.00	0.00	8.30	7.02	0.07	0.00	0.00	0.00	100.80
14SAB-L057B02-spot2-1	Plagioclase	58.60	26.62	0.00	0.00	0.16	0.00	0.00	0.00	0.00	8.25	6.90	0.06	0.00	0.00	0.00	100.59
14SAB-L080-A02-spot1-4	Plagioclase	48.01	33.19	0.00	0.00	0.03	0.00	0.00	0.00	0.00	16.50	2.19	0.02	0.00	0.00	0.00	99.95
14SAB-L080-A02-spot2-1	Plagioclase	47.20	34.44	0.00	0.00	0.05	0.00	0.00	0.00	0.00	17.47	1.54	0.01	0.00	0.00	0.00	100.71
14SAB-L080-A02-spot2-2	Plagioclase	47.12	34.17	0.00	0.00	0.03	0.00	0.00	0.00	0.00	17.54	1.61	0.03	0.00	0.00	0.00	100.50
15SAB-D196A01-spot2-6	Plagioclase	44.30	36.07	0.00	0.00	0.11	0.00	0.00	0.00	0.00	19.69	0.38	0.01	0.00	0.00	0.00	100.57
15SAB-D196A01-spot3-1	Plagioclase	43.94	35.83	0.00	0.00	0.12	0.00	0.00	0.00	0.00	19.56	0.32	0.00	0.00	0.00	0.00	99.78
15SAB-L056B01-spot1-5	Plagioclase	50.63	31.43	0.00	0.00	0.13	0.00	0.00	0.00	0.00	14.63	3.17	0.09	0.00	0.00	0.00	100.07
15SAB-L056B01-spot1-6	Plagioclase	54.43	28.82	0.00	0.00	0.05	0.00	0.00	0.00	0.00	11.58	4.86	0.17	0.00	0.00	0.00	99.91
15SAB-L056B01-spot2-1	Plagioclase	50.34	31.34	0.00	0.00	0.15	0.00	0.00	0.00	0.00	14.71	3.07	0.09	0.00	0.00	0.00	99.71
15SAB-L056B01-spot2-2	Plagioclase	50.45	31.53	0.00	0.00	0.13	0.00	0.00	0.00	0.00	14.84	3.10	0.08	0.00	0.00	0.00	100.13
15SAB-L077B01-area1-3	Plagioclase	43.30	36.32	0.00	0.00	0.01	0.00	0.00	0.00	0.00	20.67	0.11	0.00	0.00	0.00	0.00	100.40
15SAB-L077B01-area1-3	Plagioclase	43.86	35.74	0.00	0.00	0.02	0.00	0.00	0.00	0.00	19.80	0.07	0.00	0.00	0.00	0.00	99.49
15SAB-L077B01-area1-4	Plagioclase	43.78	36.23	0.00	0.00	0.05	0.00	0.00	0.00	0.00	19.94	0.10	0.00	0.00	0.00	0.00	100.12
15SAB-L077B01-area2-1	Plagioclase	43.72	36.30	0.00	0.00	0.03	0.00	0.00	0.00	0.00	20.61	0.14	0.00	0.00	0.00	0.00	100.80
15SAB-L077B01-area2-1	Plagioclase	43.62	36.22	0.00	0.00	0.01	0.00	0.00	0.00	0.00	20.07	0.14	0.00	0.00	0.00	0.00	100.07
15SAB-L077B01-area2-2	Plagioclase	43.88	36.04	0.00	0.00	0.00	0.00	0.00	0.00	0.00	19.97	0.16	0.00	0.00	0.00	0.00	100.05
15SAB-L077B01-area1-4	Plagioclase	43.35	36.36	0.00	0.00	0.05	0.00	0.00	0.00	0.00	20.82	0.07	0.00	0.00	0.00	0.00	100.65
15SAB-L080A01-area1-3	Plagioclase	46.24	34.44	0.00	0.00	0.03	0.00	0.00	0.00	0.00	18.73	1.35	0.02	0.00	0.00	0.00	100.81
15SAB-L080A01-area1-4	Plagioclase	46.47	34.04	0.00	0.00	0.09	0.00	0.00	0.00	0.00	18.63	1.39	0.02	0.00	0.00	0.00	100.65
15SAB-L080A01-area2-1	Plagioclase	46.40	34.39	0.00	0.00	0.05	0.00	0.00	0.00	0.00	18.51	1.37	0.01	0.00	0.00	0.00	100.72
15SAB-L112A01-spot1-1	Plagioclase	44.71	35.01	0.00	0.00	0.06	0.00	0.00	0.00	0.00	19.02	0.67	0.01	0.00	0.00	0.00	99.48
15SAB-L112A01-spot1-2	Plagioclase	44.52	34.89	0.00	0.00	0.09	0.00	0.00	0.00	0.00	19.19	0.68	0.01	0.00	0.00	0.00	99.38
15SAB-L112A01-spot2-4	Plagioclase	44.72	35.91	0.00	0.00	0.02	0.00	0.00	0.00	0.00	19.26	0.68	0.00	0.00	0.00	0.00	100.61
15SAB-M005A01-spot1-4	Plagioclase	54.83	28.51	0.00	0.00	0.12	0.00	0.00	0.00	0.00	11.01	5.11	0.22	0.00	0.00	0.00	99.80
15SAB-M005A01-spot2-1	Plagioclase	54.47	28.08	0.00	0.00	0.11	0.00	0.00	0.00	0.00	10.79	5.23	0.24	0.00	0.00	0.00	98.93

Plagioclase - Atoms Per Formula Unit																	
SAMPLE	Mineral	Si	Al	Ti	Cr	Fe	Mn	Mg	Ni	Zn	Ca	Na	K	Ba	Σ cations	An%	
14SAB-L001B01-spot1-3	Plagioclase	2.24	1.75	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.78	0.21	0.00	0.00	4.99	78	
14SAB-L001B01-spot1-4	Plagioclase	2.23	1.76	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.80	0.20	0.00	0.00	4.99	80	
14SAB-L001B01-spot2-6	Plagioclase	2.15	1.84	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.85	0.15	0.00	0.00	5.00	85	
14SAB-L001B01-spot2-7	Plagioclase	2.15	1.85	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.86	0.13	0.00	0.00	5.00	86	
14SAB-L003B01-spot1-1	Plagioclase	2.34	1.66	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.69	0.30	0.01	0.00	4.99	69	
14SAB-L003B01-spot1-2	Plagioclase	2.33	1.67	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.67	0.32	0.01	0.00	5.00	68	
14SAB-L003B01-spot2-6	Plagioclase	2.32	1.67	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.69	0.30	0.01	0.00	5.00	69	
14SAB-L003B01-spot2-7	Plagioclase	2.32	1.67	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.68	0.32	0.01	0.00	5.00	68	
14SAB-L005A02-spot1-4	Plagioclase	2.27	1.71	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.74	0.27	0.00	0.00	5.01	73	
14SAB-L005A02-spot2-1	Plagioclase	2.11	1.88	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.90	0.10	0.00	0.00	5.00	90	
14SAB-L005A02-spot2-2	Plagioclase	2.08	1.90	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.93	0.08	0.00	0.00	5.00	92	
14SAB-L006A01-spot1-4	Plagioclase	2.62	1.38	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.39	0.59	0.02	0.00	4.99	39	
14SAB-L006A01-spot1-5	Plagioclase	2.63	1.36	0.00	0.00	0.04	0.00	0.00	0.00	0.00	0.37	0.58	0.01	0.00	4.99	38	
14SAB-L006A01-spot2-1	Plagioclase	2.62	1.38	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.38	0.60	0.02	0.00	5.00	38	
14SAB-L006A01-spot2-2	Plagioclase	2.61	1.39	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.38	0.59	0.01	0.00	5.00	39	
14SAB-L010A01-spot1-4	Plagioclase	2.42	1.57	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.60	0.39	0.01	0.00	4.99	60	
14SAB-L010A01-spot2-1	Plagioclase	2.40	1.59	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.61	0.37	0.01	0.00	4.99	61	
14SAB-L011A02-spot1-4	Plagioclase	2.41	1.59	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.60	0.40	0.01	0.00	5.00	60	
14SAB-L011A02-spot2-1	Plagioclase	2.35	1.63	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.66	0.35	0.00	0.00	5.01	65	
14SAB-L012A02-spot1-1	Plagioclase	2.08	1.91	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.93	0.07	0.00	0.00	5.00	93	
14SAB-L012A02-spot2-5	Plagioclase	2.06	1.93	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.95	0.04	0.00	0.00	4.99	96	
14SAB-L013A02-spot1-7	Plagioclase	2.04	1.96	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.97	0.02	0.00	0.00	4.99	98	
14SAB-L013A02-spot1-8	Plagioclase	2.03	1.97	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.98	0.01	0.00	0.00	4.99	99	
14SAB-L015A01-spot1-2	Plagioclase	2.12	1.87	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.89	0.10	0.00	0.00	4.99	90	
14SAB-L015A01-spot2-1	Plagioclase	2.12	1.88	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.89	0.10	0.00	0.00	4.99	90	
14SAB-L015A01-spot2-2	Plagioclase	2.13	1.87	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.89	0.09	0.00	0.00	4.98	91	
14SAB-L016A01-spot1-3	Plagioclase	2.22	1.77	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.78	0.22	0.00	0.00	5.00	78	
14SAB-L016A01-spot1-4	Plagioclase	2.23	1.77	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.77	0.22	0.00	0.00	5.00	78	
14SAB-L016A01-spot2-1	Plagioclase	2.22	1.78	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.78	0.21	0.00	0.00	5.00	78	
14SAB-L016A01-spot2-2	Plagioclase	2.25	1.75	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.76	0.23	0.00	0.00	4.99	77	
14SAB-L017A02-spot1-1	Plagioclase	2.35	1.64	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.66	0.32	0.01	0.00	4.99	66	
14SAB-L017A02-spot2-4	Plagioclase	2.49	1.52	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.51	0.44	0.01	0.00	4.98	53	
14SAB-L017A02-spot2-5	Plagioclase	2.46	1.54	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.54	0.41	0.02	0.00	4.98	56	
14SAB-L057B02-spot1-4	Plagioclase	2.61	1.39	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.39	0.61	0.01	0.00	5.00	39	
14SAB-L057B02-spot1-5	Plagioclase	2.60	1.40	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.40	0.60	0.00	0.00	5.01	39	
14SAB-L057B02-spot2-1	Plagioclase	2.60	1.39	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.39	0.59	0.00	0.00	5.00	40	
14SAB-L080-A02-spot1-4	Plagioclase	2.20	1.79	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.81	0.19	0.00	0.00	5.00	81	
14SAB-L080-A02-spot2-1	Plagioclase	2.15	1.85	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.85	0.14	0.00	0.00	4.99	86	
14SAB-L080-A02-spot2-2	Plagioclase	2.15	1.84	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.86	0.14	0.00	0.00	5.00	86	
15SAB-D196A01-spot2-6	Plagioclase	2.04	1.96	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.97	0.03	0.00	0.00	5.00	97	
15SAB-D196A01-spot3-1	Plagioclase	2.04	1.96	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.97	0.03	0.00	0.00	5.00	97	
15SAB-L056B01-spot1-5	Plagioclase	2.30	1.69	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.71	0.28	0.01	0.00	4.99	71	
15SAB-L056B01-spot1-6	Plagioclase	2.46	1.53	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.56	0.43	0.01	0.00	4.99	56	
15SAB-L056B01-spot2-1	Plagioclase	2.30	1.69	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.72	0.27	0.01	0.00	4.99	72	
15SAB-L056B01-spot2-2	Plagioclase	2.30	1.69	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.72	0.27	0.00	0.00	5.00	72	
15SAB-L077B01-area1-3	Plagioclase	2.00	1.98	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.02	0.01	0.00	0.00	5.01	99	
15SAB-L077B01-area1-3	Plagioclase	2.04	1.96	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.99	0.01	0.00	0.00	4.99	99	
15SAB-L077B01-area1-4	Plagioclase	2.02	1.97	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.99	0.01	0.00	0.00	4.99	99	
15SAB-L077B01-area2-1	Plagioclase	2.01	1.97	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.02	0.01	0.00	0.00	5.01	99	
15SAB-L077B01-area2-1	Plagioclase	2.02	1.98	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.99	0.01	0.00	0.00	5.00	99	
15SAB-L077B01-area2-2	Plagioclase	2.03	1.96	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.99	0.01	0.00	0.00	5.00	99	
15SAB-L077B01-area1-4	Plagioclase	2.00	1.98	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.03	0.01	0.00	0.00	5.01	99	
15SAB-L080A01-area1-3	Plagioclase	2.12	1.86	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.92	0.12	0.00	0.00	5.02	88	
15SAB-L080A01-area1-4	Plagioclase	2.13	1.84	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.91	0.12	0.00	0.00	5.01	88	
15SAB-L080A01-area2-1	Plagioclase	2.12	1.85	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.91	0.12	0.00	0.00	5.01	88	
15SAB-L112A01-spot1-1	Plagioclase	2.07	1.91	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.95	0.06	0.00	0.00	5.00	94	
15SAB-L112A01-spot1-2	Plagioclase	2.07	1.91	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.96	0.06	0.00	0.00	5.00	94	
15SAB-L112A01-spot2-4	Plagioclase	2.05	1.94	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.95	0.06	0.00	0.00	5.01	94	
15SAB-M005A01-spot1-4	Plagioclase	2.48	1.52	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.53	0.45	0.01	0.00	4.99	54	
15SAB-M005A01-spot2-1	Plagioclase	2.48	1.51	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.53	0.46	0.01	0.00	5.00	53	

Spinel - Electron Microprobe Data																	
SAMPLE	Mineral	SiO ₂	Al ₂ O ₃	TiO ₂	Cr ₂ O ₃	FeO	MnO	MgO	NiO	ZnO	CaO	Na ₂ O	K ₂ O	BaO	F	Cl	Total
14SAB-L003A01-spot1-4	Spinel	0.04	53.27	0.02	8.86	22.78	0.14	12.52	0.30	0.75	0.00	0.00	0.00	0.00	0.00	0.00	98.69
14SAB-L003A01-spot2-1	Spinel	0.04	51.72	0.03	10.30	23.22	0.22	11.91	0.30	0.79	0.01	0.00	0.00	0.00	0.00	0.00	98.54
14SAB-L004A01-spot1-3	Spinel	0.14	57.11	0.08	4.26	24.49	0.14	11.11	0.46	0.67	0.06	0.00	0.00	0.00	0.00	0.00	98.52
14SAB-L006B01-spot1-4	Spinel	0.02	57.76	0.05	3.88	22.15	0.15	12.93	0.33	0.27	0.01	0.00	0.00	0.00	0.00	0.00	97.55
14SAB-L006B01-spot2-2	Spinel	0.03	57.59	0.05	3.68	24.39	0.17	11.53	0.36	0.30	0.02	0.00	0.00	0.00	0.00	0.00	98.11
14SAB-R010A01-spot1-3	Spinel	0.08	48.37	0.05	16.42	19.48	0.20	13.49	0.21	0.58	0.07	0.00	0.00	0.00	0.00	0.00	98.94
14SAB-S130A01-spot1-3	Spinel	0.01	58.54	0.03	3.59	21.15	0.14	14.37	0.31	0.11	0.00	0.00	0.00	0.00	0.00	0.00	98.25
14SAB-S130A01-spot2-3	Spinel	0.05	59.32	0.03	2.93	21.10	0.16	14.42	0.32	0.14	0.00	0.00	0.00	0.00	0.00	0.00	98.48
15SAB-D196A01-spot2-3	Spinel	0.10	60.47	0.01	0.39	22.27	0.15	13.67	0.08	1.06	0.01	0.04	0.00	0.00	0.00	0.00	98.26
15SAB-L076A01 area 1-3	Spinel	0.05	62.09	0.04	0.89	19.45	0.16	15.01	0.35	0.14	0.02	0.00	0.00	0.00	0.00	0.00	98.19
15SAB-L076A01 area 1-4	Spinel	0.07	61.84	0.00	1.18	20.85	0.12	14.88	0.33	0.18	0.01	0.00	0.00	0.00	0.00	0.00	99.47
15SAB-L112A01-spot2-3	Spinel	0.10	61.58	0.00	0.02	23.62	0.16	13.70	0.13	0.51	0.02	0.00	0.00	0.00	0.00	0.00	99.84
15SAB-L119A01-spot1-3	Spinel	0.13	60.10	0.01	2.89	19.66	0.09	14.65	0.36	0.88	0.04	0.00	0.00	0.00	0.00	0.00	98.82
15SAB-L119A01-spot2-2	Spinel	0.12	60.36	0.01	3.26	19.41	0.11	14.65	0.39	0.67	0.03	0.00	0.00	0.00	0.00	0.00	99.01
14SAB-L012A01-circle2-3	Spinel	0.01	60.53	0.00	1.26	20.29	0.16	16.05	0.26	0.21	0.00	0.00	0.00	0.00	0.00	0.00	98.78
14SAB-L013A02-spot1-3	Spinel	0.05	63.00	0.00	1.34	18.95	0.13	16.03	0.16	0.18	0.01	0.01	0.00	0.00	0.00	0.00	99.86
14SAB-L014-A01-spot2-3	Spinel	0.17	59.97	0.02	2.51	21.91	0.12	14.52	0.14	0.37	0.00	0.00	0.00	0.00	0.00	0.00	99.74
14SAB-L076-A02-spot1-5	Spinel	0.09	62.76	0.01	0.43	20.28	0.17	15.22	0.35	0.06	0.00	0.00	0.00	0.00	0.00	0.00	99.37
14SAB-L076-A02-spot2-1	Spinel	0.00	62.67	0.03	0.63	19.85	0.12	15.38	0.37	0.14	0.00	0.00	0.00	0.00	0.00	0.00	99.20
14SAB-L077-A01-spot2-3	Spinel	0.05	61.42	0.02	2.74	19.22	0.14	15.40	0.36	0.11	0.00	0.00	0.00	0.00	0.00	0.00	99.46
14SAB-L078A01-spot1-3	Spinel	0.07	60.44	0.04	3.21	17.61	0.10	15.95	0.30	0.22	0.00	0.00	0.00	0.00	0.00	0.00	97.94
14SAB-L079-A01-spot2-4	Spinel	0.02	63.59	0.01	1.21	17.31	0.13	16.63	0.33	0.06	0.00	0.00	0.00	0.00	0.00	0.00	99.30

Spinel - Atoms Per Formula Unit																		
SAMPLE	Mineral	Si	Al	Ti	Cr	Fe	Mn	Mg	Ni	Zn	Ca	Na	K	Ba	Σ cations	Cr#	Fe3+	Fe2+
14SAB-L003A01-spot1-4	Spinel	0.00	1.75	0.00	0.20	0.53	0.00	0.52	0.01	0.02	0.00	0.00	0.00	0.00	3.03	0.10	0.07	0.46
14SAB-L003A01-spot2-1	Spinel	0.00	1.72	0.00	0.23	0.55	0.01	0.50	0.01	0.02	0.00	0.00	0.00	0.00	3.02	0.12	0.07	0.48
14SAB-L004A01-spot1-3	Spinel	0.00	1.86	0.00	0.09	0.57	0.00	0.46	0.01	0.01	0.00	0.00	0.00	0.00	3.02	0.05	0.04	0.52
14SAB-L006B01-spot1-4	Spinel	0.00	1.87	0.00	0.08	0.51	0.00	0.53	0.01	0.01	0.00	0.00	0.00	0.00	3.02	0.04	0.05	0.46
14SAB-L006B01-spot2-2	Spinel	0.00	1.88	0.00	0.08	0.56	0.00	0.48	0.01	0.01	0.00	0.00	0.00	0.00	3.02	0.04	0.05	0.51
14SAB-R010A01-spot1-3	Spinel	0.00	1.60	0.00	0.36	0.46	0.00	0.56	0.00	0.01	0.00	0.00	0.00	0.00	3.01	0.19	0.04	0.42
14SAB-S130A01-spot1-3	Spinel	0.00	1.87	0.00	0.08	0.48	0.00	0.58	0.01	0.00	0.00	0.00	0.00	0.00	3.02	0.04	0.06	0.42
14SAB-S130A01-spot2-3	Spinel	0.00	1.89	0.00	0.06	0.48	0.00	0.58	0.01	0.00	0.00	0.00	0.00	0.00	3.02	0.03	0.06	0.42
15SAB-D196A01-spot2-3	Spinel	0.00	1.93	0.00	0.01	0.50	0.00	0.55	0.00	0.02	0.00	0.00	0.00	0.00	3.03	0.00	0.07	0.43
15SAB-L076A01 area 1-3	Spinel	0.00	1.95	0.00	0.02	0.43	0.00	0.60	0.01	0.00	0.00	0.00	0.00	0.00	3.01	0.01	0.04	0.40
15SAB-L076A01 area 1-4	Spinel	0.00	1.93	0.00	0.02	0.46	0.00	0.59	0.01	0.00	0.00	0.00	0.00	0.00	3.02	0.01	0.05	0.41
15SAB-L112A01-spot2-3	Spinel	0.00	1.94	0.00	0.00	0.53	0.00	0.55	0.00	0.01	0.00	0.00	0.00	0.00	3.03	0.00	0.08	0.45
15SAB-L119A01-spot1-3	Spinel	0.00	1.90	0.00	0.06	0.44	0.00	0.59	0.01	0.02	0.00	0.00	0.00	0.00	3.02	0.03	0.04	0.40
15SAB-L119A01-spot2-2	Spinel	0.00	1.90	0.00	0.07	0.43	0.00	0.58	0.01	0.01	0.00	0.00	0.00	0.00	3.01	0.04	0.03	0.40
14SAB-L012A01-circle2-3	Spinel	0.00	1.90	0.00	0.03	0.45	0.00	0.64	0.01	0.00	0.00	0.00	0.00	0.00	3.03	0.01	0.09	0.36
14SAB-L013A02-spot1-3	Spinel	0.00	1.94	0.00	0.03	0.41	0.00	0.62	0.00	0.00	0.00	0.00	0.00	0.00	3.02	0.01	0.04	0.37
14SAB-L014-A01-spot2-3	Spinel	0.00	1.89	0.00	0.05	0.49	0.00	0.58	0.00	0.01	0.00	0.00	0.00	0.00	3.03	0.03	0.07	0.42
14SAB-L076-A02-spot1-5	Spinel	0.00	1.95	0.00	0.01	0.45	0.00	0.60	0.01	0.00	0.00	0.00	0.00	0.00	3.02	0.00	0.05	0.40
14SAB-L076-A02-spot2-1	Spinel	0.00	1.95	0.00	0.01	0.44	0.00	0.60	0.01	0.00	0.00	0.00	0.00	0.00	3.02	0.01	0.05	0.39
14SAB-L077-A01-spot2-3	Spinel	0.00	1.91	0.00	0.06	0.42	0.00	0.61	0.01	0.00	0.00	0.00	0.00	0.00	3.01	0.03	0.04	0.39
14SAB-L078A01-spot1-3	Spinel	0.00	1.90	0.00	0.07	0.39	0.00	0.63	0.01	0.00	0.00	0.00	0.00	0.00	3.01	0.03	0.03	0.36
14SAB-L079-A01-spot2-4	Spinel	0.00	1.95	0.00	0.02	0.38	0.00	0.65	0.01	0.00	0.00	0.00	0.00	0.00	3.01	0.01	0.03	0.35

Biotite - Electron Microprobe Data																	
SAMPLE	Mineral	SiO ₂	Al ₂ O ₃	TiO ₂	Cr ₂ O ₃	FeO	MnO	MgO	NiO	ZnO	CaO	Na ₂ O	K ₂ O	BaO	F	Cl	Total
14SAB-L017A02-spot1-3	Biotite	37.32	14.84	2.30	0.23	13.88	0.10	16.03	0.00	0.00	0.01	0.06	9.45	0.92	0.17	0.06	95.37
14SAB-L017A02-spot2-1	Biotite	37.62	15.02	2.26	0.13	13.70	0.05	16.11	0.00	0.00	0.00	0.05	9.44	0.85	0.17	0.07	95.46
14SAB-L016A01-spot1-2	Biotite	37.19	15.24	2.15	0.22	12.51	0.06	16.15	0.00	0.00	0.00	0.03	9.57	0.00	0.29	0.00	93.41
14SAB-L009A01-area1-5	Biotite	39.15	16.57	1.68	0.20	6.96	0.00	20.64	0.00	0.00	0.01	0.56	8.26	0.17	0.14	0.00	94.34

Biotite - Atoms Per Formula Unit																	
SAMPLE	Mineral	Si	Al	Ti	Cr	Fe	Mn	Mg	Ni	Zn	Ca	Na	K	Ba	Σ cations	Al ³⁺	Al ⁴⁺
14SAB-L017A02-spot1-3	Biotite	6.12	2.87	0.28	0.03	1.90	0.01	3.92	0.00	0.00	0.00	0.02	1.98	0.06	17.14	0.99	1.88
14SAB-L017A02-spot2-1	Biotite	6.15	2.89	0.28	0.02	1.87	0.01	3.92	0.00	0.00	0.00	0.02	1.97	0.05	17.11	1.04	1.85
14SAB-L016A01-spot1-2	Biotite	6.13	2.96	0.27	0.03	1.72	0.01	3.97	0.00	0.00	0.00	0.01	2.01	0.00	17.12	1.09	1.87
14SAB-L009A01-area1-5	Biotite	6.15	3.07	0.20	0.03	0.91	0.00	4.83	0.00	0.00	0.00	0.17	1.66	0.01	17.02	1.22	1.85

Serpentine - Electron Microprobe Data																	
SAMPLE	Mineral	SiO ₂	Al ₂ O ₃	TiO ₂	Cr ₂ O ₃	FeO	MnO	MgO	NiO	ZnO	CaO	Na ₂ O	K ₂ O	BaO	F	Cl	Total
14SAB-L013A02-spot1-4	Serpentine	42.60	0.13	0.01	0.01	17.92	0.49	25.92	0.04	0.00	0.24	0.00	0.00	0.00	0.00	0.00	87.37
14SAB-L013A02-spot1-5	Serpentine	54.73	0.20	0.00	0.01	12.72	0.13	21.83	0.10	0.00	0.32	0.00	0.05	0.00	0.00	0.00	90.08
14SAB-L014A01-spot1-1	Serpentine	37.04	1.42	0.04	0.01	14.11	0.22	30.16	0.01	0.00	0.11	0.02	0.00	0.00	0.00	0.40	83.53
14SAB-L014A01-spot1-2	Serpentine	36.34	0.12	0.01	0.00	11.32	0.09	34.66	0.07	0.00	0.02	0.00	0.00	0.00	0.00	0.60	83.21
14SAB-L076-A02-spot1-4	Serpentine	41.72	0.02	0.00	0.00	22.18	0.25	19.91	0.00	0.00	0.12	0.00	0.01	0.00	0.00	0.14	84.34
14SAB-L076-A02-spot2-3	Serpentine	42.52	0.02	0.00	0.00	5.77	0.01	38.20	0.23	0.00	0.00	0.00	0.00	0.00	0.00	0.00	86.75
14SAB-L076-A02-spot2-4	Serpentine	48.88	0.02	0.04	0.02	20.05	0.13	18.17	0.25	0.00	0.13	0.00	0.11	0.00	0.00	0.00	87.80

White Mica - Electron Microprobe Data																	
Descrip.	Mineral	SiO2	Al2O3	TiO2	Cr2O3	FeO	MnO	MgO	NiO	ZnO	CaO	Na2O	K2O	BaO	F	Cl	Total
14SAB-L013A02-spot1-6	White mica	45.70	37.00	0.02	0.00	0.13	0.29	0.13	0.01	0.00	0.37	0.29	9.61	0.55	0.03	0.00	94.13

Sulphides - Electron Microprobe Data													
SAMPLE	Mineral	Fe	S	Cu	Ni	Co	Zn	As	Mn	Pb	Cd	Sb	Total
15SAB-M005A01-spot1-1	pyrrhotite	59.24	39.00	0.00	0.07	0.17	0.00	0.01	0.00	0.00	0.00	0.00	98.50
15SAB-M005A01-spot2-1	pyrrhotite	60.19	38.74	0.16	0.04	0.18	0.02	0.00	0.01	0.00	0.00	0.00	99.33
14SUB-H029B2-spot1-1	troilite (pyrrhotite)	63.16	35.71	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	98.90
14SUB-H029B2-spot1-4	pyrrhotite	62.07	35.95	0.00	0.04	0.00	0.00	0.00	0.03	0.00	0.00	0.00	98.09
14SAB-L006A02-spot1-2	pyrrhotite	57.61	40.20	0.02	1.08	0.09	0.02	0.00	0.00	0.00	0.00	0.00	99.02
14SAB-L007A01-spot1-1	pyrrhotite	58.85	38.77	0.00	0.37	0.03	0.02	0.01	0.01	0.00	0.00	0.00	98.05
14SAB-L007A01-spot1-2	pyrrhotite	59.97	39.08	0.09	0.28	0.00	0.01	0.01	0.01	0.00	0.00	0.00	99.45
14SAB-L011A01-spot1-1	pyrrhotite	57.69	40.04	0.02	0.94	0.08	0.01	0.00	0.00	0.00	0.00	0.00	98.77
15SAB-L056B02-spot1-2	pyrrhotite	59.96	38.17	0.00	0.39	0.07	0.01	0.00	0.00	0.00	0.00	0.00	98.59
15SAB-L076A02-spot3-3	pyrrhotite	59.06	39.28	0.01	0.18	0.01	0.02	0.00	0.00	0.00	0.00	0.00	98.57
15SAB-L080A03-spot1-2	pyrrhotite	59.47	39.36	0.00	0.56	0.04	0.00	0.01	0.00	0.00	0.00	0.00	99.44
15SAB-D196B01-spot1-1	troilite (pyrrhotite)	63.09	36.09	0.01	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.22
15SAB-D196B01-spot1-5	troilite (pyrrhotite)	63.13	36.07	0.04	0.00	0.03	0.00	0.02	0.00	0.00	0.00	0.00	99.28
15SAB-D196B01-spot1-7	pyrrhotite	59.85	33.47	0.04	0.50	0.06	0.03	0.00	0.02	0.00	0.00	0.00	93.96
14SAB-L009A02-spot1-1	pyrrhotite	60.06	39.07	0.03	0.31	0.01	0.01	0.01	0.00	0.00	0.00	0.00	99.50
14SAB-L009B02-spot1-1	pyrrhotite	59.96	39.22	0.00	0.32	0.05	0.01	0.00	0.01	0.00	0.00	0.00	99.57
14SAB-L017A02-spot3-1	pyrrhotite	60.13	39.09	0.00	0.73	0.00	0.03	0.00	0.00	0.00	0.00	0.00	99.98
14SAB-L017A02-spot3-5	pyrrhotite	59.26	38.92	0.00	1.23	0.01	0.00	0.00	0.00	0.00	0.00	0.00	99.41
15SAB-M005A01-spot2-2	chalcopyrite	30.62	33.87	33.95	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	98.43
14SAB-L003B01-spot3-2	chalcopyrite	30.34	33.67	34.28	0.02	0.01	0.01	0.00	0.00	0.00	0.00	0.00	98.33
14SAB-L006A01-spot3-2	chalcopyrite	29.63	33.55	35.10	0.00	0.01	0.06	0.02	0.02	0.00	0.00	0.00	98.39
14SAB-L006A02-spot1-1	chalcopyrite	29.57	34.39	34.98	0.01	0.00	0.03	0.00	0.01	0.00	0.00	0.00	98.99
14SAB-L011A01-spot1-2	chalcopyrite	29.69	34.63	34.76	0.00	0.01	0.00	0.00	0.01	0.00	0.00	0.00	99.11
14SAB-L015A02-spot 2-2	chalcopyrite	29.34	34.58	34.12	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	98.05
15SAB-L078A02-spot1-4	chalcopyrite (pure)	30.13	33.69	33.95	0.27	0.01	0.02	0.00	0.02	0.00	0.00	0.00	98.10
15SAB-L056B02-spot1-1	chalcopyrite	30.68	33.57	33.58	0.03	0.00	0.05	0.02	0.01	0.00	0.00	0.00	97.94
15SAB-L076A02-spot3-2	chalcopyrite	29.91	34.14	35.37	0.01	0.01	0.00	0.00	0.01	0.00	0.00	0.00	99.46
15SAB-D196B01-spot1-6	chalcopyrite	30.27	33.98	34.73	0.02	0.01	0.00	0.00	0.01	0.00	0.00	0.00	99.02
14SAB-L009A02-spot1-2	chalcopyrite	30.72	33.99	34.96	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	99.70
14SAB-L009B02-spot1-2	chalcopyrite	29.49	33.64	34.53	0.00	0.01	0.00	0.00	0.01	0.00	0.00	0.00	97.68
14SAB-L017A02-spot3-3	chalcopyrite	30.39	33.95	35.43	0.01	0.00	0.05	0.00	0.01	0.00	0.00	0.00	99.84
14SAB-L015A02-spot 1-2	chalcopyrite	29.98	34.78	33.93	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	98.69
15SAB-L080A03-spot1-1	chalcopyrite	30.05	34.12	35.43	0.01	0.00	0.02	0.00	0.00	0.00	0.00	0.00	99.63

Sulphides - Electron Microprobe Data													
SAMPLE	Mineral	Fe	S	Cu	Ni	Co	Zn	As	Mn	Pb	Cd	Sb	Total
14SUB-H029B2-spot1-2	pentlandite	40.37	31.52	0.04	26.12	0.60	0.01	0.03	0.01	0.00	0.00	0.00	98.69
14SUB-H029B2-spot1-3	pentlandite	36.24	33.42	0.00	30.17	0.99	0.00	0.00	0.02	0.00	0.00	0.00	100.83
14SUB-H029B2-spot1-5	pentlandite	33.10	32.65	0.00	32.82	0.71	0.00	0.00	0.02	0.00	0.00	0.00	99.30
14SAB-L007A01-spot1-2	pentlandite	31.20	33.23	0.00	34.79	1.24	0.00	0.01	0.01	0.00	0.00	0.00	100.48
14SAB-L007A01-spot1-2	pentlandite	30.81	33.19	0.00	34.83	1.17	0.05	0.00	0.00	0.00	0.00	0.00	100.05
15SAB-L078A02-spot1-1	pentlandite	28.49	32.13	0.04	37.29	1.05	0.00	0.00	0.01	0.00	0.00	0.00	99.00
15SAB-L078A02-spot1-3	pentlandite	29.24	32.26	0.17	36.02	1.15	0.00	0.00	0.02	0.00	0.00	0.00	98.87
15SAB-L076A02-spot3-1	pentlandite	31.57	32.76	0.00	33.99	1.19	0.00	0.00	0.00	0.00	0.00	0.00	99.52
15SAB-D196B01-spot1-2	pentlandite	28.07	32.68	0.00	25.69	12.34	0.00	0.00	0.00	0.00	0.00	0.00	98.78
15SAB-D196B01-spot1-3	pentlandite	31.05	32.65	0.51	27.71	6.96	0.02	0.00	0.00	0.00	0.00	0.00	98.90
15SAB-D196B01-spot1-4	pentlandite	29.70	32.60	0.23	26.66	9.49	0.00	0.00	0.00	0.00	0.00	0.00	98.68
14SAB-L009A02-spot1-3	pentlandite	33.26	33.66	0.00	30.75	2.15	0.01	0.00	0.00	0.00	0.00	0.00	99.84
14SAB-H006A01-spot1-1	pyrite	47.02	53.06	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00	100.10
14SAB-H006A01-spot1-2	pyrite	47.55	53.21	0.03	0.03	0.13	0.01	0.00	0.00	0.00	0.00	0.00	100.95
14SAB-L003B01-spot3-1	pyrite	46.44	53.28	0.02	0.10	0.08	0.02	0.00	0.00	0.00	0.00	0.00	99.94
14SAB-L006A01-spot3-1	pyrite	46.11	52.45	0.01	0.16	0.13	0.00	0.00	0.00	0.00	0.00	0.00	98.86
14SAB-L015A02-spot 1-1	pyrite	45.28	53.77	0.01	0.01	1.12	0.00	0.00	0.00	0.00	0.00	0.00	100.19
14SAB-L015A02-spot 2-1	pyrite	45.84	52.90	0.00	0.03	0.02	0.00	0.00	0.01	0.00	0.00	0.00	98.80
14SAB-L017A02-spot3-2	pyrite	46.38	53.54	0.03	0.00	0.75	0.01	0.01	0.00	0.00	0.00	0.00	100.72
14SAB-L017A02-spot3-3	pyrite	46.40	53.67	0.00	0.02	0.88	0.00	0.01	0.00	0.00	0.00	0.00	100.98
14SAB-L015A02-spot 1-3	pyrite	45.67	53.74	0.00	0.02	0.77	0.01	0.00	0.00	0.00	0.00	0.00	100.21

Appendix E - Sm-Nd Isotopic Geochemical Data

Sample ID	AGE	Nd (ppm)	¹⁴³ Nd/ ¹⁴⁴ Nd (current)	Sm (ppm)	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd initial	2 sigma error on ¹⁴³ Nd/ ¹⁴⁴ Nd initial	Eps Nd (CHUR)T
14 SAB-H080-A01	1900	20.08	0.511899	4.45	0.1340	0.510224	3.97109E-05	0.89
14 SAB-H080-B01	1900	2.62	0.512810	0.85	0.1966	0.510352	5.7288E-05	3.40
14 SAB-L001-B01	1900	11.46	0.512431	3.21	0.1695	0.510312	4.96361E-05	2.62
14 SAB-L002-A01	1900	3.19	0.512584	0.97	0.1829	0.510298	5.3408E-05	2.33
14 SAB-L002-A02	1900	16.26	0.512052	3.87	0.1438	0.510254	4.24349E-05	1.48
14 SAB-L003-A01	1900	3.18	0.512599	0.97	0.1854	0.510282	5.41079E-05	2.02
14 SAB-L003-B01	1900	8.74	0.512349	2.49	0.1726	0.510191	5.05027E-05	0.25
14 SAB-L004-A01	1900	16.53	0.511974	3.77	0.1379	0.510251	4.07902E-05	1.41
14 SAB-L006-A01	1900	48.30	0.511851	10.30	0.1289	0.510240	3.83084E-05	1.20
14 SAB-L007-A02	1900	17.71	0.512137	4.35	0.1486	0.510278	4.37878E-05	1.96
14 SAB-L008-A02	1900	21.70	0.511943	4.82	0.1343	0.510264	3.98052E-05	1.67
14 SAB-L009-A01	1900	12.87	0.512176	3.19	0.1500	0.510301	4.4175E-05	2.39
14 SAB-L010-A01	1900	10.54	0.511706	2.34	0.1344	0.510025	3.98336E-05	-3.00
14 SAB-L012-A01	1900	10.28	0.512580	3.07	0.1807	0.510321	5.27874E-05	2.79
14 SAB-L012-A02	1900	8.03	0.511608	1.75	0.1319	0.509959	3.91335E-05	-4.31
14 SAB-R017-A02	1900	11.73	0.511010	1.61	0.0828	0.509975	2.57662E-05	-3.99
14 SAB-R010-A01	1900	2.10	0.512761	0.67	0.1915	0.510367	5.58364E-05	3.69
14 SAB-S130-A01	1900	5.99	0.512132	1.47	0.1488	0.510271	4.3851E-05	1.82
15 SAB-L057-B02	1900	8.58	0.511887	2.32	0.1637	0.509840	4.80118E-05	-6.64
15 SAB-L078-A01	1900	1.06	0.513082	0.38	0.2135	0.510412	6.20663E-05	4.58
15 SAB-L080-A01	1900	1.86	0.513386	0.71	0.2315	0.510491	6.71572E-05	6.13
15 SAB-L119-A01	1900	3.90	0.512493	1.16	0.1801	0.510241	5.26206E-05	1.23