2.0 to 1.9 Ga large igneous province magmatism in northern Quebec –

U-Pb geochronology and geochemistry

By

Nico Kastek

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Nico Kastek

ABSTRACT

The northeastern margins of the Superior craton contain the Cape Smith belt and the Labrador Trough (including the Roberts Lake Syncline), which range from 2.2 to 1.9 Ga in age. This thesis examines the age of the Cape Smith belt, its correlation to other units of similar age and its possible extent to the southeast.

Seven U-Pb ages on magmatic baddeleyite and zircon and 84 geochemical analyses were obtained of magmatic units throughout the Cape Smith belt. The most precise are two ages each for the Povungnituk and Chukotat Groups.

The Povungnituk Group yields ages of 1998 ± 6 Ma and 1967 ± 7 Ma. The older age represents the main volcanic pulse, and matches previously obtained U-Pb ages for the Watts Group (Purtuniq) ophiolite of the northern Cape Smith Belt and the Minto and Lac Shpogan dyke swarms that intrude the Superior craton to the south. They define a large igneous province (LIP), extending over an area of $>400,000 \text{ km}^2$, herein defined as the Minto-Povungnituk LIP.

The Minto-Povungnituk LIP can be divided into two spatially separate geochemical domains with different mantle source characteristics that can be produced in two ways: (1) melting of different portions of Superior craton lithosphere; or (2) two distinct deep mantle sources that remained separated within the ascending plume.

A feeder dyke of the Chukotat Group yielded a U-Pb age of 1874 ± 3 Ma, representing the younger end of its age range. An age obtained from the top of the Chukotat Formation yielded 1861 ± 28 Ma and confirms the previously geochemistry-based interpretation that the Chukotat Group represents a single event.

To the southeast of the Cape Smith belt, the Roberts Lake Syncline hosts two Paleoproterozoic mafic magmatic sequences but their correlation with the units of Labrador Trough to the south, or the Cape Smith belt to the north is uncertain. 94 geochemical analyses were obtained on the two volcanic units and the data show a closer match to the chemical stratigraphy observed within the Cape Smith belt, suggesting that the Roberts Lake Syncline is the easternmost portion of the Ungava Orogen rather than the northernmost part of the Labrador Trough.

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I would like to first thank my primary supervisors, R.E. Ernst and B.L. Cousens. You have been an incredible help and I am thankful that you took the patience for the prolonged time it has taken me to get to this point. I also want to thank P. Sylvester, who supervised me during my first year at Memorial University of Newfoundland and dealt with me passing all of my PhD requirements.

I owe a large debt of gratitude to W.R.A. Baragar, M.R. St-Onge and J.E. Mungall for allowing me to use some of their samples. I want to especially thank J.E. Mungall, who gifted me with a significant amount of time and helped me greatly in developing the fifth chapter of the thesis.

I want to thank W. Bleeker, who guided me in the field and to whom I owe a deep understanding of my study area. He has also been an indispensable contributor to the development of our first publication on the geochronology and geochemistry of the Povungnituk Group.

I want to thank Glencore for providing travel and access to the Raglan mine. I cherish the time I was able to spend on their facility and the samples they let me take. I also want to thank Anglo American and especially J.-F. Belanger, for welcoming me on their drill site in the Roberts Lake Syncline, access to all their samples and time on their on-site helicopter.

I would not have been able to collect any of the data without M. Shaffer and D. Goudie, who assisted me with work on the SEM at Memorial University; U.

Söderlund, who let me visit him at Lund University to learn his technique of separating baddeleyite; S. Kamo and M. Hamilton, who welcomed me in their laboratory at the University of Toronto and taught me how to analyze samples using ID-TIMS, sacrificing multiple weekends; K. Chamberlain who took me on as an assistant during one of his trips to UCLA for IN-SIMS dating and giving me enough of his measuring time to date two of my own samples; S. Jantzi for assistance with LA-ICP-MS geochronology at Memorial University; and D. Davis for assistance with LA-ICP-MS geochronology at the University of Toronto.

S. Zhang has been a great help, providing guidance in the clean lab at Carleton University and amazing assistance during TIMS measurements.

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STATEMENT OF CONTRIBUTION

Fieldwork and sample collection

I (Nico Kastek) visited field locations the Roberts Lake Syncline in northern Quebec in the summer of 2013 to collect 10 samples. I was accompanied on the trip by W. Bleeker (Geological Survey of Canada), who helped me identify suitable samples and provided expertise on the regional geology. We were hosted by J.-F. Belanger from Anglo American, who led a drill project in the Roberts Lake Syncline and who provided us with transportation (air and land).

An additional 7 rock samples and 84 rock powders were provided by W.R.A. Baragar and 2 rock samples from M.R. St-Onge for the Povungnituk Group. J.E. Mungall provided 84 unpublished major element analyses from the Roberts Lake Syncline, from which powders of 39 samples were taken for further trace element and platinum group element (PGE) analyses.

Sample preparation

Single polish thin-sections for selected coarse grained samples of the Povungnituk and Chukotat Group were prepared at Memorial University of Newfoundland.

Samples not obtained as powder or crushed material were cleaned, cut into slabs, crushed to <2 mm and subsequently powdered in an agate mill by myself under the supervision of T. Mount (Carleton University).

Major and trace element analyses

All samples for the Povungnituk Group were sent to ALS Geochemistry laboratories in North Vancouver, British Columbia for major element analysis via Inductively Coupled Plasma Atomic Emissions Spectrometry (ICP–AES) as well as trace and rare earth element (REE) analysis via Inductively Coupled Plasma Mass Spectrometry (ICP-MS).

Major element concentrations for samples that J.E. Mungall collected from the Roberts Lake Syncline in 1998, 1999 and 2000 were determined by borate fusion X-ray fluorescence analysis at the Geoscience Laboratories (Geo Labs) of the Ontario Geological Survey in Sudbury, Ontario. Samples collected by myself in 2013 from the Roberts Lake Syncline and a subset of the sample suite provided by J.E. Mungall were sent to ALS Geochemistry laboratories in North Vancouver, British Columbia for ICP-AES.

Isotopic analyses

I performed sample preparation and all of the isotopic analyses for thermal ionization mass spectrometry (TIMS) at Carleton University's Isotope Geochemistry and Geochronology Research Facility (IGGRF) under the supervision of S. Zhang.

Geochronology

I identified zirconium-bearing phases in thin section at the scanning electron microscope (SEM) at the TERRA facility – CREAT (Memorial University Earth Science Department). I took promising samples to Lund University Geochronology laboratory, where I mechanically separated baddeleyite grains under the supervision of U. Söderlund. I separated additional baddeleyite grains at the Jack Satterly Geochronology Laboratory at the University of Toronto under the supervision of B. Foursenko. A difficult sample (BLS-73-31) was further processed by S. Kamo (also at the Jack Satterly Geochronology Laboratory) using heavy liquid (methylene iodide) and I picked the baddeleyite crystals from the remaining grains. I then assisted S. Kamo in Isotope Dilution Thermal Ionization Mass Spectrometer (ID-TIMS) U-Pb analysis. The data was edited by S. Kamo, who also interpreted the respective ages.

I provided K. Chamberlain (University of Wyoming) with polished thin sections, in which zirconium-bearing phases had been identified. He re-analysed the samples via wavelength dispersive spectroscopy (WDS) at the University of Wyoming and identified zirconium-bearing phases using energy dispersive spectroscopy (EDS). I assisted K. Chamberlain with secondary ion mass spectrometry (SIMS) analyses at the University of California at Los Angeles. Data editing and age interpretation was done by K. Chamberlain.

I took two polished thin sections to the TERRA facility – CREAT at Memorial University of Newfoundland, where I measured in situ U-Pb ages via laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) under the

supervision of S. Jantzi. Data editing was done by S. Jantzi and age interpretation of the data was done by myself.

I submitted three quartzite samples to the Jack Satterly Geochronology Laboratory at the University of Toronto, where detrital zircon grains were separated. I separated selected zircon grains from the full suite of zircons at the University of Toronto measured the U-Pb ratios of the samples via LA-ICP-MS under the supervision of D. Davis. Data editing and some age interpretation was done by D. Davis. Further age interpretation was done by myself.

Thesis and manuscript construction

I wrote each chapter and designed and drafted all figures, with exception of the concordia and Tera-Wasserburg diagrams for ID-TIMS and SIMS (U-Pb) analyses or those otherwise referenced. All aspects of the dissertation benefited from extensive discussion with R.E. Ernst, and B.L. Cousens (co-supervisors) and also J.E. Mungall (Carleton University), W. Bleeker (Geological Survey of Canada) and other collaborators with this project including but not limited to: S. Kamo (University of Toronto), W.R.A. Baragar (Geological Survey of Canada), U. Söderlund (Lund University), K. Chamberlain (University of Wyoming), C.C. Rogers (Carleton University), and S. Davey (Carleton University).

The thesis is constructed around three interrelated chapters (Chapter 3-5). Parts of the contents of Chapters 3 and 4 were published in *Lithos* (Kastek et al., 2018). Chapter 5 has been submitted to *Lithos*.

The information in Chapter 3 can be published after some additional U-Pb dating is obtained.

Statement of originality

This thesis was born out of a need to understand the geological history of the Cape Smith belt and the exact timing and duration of its magmatic events in order to constrain possible correlations with other magmatic events within and around the Superior craton.

The present study represents multiple ages on the magmatic succession of the Cape Smith belt using various U-Pb geochronological techniques best suited to the individual quality of the samples. Together with an extensive geochemical dataset on the Povungnituk Group of the Cape Smith belt, the Minto dykes of the interior of the Superior craton and the lavas of the Roberts Lake Syncline, the duration and extend of a Minto-Povungnituk large igneous province (LIP) can be constrained with implications for the role of a mantle plume in its formation.

New ages on the Chukotat Group of the Cape Smith belt are presented confirming that the whole group represents a single short-lived event.

Chapter 3

For this chapter I identified and separated the zirconium-bearing phases and assisted in all U-Pb dating techniques. Data editing was done by S. Kamo (University of Toronto), K. Chamberlain (University of Wyoming), S. Jantzi (Memorial University), and D. Davis (University of Toronto). The age interpretation of the ID-TIMS data was done by S. Kamo. The age interpretation of the IN-SIMS data was done by K. Chamberlain. The age interpretation for the

LA-ICP-MS data for magmatic baddeleyite and zircon was done by myself. The age interpretation for the LA-ICP-MS data was done jointly by myself and D. Davis.

I tested the correlation of all the magmatic units with other magmatic events within and around the Superior craton.

Chapter 4

I obtained and compared the geochemical data from the Povungnituk Group with other units in the northeastern Superior craton of similar age and the petrogenesis of this 2.0 Ga Minto-Povungnituk LIP was presented. I developed possible models to explain the identified bilateral asymmetry within the two different domains of this LIP. These new insights into the Povungnituk Group and the broader Minto-Povungnituk LIP are part of an article published in *Lithos* in 2018 and this publication benefited from the input of all co-authors. It should be cited as follows:

Kastek, N., Ernst, R.E., Cousens, B.L., Kamo, S.L., Bleeker, W., Söderlund, U., Baragar, W.R.A., Sylvester, P., 2018. U-Pb Geochronology and Geochemistry of the Povungnituk Group of the Cape Smith Belt: Part of a Craton-Scale Circa 2.0 Ga Minto-Povungnituk Large Igneous Province, Northern Superior Craton. *Lithos* 320-321, 315-331.

Chapter 5

The geochemical dataset on the Roberts Lake Syncline is obtained from analyses on a combination of samples from my own field season, major element concentrations provided by J.E. Mungall and additional trace element concentrations measured at ALS. I performed all the modelling to explain the geochemical characteristics observed in the rocks and compared the units of the Roberts Lake Syncline to possible equivalents in the Cape Smith belt and the Labrador Trough. Chapter 5 is presented as a manuscript submitted to *Lithos*.

Kastek, N., Mungall, J.E., Ernst, R.E., Cousens, B.L., 2019. Geochemistry of the Roberts Lake Syncline mafic lavas (NE Superior craton): Comparison with Paleoproterozoic volcanic sequences of the Cape Smith belt and the Labrador Trough. *Lithos*, submitted.

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1 INTRODUCTION

Large igneous provinces (LIPs) are huge volumes of magma (intrusive and extrusive) that are typically thought to occur when hot, deep-sourced mantle plumes rise to the base of the Earth's lithosphere and partially melt, although non-plume origins are also proposed (Coffin and Eldholm, 1994; Ernst and Buchan, 2001; 2003; Courtillot and Renne, 2003). LIPs have become an important focus for research in recent years due to their use in paleocontinental reconstructions (e.g., Bleeker and Ernst, 2006; Ernst et al., 2013), in exploration targeting (e.g., Ernst and Jowitt, 2013, 2017), and as a result of their links to dramatic climate change (e.g., Wignall, 2001, 2005; Bond and Grasby, 2017; Ernst and Youbi, 2017). More precisely, LIPs represent large volume ($> 0.1 \text{ Mkm}^3$; frequently above 1 Mkm^3), mainly mafic (-ultramafic) magmatic events of intraplate affinity (based on tectonic setting and/or geochemistry) that occur in both continental and oceanic settings, and are typically either of short duration ($< 5 \text{ Ma}$; sometimes $< 1 \text{ Ma}$) or consist of multiple short pulses over a maximum of a few 10s of Myr. LIPs comprise volcanic packages (flood basalts) and a plumbing system of dyke swarms, sill complexes, layered intrusions, and a magmatic underplate (Coffin and Eldholm, 1992; Bryan and Ernst, 2008; Ernst, 2014; Ernst and Youbi, 2017).

1.1 Project rationale

This thesis focuses on the Cape Smith belt in northernmost Quebec (Fig. 1.1). The southern part of the belt comprises massive volumes of basaltic lava flows that erupted between ca. 2.0 and 1.9 Ga onto the rifted margins of the northern Superior craton. These flows represent one of the most extensive extrusive LIP remnants of the Superior craton. They are dominated by the Povungnituk Group (Hynes and Francis, 1982; Francis et al., 1983), interpreted as continental flood basalts, and their associated dolerite sills which were constrained in age between 2.04 and 1.91 Ga (Machado et al., 1993), and the younger ca. 1880 Ma Chukotat Group (Wodicka et al., 2002; Randall, 2005; Bleeker and Kamo, 2018), which forms part of the Circum-Superior LIP (e.g., Baragar and Scoates, 1981; Ernst and Bleeker, 2010; Ciborowski et al., 2016). The extrusive components of LIPs around the Superior craton have been linked to sills and dykes in the interior of the craton (Ernst and Bleeker, 2010; Minifie et al., 2013; Hamilton et al., 2017). Modeland et al. (2003) examined the geochemical character of the Povungnituk Group. They identified two endmember compositions involved in the formation of the magma that produced the flood basalts.

Some authors have suggested that the Cape Smith belt extends further than previously mapped (Ferron et al., 2000; Madore and Larbi, 2001; Bleeker and Kamo, 2018). Along the northeastern margin of the Superior craton, southeast of the Cape Smith belt, lies the Roberts Lake Syncline (Hardy, 1976). The volcanic units of the Roberts Lake Syncline have been previously mapped as a

continuation of the Labrador Trough (Hardy, 1976; Clark and Wares, 2006), of the New Quebec orogen (Hoffman, 1988). The supracrustal rocks of the Labrador Trough (Fig. 1.1) constitute the foreland of the New Quebec orogen and form a thrust and fold belt on the margin of the Superior craton (Clark and Wares, 2006) and comprise magmatism that ranges from 2.2 and 1.8 Ga (Rohon et al., 1993; Skulski et al., 1993; Findlay et al., 1995; Machado et al., 1997). Some have correlated the volcanic units of the Roberts Lake Syncline with units of the Cape Smith belt (Ferron et al., 2000; Madore and Larbi, 2001; Bleeker and Kamo, 2018), while others have correlated the Roberts Lake Syncline units with magmatism of the Labrador Trough further south (Hardy, 1976, Clark and Wares, 2006).

A more detailed geological history of the region is presented in the geological overview of Chapter 2.

Figure 1.1. Overview map of the geological units that this thesis focuses on. On the northern margin is the Cape Smith belt. Along the eastern margin is the Labrador Trough with the Roberts Lake Syncline at the northernmost tip. Red boxes in the map indicate the geological units described in chapter 2. Small map shows the outline of North America with the Superior craton in grey and the location of the map in red. Modified from St-Onge et al. (2004) and Clark and Wares (2006). Small overview map includes in grey the location of the Superior craton (Goodfellow, 2007).

1.2 Objectives of the project

- Obtain more detailed geochronology for the Povungnituk Group (Chapter 3), for which the age is currently poorly constrained over a range of 80 myr. A more precise age is necessary to confidently link these large flood basalt units to other events in and around the Superior craton.
- Obtain additional geochemical analyses to better identify possible endmembers and the role of mantle plumes, depleted upper Proterozoic mantle, and subcontinental lithospheric mantle in the generation of the Povungnituk Group (Chapter 4).
- Compare this detailed geochemistry with other units that have been linked to the Povungnituk Group based on geochronology. Only with the new data can a full geochemical fingerprint of a possible LIP be identified (Chapter 4).
- Obtain additional ages on the feeder systems of the Chukotat Group (Chapter 3), which can provide an estimate of the duration of the primitive Chukotat magmatism, which is the host for Ni-Cu-PGE economic deposits of the Raglan and Expo Intrusive suite (Leshner, 2007; Mungall, 2007).
- Date the uppermost, most evolved units of the Chukotat Group to confirm whether the entire basalt sequence belongs to the same event and represents the remnants of a single LIP (Chapter 3), as earlier suggested on the basis of geochemistry (Francis and Hynes, 1979; Hynes and Francis, 1982; Francis et al., 1983; Ciborowski et al., 2016).

- Obtain geochemistry data on the Roberts Lake Syncline for comparison with the Cape Smith belt to the north and the Labrador Trough to the south (Chapter 5).

1.3 Thesis structure

This thesis is split into six chapters. Because of their individual character, references are given at the end of each chapter.

After this short introductory chapter, Chapter 2 describes the geology of the Cape Smith belt, the Labrador Trough and the Roberts Syncline and summarizes previous geochronological work done on these.

Chapter 3 introduces U-Pb geochronology analyses that were undertaken for this thesis. The chapter presents the results of seven magmatic zircon/baddeleyite ages, three detrital zircon ages and their implications for our understanding of the overall geology of the Cape Smith belt. The new ages obtained for units in the Cape Smith belt are compared with other units around the Superior craton.

Chapter 4 includes major and trace element and Sm-Nd radiogenic isotopic analyses for the mafic units of the Povungnituk Group. It describes the geochemical evolution and mantle source of these units. A detailed geochemical comparison with associated units identified in Chapter 3 is made and possible models for the generation of a Minto-Povungnituk LIP are made. These aspects from Chapters 3 and 4 are integrated into my 2018 publication in the journal *Lithos*.

Kastek, N., Ernst, R.E., Cousens, B.L., Kamo, S.L., Bleeker, W., Söderlund, U., Baragar, W.R.A., Sylvester, P., 2018. U-Pb Geochronology and Geochemistry of the Povungnituk Group of the Cape Smith Belt: Part of a Craton-Scale Circa 2.0 Ga Minto-Povungnituk Large Igneous Province, Northern Superior Craton. *Lithos* 320-321, 315-331.

Chapter 5 presents data for the Roberts Lake Syncline volcanism, including a detailed traverse along the western limb of the syncline. Major element, trace element and platinum group element concentrations were obtained for all samples and used to identify unique geochemical signatures of the Roberts Lake Syncline that are compared with the 2170 Ma Biscotasing-Cycle 1 and 1880 Ma Cycle 2 events of the Labrador Trough, and the 1998 Ma Povungnituk Group and 1880 Ma Chukotat Group of the Cape Smith belt. The conclusion is that the Roberts Lake Syncline represents an extension of the Cape Smith package, in addition to the rocks of the Labrador Trough to which it has been associated with in the past. The content of Chapter 5 is currently under review in the journal *Lithos*.

Kastek, N., Mungall, J.E., Ernst, R.E., Cousens, B.L., 2019. Geochemistry of the Roberts Lake Syncline mafic lavas (NE Superior craton): Comparison with Paleoproterozoic volcanic sequences of the Cape Smith belt and the Labrador Trough. *Lithos*, submitted.

A final short summary of the thesis and conclusions is presented in Chapter 6.

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2 GEOLOGICAL OVERVIEW

This thesis considers magmatic units in two areas. The first one is the Cape Smith belt, which is a part of the Paleoproterozoic Ungava Orogen, located at the northern tip of Quebec (St-Onge et al., 1992; Bleeker and Kamo, 2018). The second research area comprises the Roberts Lake Syncline (Payne Zone of Clark and Wares, 2006) and is located between the northern extremity of the Labrador Trough of the New Quebec orogen and the eastern extremity of the Cape Smith belt (Hardy, 1976). In this chapter the geological setting and previous geochronology is given for the Cape Smith belt, the Labrador Trough, and the Roberts Lake Syncline.

2.1 The Cape Smith belt

2.1.1 Geological setting of the Cape Smith belt

The Cape Smith fold and thrust belt (Fig. 2.1) contains well-preserved volcano-sedimentary and plutonic suites, which overlie Archean tonalites and gneisses of the northern Superior craton. The Cape Smith fold and thrust belt can be subdivided into northern allochthonous, and southern allochthonous to parautochthonous domains. These two domains have contrasting metamorphic and tectonic histories, until they were amalgamated during the climax of the Trans-Hudson Orogeny. The northern domain contains allochthonous crustal components of an interpreted ophiolite complex (Watts Group) (Scott et al., 1999) and a forearc complex (Parent Group, Spartan Group) (St-Onge et al.,

1992). The southern domain, comprising the Povungnituk and Chukotat Groups, has been interpreted to represent a long-lived passive continental margin that laps on the northern margin of the Archean Superior craton (Scott et al., 1999). These domains were deformed and imbricated during terminal collision of the Superior Province with the Archean Rae Province to the north (St-Onge et al., 2006; Corrigan, 2012).

This thesis concentrates on the southern part of the belt, hosting the Povungnituk and Chukotat Groups.

At its base the Cape Smith supracrustal sequence consists of the Dumas Formation of the Lamarche Group, which is an autochthonous to parautochthonous assemblage of ferruginous sandstone, conglomerate, quartzite and iron formation, overlain by progressively finer-grained wackes and rare carbonates. It unconformably overlies the Superior craton basement. These rocks represent continent-derived sediments accumulated on a developing rifted continental margin (St-Onge et al., 1992).

The Dumas Formation is overlain by the predominantly basaltic Beuparlant Formation, which in turn is overlain by the sedimentary Nuvilik Formation. The Beuparlant Formation comprises tholeiitic basalts that occur both as pillowed and tabular basalt flows, intercalated with volcanoclastic and locally graphitic siltstone horizons and minor carbonate sedimentary rocks. Geochemical constraints on the Povungnituk volcanism have been obtained in previous studies (Hynes and Francis, 1982; Francis et al., 1983; Picard et al., 1989a;

1989b; Hegner and Bevier, 1991; Legault et al., 1994; Dunphy et al., 1995; Modeland et al., 2003) and are discussed in Chapter 4 of the thesis.

Although tholeiitic basalts dominate the Povungnituk Group, small alkaline volcanic suites have also been documented. Near the base of the succession, in the western portion of the Cape Smith fold belt, occur carbonate-rich alkaline lavas as described by Baragar et al. (2001). This Lac Leclair volcanic suite comprises a 500 m-thick sequence of carbonatitic, ultramafic lapilli tuffs and lavas that outcrop along the southern extremity of the belt, near the unconformity with the Superior craton. The presence of the Lac Leclair volcanic suite among the platformal sedimentary rocks of the lowermost Povungnituk Group indicates that it may be one of the first volcanic manifestations of rifting and subsidence during the development of the Cape Smith fold belt (Baragar et al., 2001).

Another alkaline volcanic package is exposed in two adjacent thrust blocks near the top of the Beuparlant Formation in the Cecilia and Kenty Lake areas (Gaonac'h et al., 1992) (Fig. 2.1). This alkaline suite is several hundred meters thick, and comprises a lower sequence of basanites and nephelinites intercalated with pyroclastic deposits and an upper sequence of trachytic lavas with rare alkali-feldspar phenocrysts (Gaonac'h et al., 1992).

The Beuparlant Formation is overlain by a succession of deep-water sedimentary rocks of the Nuvilik Formation, including greywackes, cherts, and graphitic argillites and rare pebbly quartz arenites (Mungall, 2007). In some parts of the Cape Smith Belt, the Nuvilik Formation overlies alkaline volcanic rocks of the Cecilia Formation (Picard, 1995).

Based on re-evaluations of the Povungnituk stratigraphy (Mungall, 2007; Bleeker and Kamo, 2018) it was concluded that the contact between the uppermost pillow basalt of the Beuparlant Formation and the base of the Nuvilik Formation is conformable. Furthermore, there is no evidence for a thrust fault at the top of the Nuvilik Formation, and therefore the Nuvilik Formation does not represent a tectonically transported metasedimentary package equivalent to the Dumas Formation, as was previously proposed by Lucas and St-Onge (1992); indeed, the majority of thrust faults inferred by earlier workers in the Cape Smith belt can be shown not to exist (cf. Mungall, 2007 and Bleeker and Ames, 2017 for discussion).

The Nuvilik Formation is overlain by the accumulation of predominantly volcanic deep-water assemblage of the Chukotat Group (Francis and Hynes, 1979; Hynes and Francis, 1982; Francis et al., 1983).

The Chukotat Group comprises a succession of mafic volcanic rocks that was formerly suggested to record a transition in chemical composition from rift-related to mid-ocean ridge basalt (MORB)-like composition (Francis et al., 1981, 1983; Picard, 1989a; 1989b; St-Onge et al., 1992). However recent consideration of its tectonic and stratigraphic setting has resulted in a consensus that the komatiitic and basaltic Chukotat Group is part of the Circum Superior LIP, emplaced through thinned lithosphere surrounding the Superior Province about 150 Ma after rifting and therefore not directly related to extensional tectonics (e.g., Mungall, 2007; Heaman et al. 2009; Minifie et al. 2013; Ciborowski et al., 2017).

Figure 2.1. Geological map of the Ungava Orogen, northern Quebec, Canada (after St-Onge et al., 2004). Stars represent the location of the Lac Leclair alkaline complex from Baragar et al. (2001) and the Kenty Lake alkaline complex from Gaonac'h et al. (1992). Small map shows the outline of northern Quebec with the Cape Smith belt and the Labrador Trough in grey (after St-Onge et al., 2004; Clark and Wares, 2006; Goodfellow, 2007). Red box indicates the location of Fig. 1.1.

2.1.2 Previous geochronology of the Cape Smith belt

A mafic sill in the upper Dumas Formation is 2038 ± 3 Ma (Machado et al., 1993) and was originally considered to be comagmatic with volcanic rocks of the Beuparlant Formation. At a higher stratigraphic level, a small granodiorite intrusion in Povungnituk pillow basalts has a robust zircon age of 1991 ± 2 Ma (Machado et al., 1993). Therefore, the main part of the Beuparlant Formation must predate 1991 Ma. These age constraints show that Povungnituk magmatism occurred between 2038 and 1991 Ma.

A rhyolite sample from alkaline Kenty Lake suite at the top of the Beuparlant Formation has been dated at 1958 ± 3 Ma (Parrish, 1989). This age marks a late-stage magmatic pulse and constrains the overlying sedimentary Nuvilik Formation to be younger than 1959 Ma.

The depositional age and detrital source of the Nuvilik Formation is not known. Its time of deposition may span the interval from ca. 1959 Ma to ca. 1883 Ma when the overlying basaltic Chukotat Group was deposited.

The strata of the Povungnituk Group were intruded at 1882.7 ± 1.3 Ma (Randall, 2005) by a series of mafic to ultramafic bodies of the Expo Intrusive Suite (Mungall, 2007). The Expo Intrusive Suite, the Raglan Formation and the Chukotat Group can be considered a single magmatic suite, based on age similarity and the match between the observed and proposed primitive magmas (Mungall, 2007).

U-Pb dating of the ultramafic and gabbroic intrusions of the Raglan and Expo Intrusive Suites, which are interpreted to have fed the Chukotat volcanism, has yielded ages of $1887 \pm 39/-11$ Ma (Wodicka et al., 2002), 1882.7 ± 1.3 Ma (Randall, 2005) and 1881.5 ± 0.9 Ma (Bleeker and Kamo 2018, re-dated from Parrish, 1989). These ages mark the onset of Chukotat emplacement. The duration of Chukotat magmatism can be constrained by a younger age of 1870 ± 4 (Parrish, unpublished data; mentioned in St-Onge et al., 1992).

Published dates are overlain on a stratigraphic column of the Cape Smith belt in Figure 2.2.

Figure 2.2. Stratigraphic column for the Cape Smith belt. Modified from Bleeker and Ames (2017) and Mungall (2007). Age references: Parrish (1989); St-Onge et al. (1992); Machado et al. (1993); Wodicka et al. (2002); Randall (2005), Bleeker and Kamo (2018).

2.2 Labrador Trough

2.2.1 Geological setting of the Labrador Trough

The Paleoproterozoic New Quebec Orogen (Hoffman, 1988) is a fold and thrust belt situated between the Archean Superior and Rae provinces in northeastern Quebec and western Labrador (Fig. 2.3). It extends over 850 km, from the Grenville Front in the south, to as far north as Ungava Bay (Clark, 1994). The orogen constitutes part of the marginal sequence formerly described as the “Circum-Ungava Geosyncline”, which surrounds the northeastern part of the Superior Province, including the Ungava Orogen (Cape Smith belt) and the Belcher Islands (Dimroth et al., 1970; Baragar and Scoates, 1981).

The first stratigraphic and structural synthesis was established on the central part of the New Quebec orogen (Dimroth, 1970, 1972, 1978, 1981; Dimroth et al., 1970; Dimroth and Dressler, 1978; Wardle and Bailey, 1981). It was subdivided into three supracrustal belts: (1) a western, parautochthonous to allochthonous “miogeosynclinal” belt composed mainly of platform sedimentary rocks: (2) a central, allochthonous “eugeosynclinal” belt composed mainly of greenschist facies, deep-water, volcano-sedimentary rocks intruded by numerous gabbro sills, and (3) an eastern allochthonous belt marking the beginning of the hinterland and composed of amphibolite facies rocks, in part equivalent to rocks of the central belt, and tectonic nappes of Archean basement rocks. Later correlations between the central and northern parts of the New Quebec orogen resulted in the first lithotectonic models for certain parts (Wares et al., 1989; Wardle et al., 1990; Wares and Goutier, 1990; Skulski, 1993).

The supracrustal rocks of the Labrador Trough constitute the foreland of the New Quebec orogen. These rocks form a thrust and fold belt on the margin of the Superior Province (Clark and Wares, 2006) and make up the Kaniapiskau Supergroup (Frarey and Duffell, 1964), which comprises three main depositional cycles. The cycles are separated from each other by regional erosional unconformities (Dimroth et al., 1970; Clark, 1994). The three cycles have later been revised and the different stratigraphic formations re-categorized by Clark and Wares (2006); their revised nomenclature is used here. Because of its historical usage, I use the term Labrador Trough for the remainder of the thesis.

The first cycle occurs mainly in the central part of the orogen and is composed of an intracratonic rift basin sequence overlain by a passive margin sequence. The sediments lie discordantly on the Archean Superior craton margin. The first sequence (Seward Group) constitutes of sandstones and conglomerates and represents an immature, continental rift sequence deposited ~2.2 Ga ago as the result of the rifting of the Archean continent along the northeastern margin of the Superior Province (Hoffman, 1988; Wardle et al., 2002). Mafic, weakly alkaline volcanic activity was contemporaneous with sedimentation. Following the deposition of these units, sandstone and dolomites were deposited on a passive margin platform (Pistolet Group). This platform eventually foundered, and basalt (Bacchus Formation) and flysch of the Swampy Bay Group were deposited in a marine basin, which are overlain by the shallow-water rocks of a dolomitic reef complex (Attikamagen Group), indicating the

establishment of a platform and a marine regression at the end of the first cycle (Hoffman and Grotzinger, 1989).

The second cycle unconformably overlies the Superior craton and first cycle rocks (Dimroth, 1978). It includes a transgressive sequence composed of platform sediments, alkalic volcanic rocks and turbidites (Ferriman Group). In the northern part of the orogen, the Baby Iron Formation and overlying Baby turbidites of the eastern allochthonous belt are correlated with the Ferriman Group (Clark and Thorpe, 1990). The Baby Formation is overlain by a thick basaltic sequence (Hellencourt Formation; Skulski et al., 1993), that together form the Koksoak Group. In the southern part of the northern sector, the lower part of the Koksoak Group interfingers with a sequence of mafic and felsic pyroclastic rocks, detrital sediments, dolomites, and tuffaceous dolomites (Le Moyne Group). The Le Moyne Group is intruded by a large carbonatite complex (Birkett and Clark, 1991). In the central sector the Doublet Group consists of mafic pyroclastic rocks, overlain by a unit of sandstone, siltstone, and shale, followed by a thick sequence of basalt lavas (Willbob Formation). Because of their geochemical similarities, second cycle mafic rocks and gabbro sills are interpreted to be comagmatic (Rohon et al., 1993; Skulski et al., 1993; Findlay et al., 1995). Because of the presence of major faults, correlation between western and eastern units is uncertain. North-south correlations are difficult because of the discontinuity of certain units, and the different stratigraphic names used by the various authors (Machado et al., 1997). Rocks of Cycles 1 and 2 are intruded by numerous tholeiitic, mafic-ultramafic sills classified under the general name of

“Montagnais Sills”. These sills are contemporaneous and comagmatic with volcanic rocks of Cycle 2 (Rohon et al., 1993; Skulski et al., 1993; Findlay et al., 1995).

The second cycle is overlain by clastic sediment interpreted as synorogenic molasses and assigned to a third sedimentary cycle (Hoffman, 1988).

Figure 2.3. Geological map of the Labrador Trough. Modified from Clark and Wares (2006). Small map shows the outline of northern Quebec with the Cape Smith belt and the Labrador Trough in grey (after St-Onge et al., 2004; Clark and Wares, 2006; Goodfellow, 2007). Red box indicates the location on Fig. 1.1.

2.2.2 Geochronology of the Labrador Trough

Limited geochronological data provide important constraints on the temporal evolution of the orogen. An age of 2169 ± 4 Ma was obtained for a gabbro sill intruding immature sediments of the Seward Group in cycle 1 (Rohon et al., 1993). This agrees with the interpretation that the rifting of the Archean continent along the northeastern margin of the Superior Province started ~ 2.2 Ga ago (Hoffman, 1988; Wardle et al., 2002) and is associated with the 2.22-2.21 Ga Ungava-Nipissing LIP with a plume centre defined by a radiating dyke swarm on the northeastern side of the Superior craton (Buchan et al. 1998; Clark and Wares, 2006; Ernst and Bleeker, 2010). This age indicates that rifting of the Archean continent began before 2169 Ma. A rhyolite dyke cutting the upper part of the Bacchus Formation in the Attikamagen Group was dated at $2142 \pm 4/-2$ Ma (unpublished data cited by Clark, 1984) and shows that most of Cycle 1 was deposited in a timespan of at least 27 myr (2169-2142 Ma). Carbon isotopic compositions of dolomites of the upper Cycle 1 (Melezhik, 1997) might indicate that sedimentation has continued until 2.06 Ga (Karhu and Holland, 1996).

A few ages are also available for the second cycle. A carbonatite dyke, contemporaneous with the Ferriman Group, was dated at 1880 ± 2 Ma (Chevé and Machado, 1988). This age overlaps with the age of 1878 ± 1 Ma obtained on felsic volcanic rocks in the lower Ferriman Group and 1884 ± 4 Ma for a Montagnais glomeroporphyritic gabbro sill cutting turbidites in the upper Ferriman Group (Findlay et al., 1995). The latter sample included discordant zircon data and has been re-dated by Bleeker and Kamo (2018), who obtained a new age of

1878.5 $\pm$ 0.8 Ma. A glomeroporphyritic gabbro sill intruding basalts at the top of the Hellancourt Formation gave an age of 1874 $\pm$ 3 Ma (Machado et al., 1997). A Pb-Pb age of 1885 $\pm$ 67 Ma was obtained for Willbob Formation basalts in the southern part of the orogen (Rohon et al., 1993). This age is consistent with the interpretation that the basalts correlate with Hellancourt Formation basalts in the north. The youngest age for the second cycle was obtained on a pyroclastic rhyodacite in the Le Moyne Group and yielded an age of 1870 $\pm$ 4 Ma (Machado et al., 1997). Taken together, ages obtained from second-cycle rocks indicate that deposition occurred over a period of at least 10 Ma (1880-1870 Ma).

Published dates are overlain on a stratigraphic column of the Labrador Trough in Figure 2.4.

Figure 2.4. Stratigraphic column for the Labrador Trough, after Clark and Wares (2006). Age references: Clark (1984); Chevé and Machado (1988); Rohon et al. (1993); Findlay et al. (1995); Machado et al. (1997); Bleeker and Kamo (2018).

2.3 Roberts Lake Syncline

2.3.1 Geological setting of the Roberts Lake Syncline

The Roberts Lake Syncline (Payne Zone of Clark and Wares, 2006) is located between the northern extremity of the New Quebec orogen (Fig. 1.1) and the eastern extremity of the Ungava orogen. It was mapped in detail by Hardy (1976) at a scale of 1:63,360. In this area, the contact between Paleoproterozoic supracrustal rocks and Archean gneisses of the Superior Province may be marked by a thrust fault, indicating that the entire supracrustal sequence represents a klippe. Both the basement and the allochthon were folded into a synclinal structure 20 km wide and 80 km long that plunges gently southeastward (Fig. 2.5). The base of the package is folded into kilometer-scale parasitic folds around the margin of the Roberts Lake Syncline. Schistosity dip moderately steeply to the northeast on the west limb and are subvertical on the east limb of the syncline, indicating that it is slightly overturned toward the west. Metamorphic grade increases from greenschist facies along the west limb to amphibolite facies along the east limb of the syncline (Hardy, 1976).

The Roberts Lake Syncline is rimmed by sedimentary rocks (iron formation, turbidites, sulfidic/graphitic mudstones and minor dolomite) and cored by a thick sequence of basalt containing minor sulfidic/graphitic mudstone. The stratigraphic correlation to formations further south is uncertain and only one large-scale fault structure is evident at the map scale, occurring along the base of a major layered sill complex east of Lac Chaunet. The basaltic pile is intruded by

abundant gabbro sills and by several tabular, undulose mafic-ultramafic complexes composed of layered gabbro-peridotite with minor pyroxenite and diorite.

Figure 2.5. Geological map of the Roberts Lake Syncline. Modified from Hardy (1976). Small map shows the outline of northern Quebec with the Cape Smith belt and the Labrador Trough in grey (after St-Onge et al., 2004; Clark and Wares, 2006; Goodfellow, 2007). Red box indicates the location on Fig. 1.1.

2.3.2 Previous geochronology of the Roberts Lake Syncline

A mineralized, mafic-ultramafic sill located in the Roberts Lake Syncline, termed the Qarqasiaq complex (Mungall, 1998), gave a U-Pb zircon age of 1882 ± 4 Ma (Wodicka et al., 2002), which also falls into the age bracket of Cycle 2 mafic magmatism (basalts and gabbros).

At its southeastern extremity, bordering the coast of Ungava Bay, the Roberts Lake Syncline is intruded by a late syntectonic to post-tectonic (Mungall, 2000) layered mafic-ultramafic intrusion called the Kyak Bay intrusion (Hardy, 1976), dated at 1838.3 ± 3.1 Ma via U-Pb on zircons by Randall (2005). The Kyak Bay intrusion is considered to have been emplaced during the culmination of the collision between the Superior and Rae Provinces (Mungall, 2000), somewhat later than the 1.82-1.77 Ga timing suggested by Wardle et al. (2002) for the climax of convergent tectonics.

Published dates are overlain on a stratigraphic column of the Roberts Lake Syncline in Figure 2.6.

Figure 2.6. Stratigraphic column for the Roberts Lake Syncline, after Hardy (1976). Age references: Wodicka et al. (2002); Randall (2005).

2.4 References

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3 GEOCHRONOLOGY OF THE CAPE SMITH BELT AND THE ROBERTS LAKE SYNCLINE

3.1 Introduction

Several precise U-Pb geochronological ages have been published throughout the Cape Smith belt, but the exact timing and duration of the Povungnituk and Chukotat magmatic events are still unconstrained (see Section 2.1.1). To assess the precise geochronology of magmatism in the Cape Smith belt, U-Pb dating was attempted on several units of the Cape Smith belt. Several coarse grain gabbroic samples from the Povungnituk and Chukotat Group have been selected from the collections of W.R.A Baragar (Geological Survey of Canada) and M.R. St. Onge (Geological Survey of Canada). Thin sections were prepared at Memorial University and were further imaged using the JSM-7100F field emission scanning electron microscope (SEM) from the TERRA facility – CREAT (Memorial University Earth Science Department) in order to identify Zr-bearing phases and to image internal morphologies and textures. Many mafic rocks, such as diabase dykes and gabbro sills, contain trace amounts of primary U-bearing minerals such as baddeleyite (ZrO_2), zircon (ZrSiO_4) and zirconolite ($\text{CaZrTi}_2\text{O}_7$) (Heaman and LeCheminant, 1993). The importance and potential of U-Pb baddeleyite dating has been outlined by several authors (e.g., Krogh et al., 1987; Heaman and LeCheminant, 1993). Its use has increased since separation techniques improved (e.g., Söderlund and Johansson, 2002) and it has become a cornerstone of supercontinent reconstruction (Bleeker, 2003; Bleeker and Ernst, 2006; Ernst and Bleeker, 2010). Dating with magmatic baddeleyite, however, can be complicated for terranes that have seen metamorphism, where zircon and zirconolite overgrowths can yield variably discordant results. In the

Cape Smith belt, greenschist to amphibolite facies metamorphism related to the Trans Hudson orogeny has caused significant overgrowth in which baddeleyites are either poorly preserved or completely destroyed. From the samples where baddeleyite survived, many have secondary zircon overgrowths, preventing dating using dissolved whole grains.

The results of all samples where zirconium-bearing phases could be identified are shown in Table 3.1. Samples, containing baddeleyite without signs of secondary overgrowth, were taken to Lund University for mechanical separation using the Wilfley water-shaking table separation method (Söderlund and Johansson, 2002). Three samples yielded baddeleyite during mechanical separation. From these samples, the first (BLS-73-197) is a mafic sill intruding the central Povungnituk Group. The second sample (BLS-73-31) is from a mafic sill intruding the upper Povungnituk Group. The third sample (BL-73-331) was collected from a large mafic dyke within the Nuvilik Formation that has been interpreted as a feeder dyke to the overlying Chukotat Group, based on field petrography and geochemistry (by W.R.A. Baragar). These samples were analysed via isotope dilution thermal ionization mass spectrometry (ID-TIMS).

In addition, two samples (BL-73-M260, BL-73-180) that had larger baddeleyite grains ($> 20 \mu\text{m}$), but showed zircon overgrowths were chosen for in-situ analyses using secondary ion mass spectrometry (IN-SIMS) using the technique described in Chamberlain (2010). Sample BL-73-M260 was collected from the central Povungnituk Group and sample BL-73-180 was collected from the top of the Chukotat Group.

Two additional samples (BL-73-M326, SAB-87-D273A) of dykes that intrude into the Dumas Formation were dated using laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) analyses. One of these samples contains very few baddeleyite grains with zircon overgrowth (Sample BL-73-M326) and the other contains only zircon grains (Sample SAB-87-273A).

Another question that is still debated is the relationship between the two major sedimentary horizons within the southern Cape Smith belt (St-Onge and Lucas, 1993). Constraints on a minimum age on the lower sedimentary horizon, which comprises the Dumas Formation, are possible due to the intrusion of the Korak sill, which has an age of 2038 ± 3 Ma (Machado et al., 1993). St-Onge and Lucas (1993) suggested that the sedimentary panel of the Nuvilik Formation represents a separate thrust slice of Dumas Formation sedimentary rocks deposited on top of the basaltic rocks of the Beuparlant Formation. An alternative interpretation is the Nuvilik Formation represents the final, uppermost unit of the Povungniut Group when waning magmatism and post-magmatic cooling led to subsidence and deposition of mostly distal turbidites across the belt (Mungall, 2007; Bleeker and Ames, 2017). These authors argue that minor thrusts and fold repetitions are likely but there is no compelling evidence for a major thrust at the base of the Nuvilik Formation. If the Nuvilik Formation represents a horizon younger than the lower Povungnituk Group, potential detrital zircon grains with ages of 2040 Ma and younger could be recovered from sedimentary rocks within the Nuvilik Formation. Therefore, a turbidite layer from

the Nuvilik Formation was sampled and detrital zircon grains were separated and analysed at the Jack Satterly Geochronology Laboratory (University of Toronto).

As stated in Section 2.3, the Roberts Lake Syncline in the northernmost Labrador Trough might represent a continuation of the Cape Smith belt. Therefore, sedimentary horizons should be correlative to the Cape Smith belt. To further test this hypothesis, two quartzite samples from different parts of the stratigraphy within the Roberts Lake Syncline were collected and processed for detrital zircon separation at the University of Toronto.

All sample locations are shown in Figure 3.1.

Table 3.1. Scanning electron microscope results of zirconium-bearing phases for samples of the Cape Smith belt

Stratigraphic unit	Locality	Sample	Number of baddeleyite imaged	biggest grain (µm)	smallest grain (µm)	Number of zircon imaged	biggest grain (µm)	smallest grain (µm)
Chukotat Group	K4-04	BL-73-179	3	16x24	20x16	16	80x76	30x25
Chukotat Group	K4-04	BL-73-180	7	28x76	24x28	2	76x76	36x32
Chukotat Group	K4-08	BL-73-185	0	/	/	28	40x44	24x32
Chukotat Group		SAB-87-L280	1	8x4	8x4	0	/	/
Nuvilik Fm	K2-02	BL-73-017	4	20x48	20x16	3	28x32	24x16
Nuvilik Fm	N2-17	BL-73-326	0	/	/	1	40x44	40x44
Nuvilik Fm	N2-17	BL-73-331	4	56x44	24x24	5	116x104	40x36
Nuvilik Fm		SAB-87-S186	1	24x16	24x16	7	40x52	20x16
Nuvilik Fm		SAB-86-D236S	11	40x36	24x16	7	84x76	24x24
Nuvilik Fm		SAB-86-D236H	2	44x32	24x28	2	40x48	24x28
Beauparlant Fm	K1-04	BLS-73-31	7	76x28	44x40	7	32x24	24x20
Beauparlant Fm	C1-41	BLS-73-183	2	16x16	12x12	2	40x28	24x36
Beauparlant Fm	N5-10	BL-73-M326	6	152x92	16x24	12	60x28	16x32
Beauparlant Fm	N5-15	BL-73-M260	2	60x92	16x12	3	24x116	25x25
Beauparlant Fm	N5-15	BL-73-M261	0	/	/	13	68x76	36x32
Dumas Fm		SAB-87-D273A	0	/	/	3	116x128	48x76
Dumas Fm		SAB-87-D273B	0	/	/	12	88x108	28x20
Dumas Fm		SAB-87-D279A	0	/	/	7	56x36	24x24
Dumas Fm		SAB-87-D283	0	/	/	3	24x24	20x16

Figure 3.1. Map of the Cape Smith belt (after St-Onge et al., 2004) and the northernmost part of the Labrador Trough, including the Roberts Syncline (with nomenclature after Clark and Wares, 2006). Sample locations from previous geochronological work and this thesis are shown as circles and numbers correlated to the legend. Overview map of North America includes the location of the Superior craton in grey (Goodfellow, 2007) and the two map areas in red outline. Age references: Parrish (1989); Machado et al. (1993); Wodicka et al. (2002); Randall (2003); Bleeker and Kamo (2018).

3.2 U-Pb ID-TIMS on magmatic baddeleyite

3.2.1 Samples

3.2.1.1 BLS-73-31

Sample BLS-73-31 is from the lower part of a dolerite sill at the very top of exposed western Povungnituk Group lavas, located at 61.156536 N 76.934639 W. It is separated from the Chukotat Group lavas to the north by a highly contorted shale unit that has been interpreted by W.R.A. Baragar (2014 pers. comm.) to mark the presence of a thrust fault in close proximity.

Major rock forming minerals in sample BLS-73-31 include plagioclase (25 %) and amphibole (45 %). The remaining 30 % of the sample includes brown radial aggregate rims surrounding plagioclase and amphibole, fine matrix of saussurite, chlorite and actinolite, and accessory phases of titanite, apatite and baddeleyite. With a medium-grained ophitic texture, the gabbro includes grains up to 5 mm (Fig. 3.2). Plagioclase is up to 3 mm long, mostly preserved and shows saussurite alteration. Amphiboles, up to 1 mm, show remnant ophitic texture. Titanite forms 0.5 mm wide grains and is distributed throughout the sample. Baddeleyite grains separated from this sample are pale brown, mildly translucent, prismatic to platy fragments, ~20-40 microns in width and up to 80 microns in length that do not exhibit incipient zircon growth on their surfaces.

Figure 3.2. Representative thin section photomicrograph for sample BLS-73-31 in plane polarized light (left) on cross polarized light (right).

3.2.1.2 BLS-73-197

Sample BLS-73-197 is a fine to medium grained gabbro from either a dolerite sill or massive flow from the lower to middle part of the Beuparlant Formation, located at 61.382381 °N 74.650412 °W. It is located within a segment composed of interlayered greywacke-shales and thin (5 to 20 m thick) dolerite sills or massive lava flows. Some of the dolerite/basalt layers have been identified as sills, whereas the identity of other intrusive bodies remains uncertain. The matrix of the sample makes up 60 % of the rock and is foliated and composed of fine grained saussurite, chlorite, and actinolite. Large amphibole crystals (2 mm) make up 30 % of the rock. They show faint remnants of ophitic texture and some (<10 %) small (0.2 mm) subhedral orthopyroxene crystals are still present (Fig. 3.3). The sample contains no identifiable plagioclase. Baddeleyite grains separated from this sample are dark brown, mildly translucent, platy fragments, ~20-40 microns in width and length that do not exhibit incipient zircon growth on their surfaces.

Figure 3.3. Representative thin section photomicrograph for sample BLS-73-197 in plane polarized light (left) on cross polarized light (right).

3.2.1.3 BL-73-331

Sample BL-73-331 is a medium grained gabbroic sample (Fig. 3.4) from the pegmatitic zone of a differentiated sill at the base of the Chukotat Group, located at 61.561563 N 74.676763 W. It has an olivine-rich base with the majority of the sample (60 %) composed of up to 2 mm long amphiboles. Plagioclase has been completely altered to saussurite that represent 20 % of the sample but still show plagioclase pseudomorphs. Fibrous aggregates of chlorite and actinolite represent the remaining 20 % of the sample. Baddeleyite grains separated from this sample are typical dark brown, translucent, platy fragments, ~20-40 microns in width and length that do not exhibit incipient zircon growth on their surfaces.

Figure 3.4. Representative thin section photomicrograph for sample BL-73-331 in plane polarized light (left) on cross polarized light (right).

3.2.2 Analytical procedure

Baddeleyite from the three samples was separated at Lund University and the University of Toronto Jack Satterly Geochronology Laboratories. U-Pb ID-TIMS dating was carried out at the University of Toronto Jack Satterly Geochronology Laboratories.

At the Lund University laboratory, ~500 g of sample was crushed into cm-size pieces using a sledgehammer and placed in a mill tray to produce a fine powder. The suspended sample was divided into ~20-40 g portions and loaded on a Wilfley water-shaking table following the procedures in Söderlund and Johansson (2002). After 2-3 cycles in which suspended powder was loaded onto the shaking table and the lighter material washed away, the remaining heavier fraction was transferred into a small glass petri dish. Magnetic minerals were removed using a pencil magnet. Baddeleyite grains were hand-picked under a binocular microscope and the best quality grains were selected for isotopic analysis. The selected baddeleyite grains were dark brown, crystalline, and with no trace of alteration (i.e., no visible frosty zircon rims).

Additional baddeleyite grains were also separated at the Jack Satterly Geochronology Laboratory of the University of Toronto using a similar procedure to that of Lund University (see above). The mineral fractions chosen for ID-TIMS analysis at the Jack Satterly Geochronology Laboratory at the University of Toronto comprised between 2 and 7 grains of baddeleyite.

The selected baddeleyite crystals were given a brief wash in HNO₃, spiked using a mixed ²⁰⁵Pb–²³⁵U isotopic tracer, and dissolved in a 12:1 volume of HF-

HNO₃ acid in Teflon® bombs at 200°C (Krogh, 1973). Samples were redissolved in 3N HCl in preparation for anion exchange chemistry, dried down with dilute phosphoric acid and loaded with silica gel onto degassed rhenium filaments. Isotopic compositions of Pb and U were determined on a M354 mass spectrometer using a Daly detector in digital pulse-counting mode. A Daly mass fractionation correction of 0.05 % per atomic mass unit (AMU), and a thermal mass discrimination correction of 0.1 % per AMU for Pb and U, was applied. Dead-time of the counting system during the analytical interval was 16 ns based on measurements of the NBS SRM 982 Pb standard. Laboratory procedural blanks at the Jack Satterly Geochronology Laboratory are routinely at the 0.5 pg and 0.1 pg level or less for Pb and U, respectively.

3.2.3 Results

3.2.3.1 BLS-73-31

Results of U-Pb ID-TIMS analysis of five fractions of baddeleyite (B1 to B5) from dolerite sill sample BLS-73-31 are summarized in Table 3.2 and plotted in Figure 3.5. The U-Pb data are variably discordant and define a Pb loss array (red ellipses in Fig. 3.5) that intersects the concordia curve at 2004 ± 24 Ma and 1303 ± 410 Ma, with an MSWD of 2.0. Fraction B3 plots below the linear array and if omitted from the regression line calculation results in reduced scatter (MSWD = 0.17), and upper and lower intercept ages of 1998 ± 20 Ma and 1251 ± 420 Ma.

The baddeleyite grains were apparently affected by variable degrees of ancient Pb loss (for example, from Raglan events ca. 1881-1882 Ma and younger metamorphism), perhaps in addition to minor later Pb loss during final uplift and exposure. The apparent lower intercepts likely result from time-averaged Pb loss and are not considered geologically significant. Anchoring the Pb loss line at ca. 1250 Ma does not affect the calculated age and yields 1998 ± 6 Ma (Fig. 3.5), which is interpreted to represent a maximum age for the emplacement of the dolerite sill.

A regression line through fractions B1 and B3 defines a steeper Pb loss pattern defining an upper intercept of 1988 ± 10 Ma, which can be interpreted as a minimum age estimate for the dolerite. As fractions B1 and B3 are less discordant and less affected by earlier open system behaviour (in contrast to fractions B4 and B5), the corresponding Pb loss line may provide a more precise

upper intercept age (e.g., Bleeker, 2014). The $^{207}\text{Pb}/^{206}\text{Pb}$ age of least discordant fraction B1, at 1986 ± 5 Ma, defines an absolute minimum age.

An age interpretation of 1998 ± 6 Ma, or one intermediate between 1988 Ma and 1998 Ma (based on the above discussion), is in agreement with the earlier age determination for Povungnituk Group magmatism of 1991 ± 2 Ma (Machado et al., 1993). This dated sample is interpreted (by W.R.A. Baragar) as a sill or coarse-grained flow. If it is a sill, then strictly speaking this U-Pb date is a minimum age for the age of the flood basalts. However, the geochemistry of this sill (and of other similar sills/coarse-grained flows) matches the chemistry of the flows, and therefore confirms this U-Pb age as representing an age for the flood basalts (specifically the enriched magma type).

Figure 3.5. U-Pb concordia diagram for dolerite sample BLS-73-31 within the Povungnituk Group volcanic sequence, showing results for five non-abraded baddeleyite fractions (red ellipses, B1 to B5). Small image shows separated baddeleyite fractions for ellipses B1, B2, B4, and B5, from left to right. The results are variably discordant along a Pb loss line (B1 is least discordant at 0.8 % from its $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1985 Ma). The upper intercept through four data points, at 1998 ± 6 Ma, represents a reasonable age interpretation. However, the true crystallization age may be slightly younger, intermediate between 1998 ± 6 Ma and 1988 ± 10 Ma, a minimum age estimate based on B1 and B3. The age is in agreement with zircon data (grey ellipses, shown for reference) on a small granodiorite intrusion in pillow basalts of the Povungnituk Group (Machado et al., 1993).

3.2.3.2 BLS-73-197

Results of U-Pb ID-TIMS analysis of three fractions of baddeleyite (B1 to B3) from dolerite sill sample BLS-73-197 are summarized in Table 3.2 and plotted in Figure 3.6. The U-Pb data are near concordant ranging between 0.8 and 2.1% discordant. The three fractions are included in a regression calculation anchored at 0 Ma that has an upper intercept of 1967 ± 7 Ma, with an MSWD of 2.3.

Figure 3.6. U-Pb concordia diagram for dolerite sample BLS-73-197 within the Beauparlant Formation, showing results for three non-abraded baddeleyite fractions (red ellipses). The results are concordant, or near concordant and have a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1967 ± 7 Ma. Regression line anchored to 0 Ma intersects the concordia at a similar age.

3.2.3.3 BL-73-331

Results of U-Pb ID-TIMS analysis of six fractions of baddeleyite (B1 to B6) from dolerite sill sample BLS-73-331 are summarized in Table 3.2 and plotted in Figure 3.7. The U-Pb data are near concordant. An upper intercept line, with the lower intercept anchored at 0 Ma, drawn through four fractions (B1 to B4, red ellipses) has an upper intercept of 1874 ± 3 Ma, with an MSWD of 0.65.

Figure 3.7. U-Pb concordia diagram for dolerite sample BL-73-331 within the Nuvilik Formation, showing results for six non-abraded baddeleyite fractions (red ellipses). The four results that are most concordant and precise have a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1874 ± 3 Ma. A regression line anchored at 0 Ma intersects the concordia at a similar age.

Table 3.2. U-Pb isotopic results for ID-TIMS

No.	U (ppm)	Th/U	Pb _{total} (pg)	PbC (pg)	²⁰⁶ Pb/ ²⁰⁴ Pb meas.	²⁰⁷ Pb/ ²³⁵ U	2 σ	²⁰⁶ Pb/ ²³⁸ U	2 σ	Error Corr.	²⁰⁷ Pb/ ²⁰⁶ Pb	2 σ	²⁰⁶ Pb/ ²³⁸ U Age (Ma)	2 σ	²⁰⁷ Pb/ ²³⁵ U Age (Ma)	2 σ	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	2 σ	% Disc.
BLS-73-31 - 61.156536 N 76.934639 W																			
B1	1767	0.16	19.0	0.6	2222	6.020	0.025	0.3579	0.0011	0.794	0.1220	0.0003	1972	5	1979	4	1986	5	0.8
B2	720	0.25	15.7	0.3	2992	5.933	0.024	0.3546	0.0011	0.839	0.1213	0.0003	1957	5	1966	3	1976	4	1.1
B3	160	0.20	11.3	0.2	3651	5.944	0.027	0.3540	0.0015	0.921	0.1218	0.0002	1954	7	1968	4	1982	3	1.7
B4	582	0.48	17.8	0.6	1777	5.883	0.027	0.3524	0.0012	0.792	0.1211	0.0003	1946	6	1959	4	1972	5	1.5
B5	416	0.23	11.8	1.4	551	5.800	0.078	0.3489	0.0019	0.651	0.1206	0.0013	1929	9	1946	12	1965	19	2.1
BLS-73-197 - 61.382381 °N 74.650412 °W																			
B1	330	0.77	1.4	0.1	684	6.00	0.24	0.3571	0.0138	0.964	0.1218	0.0013	1969	66	1975	35	1982	19	0.8
B2	179	0.16	3.2	1.1	211	6.01	0.18	0.3565	0.0047	0.663	0.1223	0.0028	1966	22	1978	26	1990	41	1.4
B3	197	0.61	3.8	0.2	1265	5.790	0.069	0.3485	0.0038	0.935	0.1205	0.0005	1928	18	1945	10	1964	8	2.1
BLS-73-331 - 61.561563 N 74.676763 W																			
B1	31	0.14	5.2	0.2	2236	5.304	0.041	0.3352	0.0024	0.931	0.1147	0.0003	1864	12	1869	7	1876	5	0.7
B2	30	0.05	2.0	0.8	173	5.258	0.239	0.3347	0.0063	0.627	0.1139	0.0042	1861	31	1862	39	1863	67	0.1
B3	14	0.19	2.4	0.3	559	5.267	0.100	0.3346	0.0057	0.842	0.1142	0.0012	1861	28	1863	16	1867	19	0.4
B4	60	0.09	10	0.3	2408	5.258	0.027	0.3328	0.0015	0.868	0.1146	0.0003	1852	7	1862	4	1873	5	1.3
B5	26	0.31	4.4	0.3	895	5.214	0.056	0.3296	0.0028	0.850	0.1147	0.0007	1837	14	1855	9	1876	10	2.4
B6	10	0.25	2.2	0.5	290	5.034	0.129	0.3222	0.0052	0.728	0.1133	0.0020	1801	25	1825	22	1853	32	3.2

Th/U calculated from radiogenic ²⁰⁸Pb/²⁰⁶Pb ratio and ²⁰⁷Pb/²⁰⁶Pb age assuming concordance.

Pb_{total} is total amount of Pb excluding blank.

PbC is total amount of common Pb in picograms; assigned the isotopic composition of laboratory blank.

²⁰⁶Pb/²⁰⁴Pb corrected for fractionation and common Pb in the spike.

Pb/U ratios corrected for fractionation, common Pb in the spike, and blank.

Correction for ²³⁰Th disequilibrium in ²⁰⁶Pb/²³⁸U and ²⁰⁷Pb/²⁰⁶Pb assuming Th/U of 4.2 in the magma.

Error Corr. is correlation coefficients of X-Y errors on the concordia plot.

Disc. is percent discordance for the given ²⁰⁷Pb/²⁰⁶Pb age.

3.3 U-Pb SIMS on magmatic baddeleyite and zircon

3.3.1 Samples

3.3.1.1 BL-73-180

Sample BL-73-180 is from the upper margin of a 200 m thick sill, located at 61.394154 N 76.977077 W, and is a medium to coarse-grained gabbro. The sill intrudes upper levels of what is interpreted to be the tholeiitic plagioclase-phyric upper part of the generally komatiitic Chukotat Formation. It is only a few 10's of metres below the major thrust that marks the northern boundary of the Cape Smith Belt. Therefore, the specimen provides minimum age constraints for the Chukotat Formation. The sample is composed to 30 % of 2 to 5 mm large ophitic orthopyroxenes and amphibole, the latter is a secondary phase replacing pyroxene that enclose plagioclases smaller than 0.5 mm (Fig. 3.8a/b). Plagioclase is completely altered to saussurite but still exhibits igneous pseudomorphs. The matrix is composed of saussuritised plagioclase and patches of chlorite and actinolite. Small (<0.5 mm) rounded blebs of serpentine making up less than 5 % of the sample indicate the presence of previous olivine. Titanite is an accessory phase.

This sample is relatively poor in baddeleyite and zircon with only about 60 targets identified in a single thin section, although many of them are large, up to 100 microns, and concentrated in randomly dispersed regions. Approximately 1/6 of the identified targets are baddeleyite (Fig. 3.8c), 1/3 of the targets are zircon and the rest are mixed phase grains ranging from relatively thin zircon rims on

large baddeleyite cores (e.g., Fig. 3.8d) to nearly complete zircon surrounding minor baddeleyite cores. Where possible, single-phase grains were analysed preferentially, and, when overgrowths were present, pure baddeleyite portions of the grains were targeted.

Figure 3.8. Photomicrographs in plane polarized light (a) and cross polarized light (b), as well as back-scattered electron (BSE) images of typical baddeleyite from sample BL-73-180. (c) Typical grains from sample BL-73-180 with sharp boundaries and no zircon rims. (d) However, many baddeleyite grains in sample BL-73-180 have zircon fringes and rims (pale grey on lower image d). Analyses targeted the portions of pure baddeleyite and limited the analyzed regions to 3-4 microns whenever possible. The data from both baddeleyite and individual, pure zircon grains yield the same age of ca. 1861 Ma, so the zircon fringes are interpreted to reflect a late magmatic increase in silica activity.

3.3.1.2 *BL-73-M260*

Sample BL-73-M260 is a medium-grained gabbroic sample from a sill located at 61.385387 N 74.649948 W that intrudes a unit of interlayered basaltic volcanic rocks and dark shales and greywackes of the Povungnituk Group. This unit is near the base of the deep-water, magmatic-dominant part of the Povungnituk Group and overlies a continental shelf facies that adjoins the Archean basement. The sample is composed mainly of 2 mm plagioclase, which makes up 60 % of the mineral assemblage (Fig. 3.9a/b), that show only minor sericite alteration. Euhedral clinopyroxene grains are up to 1 mm large and make up 10 % of the sample. Opaque minerals (0.5 - 1 mm) comprises 10 % of the sample and are probably Fe-Ti-oxides. The remaining 20 % of the sample is composed of amphibole (1 mm), recrystallized after pyroxene.

This sample is extremely rich in baddeleyite grains with over 250 grains located in a single thin section. Zircon grains are much scarcer, with only 15 grains located. Baddeleyite grains are unaltered, no zircon fringes were observed (e.g., Fig. 3.9c). Zircon grains appear to have grown later along fractures and fluid pathways (e.g., Fig. 3.9d).

Figure 3.9. Photomicrographs in plane polarized light (a) and cross polarized light (b), as well as back-scattered electron (BSE) images of typical baddeleyite from sample BL-73-M260. (c) Typical baddeleyite from sample BL-73-M260. (d) Typical zircon from BL-73-M260. All the baddeleyite grains have smooth internal textures and no evidence of zircon overgrowths or alteration. The zircon grains in this sample (less than 10% of zirconium phases) occur along fractures and fluid pathways and reflect late, localized alteration ca. 330 Ma (as discussed later), which did not appear to affect the baddeleyite grains.

3.3.2 Analytical procedure

Zirconium-bearing phases were located in polished thin sections by wavelength dispersive spectroscopy (WDS, University of Wyoming). Each map required up to 10 hours of automated instrument time per section to locate the ~10 to 30 micron diameter target minerals. Manual energy dispersive spectroscopy (EDS) was used to identify the zirconium-bearing mineral phases. A series of back-scattered electron (BSE) and reflected light images at various magnifications were prepared to facilitate relocating the targets for isotopic analyses. Target-rich portions of the thin sections up to 10 mm squared, were mounted in 2.54 cm diameter, epoxy disks along with pre-polished standard grains for U and Pb isotopic analysis by secondary ion mass spectrometry (SIMS). SIMS analysis followed the methods described in Schmitt et al. (2009) and Chamberlain et al. (2010) using the CAMECA 1270ims at UCLA. The field aperture was matched to grain dimensions to screen out ions from host phases, especially common Pb. Sampling regions were generally limited to 3-4 microns of the 20-micron primary pit. Percent radiogenic ^{206}Pb and Th/U values were used to screen the data (Table 3.3). Any analyses with radiogenic values below 98% often reflect altered grains, or mixed phases, especially if the Th/U is greater than 0.2.

Baddeleyite data from BL-73-M260 had significant amounts of common Pb (up to 68 %, Table 3.4). A total-Pb (radiogenic and common), linear regression strategy was used to estimate the ages of the minerals using a 3-dimensional Tera-Wasserburg (1972) concordia approach. This strategy minimizes the

potential effects of poorer precision on the ^{204}Pb determination and avoids potential bias introduced by the choice of common Pb isotopic compositions, although it assumes a single age for the selected populations. Discordance on total-Pb concordia plots is largely due to the amount of common Pb in the analyses; inheritance cannot be easily identified. Common Pb-corrected $^{206}\text{Pb}/^{238}\text{U}$ dates were also calculated for BL-73-M260 using the ^{204}Pb method to test whether age variation existed.

Data reduction and calculation of slopes and intercepts used the in-house ZIPS program at UCLA and ISOPLOT/EX (based on Ludwig, 1991), respectively.

3.3.3 Results

3.3.3.1 BL-73-180

Fourteen baddeleyite grains and ten zircon grains were analysed from this sample. Many of the baddeleyite grains have fringes of zircon, these were avoided whenever possible to try to determine the magmatic date of the baddeleyite. Uranium concentrations varied from 100 to 3000 ppm for baddeleyite; most were less than 1000 ppm (Table 3.3). Th/U values were all 0.2 or less, with two exceptions. Uranium concentrations in zircon grains were considerably higher, from 800 to 12,000 ppm. Both baddeleyite and zircon analyses were nearly free of common Pb, with less than 5% in all but 2 analyses and many had less than 1%. The data from baddeleyite, zircon, and baddeleyite grains with thin zircon rims yield similar ages; all plot on a single chord with an upper intercept date of 1861 ± 28 Ma (Fig. 3.10), so growths of both phases are interpreted as magmatic. The zircon data in this sample is more discordant. The best magmatic age estimate for this rock is the weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ baddeleyite date of 1861 ± 28 Ma (Fig. 3.10).

Figure 3.10. (top) Concordia plot of in-situ SIMS baddeleyite and zircon analyses from BL-73-180. Linear regression includes all data from both phases and yields an age of $1861 \pm 21 \text{ Ma}$. (bottom) Weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date from in-situ SIMS baddeleyite analyses from BL-73-180.

Table 3.3. U-Pb SIMS results for sample BL-73-180

Sample	grain size (microns)	Ages (Ma)							U conc. ppm	Ratios				rho	U O2/ U	U O/ U	Th/ U
		²⁰⁶ Pb/		²⁰⁷ Pb/		²⁰⁷ Pb/		r ²⁰⁶ Pb		²⁰⁷ Pb*/		²⁰⁶ Pb*/					
		²³⁸ U	1 σ	²³⁵ U	1 σ	²⁰⁶ Pb	1 σ			%	²³⁵ U	1 σ %	²³⁸ U				
BL-73-180 - 61.394154 N 76.977077 W																	
53bz	24x12	1977	124	1916	64	1851	24	99.4	474	5.600	(7)	0.359	(7)	0.984	7.1		0.18
31bz	32x24	2076	174	1978	86	1878	20	99.3	401	6.017	(10)	0.380	(10)	0.994	5.8		0.14
39b	11x3	2363	243	2116	111	1884	44	99.2	1032	7.036	(13)	0.443	(12)	0.980	5.2		0.64
29bz	19x15	2050	200	1957	100	1859	31	98.9	357	5.868	(12)	0.375	(11)	0.989	5.3		0.20
58bz	30x15	2071	110	1954	56	1833	24	98.8	491	5.853	(6)	0.379	(6)	0.979	8.0		0.15
37bz ^x	18x18	5777	3240	3215	826	1858	41	98.8	878	22.710	(85)	1.450	(85)	1.000	2.6		0.12
52bz	33x28	2084	140	2010	74	1934	55	97.9	284	6.237	(8)	0.382	(8)	0.932	6.8		0.15
51bz	23x17	1866	105	1832	62	1794	60	97.8	390	5.077	(7)	0.336	(6)	0.895	7.9		0.25
23b	25x5	2205	138	2021	85	1839	105	96.6	157	6.322	(10)	0.408	(7)	0.804	7.3		0.15
56bz	54x10	2047	235	1979	125	1910	73	96.6	388	6.025	(14)	0.374	(13)	0.959	4.9		0.11
48b	12x11	1998	77	1973	72	1947	111	96.4	117	5.979	(8)	0.363	(4)	0.674	11.2		0.14
50bz	21x14	2012	146	1889	106	1757	142	95.9	178	5.427	(12)	0.366	(8)	0.785	6.6		0.21
59bz	12x9	1468	112	1576	87	1723	111	94.9	2688	3.720	(11)	0.256	(9)	0.836	6.4		3.64
32bz	8x2	1852	129	1868	154	1885	257	82.7	650	5.294	(18)	0.333	(8)	0.646	7.0		0.25
42z	58x30	1376	107	1568	69	1837	11	99.8	2068	3.686	(9)	0.238	(9)	0.998		7.5	6.81
38z	5x3	1298	75	1320	250	1355	589	84.9	847	2.668	(34)	0.223	(6)	0.589		11.6	2.35
35z@1	82x68	1209	48	1412	34	1734	16	100.0	2429	3.019	(4)	0.206	(4)	0.981		9.6	3.10
41z	36x21	1183	84	1416	59	1785	4	100.0	5321	3.032	(8)	0.202	(8)	1.000		7.7	2.51
47z	5x4	1177	46	1410	57	1782	102	99.2	1974	3.009	(8)	0.200	(4)	0.684		10.4	0.45
35z@2	82x68	960	38	1173	33	1591	26	99.8	2605	2.176	(5)	0.161	(4)	0.958		9.7	3.83
36z	27x16	915	41	1171	36	1681	26	99.6	4373	2.170	(5)	0.153	(5)	0.961		9.4	3.38
44z	55x18	855	34	1141	80	1735	179	98.7	2880	2.077	(12)	0.142	(4)	0.594		10.1	2.78
40z	100x58	793	60	1069	58	1685	43	99.7	12569	1.866	(9)	0.131	(8)	0.965		7.7	2.67
43z	33x19	545	30	691	32	1199	56	99.8	11516	0.974	(6)	0.088	(6)	0.895		8.5	3.05

values in parentheses are absolute errors at one sigma level for ages, percent for ratios.

b = baddeleyite, z = zircon, bz = baddeleyite rimmed with zircon, ^x excluded from concordia intercept calculation.

^{r²⁰⁶Pb} = radiogenic ²⁰⁶Pb in percent.

* = radiogenic Pb value corrected for initial Pb using ²⁰⁴Pb method.

rho = correlation coefficient of error ellipses.

3.3.3.2 BL-73-M260

Twenty-six baddeleyite grains and three zircon grains were analysed from this sample. Uranium concentrations are quite variable and relatively high, 151 ppm to 3770 ppm for baddeleyite, 1900 to 8300 ppm for zircon. Th/U values for baddeleyite are quite variable as well, from 0.1 to 3.7.

Due to the high proportion of common Pb, a 3-dimensional, total Pb, Tera-Wasserburg (1972) concordia approach was used (Fig. 3.11), which keeps the common Pb in the concordia coordinates and solves for the common Pb isotopic composition. The lower intercept date from this approach is 819 ± 36 Ma (95% confidence) with common Pb compositions that are reasonable for a Neoproterozoic sample. Zircon data are distinctly younger with a lower intercept of 339 ± 39 Ma. Common Pb-corrected $^{206}\text{Pb}/^{238}\text{U}$ weighted mean dates overlap this intercept date within error (Fig. 3.11). The age of 819 ± 36 Ma is interpreted as the magmatic age of this mafic rock. The zircon grains are minor within the rock and seem to be limited to fractures and other fluid pathways, so 339 Ma is interpreted as a time of alteration leading to lead loss.

Figure 3.11. (top) Total-Pb, 3-dimensional Tera-Wasserburg Concordia linear regressions (Ludwig, 1991) for baddeleyite and zircon in-situ SIMS data from BL-73-M260; the third dimension is $^{204}\text{Pb}/^{206}\text{Pb}$. Baddeleyite analyses were relatively high in common Pb (2 to 32 %, Table 3.4), so a total Pb approach was deemed to be the most robust. Calculated common Pb isotopic compositions are reasonable for Neoproterozoic rocks. Zircon data may relate to minor, localized fluid flow within the rock. (bottom) Common Pb-corrected $^{206}\text{Pb}/^{238}\text{U}$ dates for baddeleyite (red) and zircon (blue) from BL-73-M260. Three baddeleyite analyses were rejected as outliers (brown).

Table 3.4. U-Pb SIMS results for sample BL-73-M260

Sample	grain size (μm)	Age (Ma)		²⁰⁶ Pb %	U conc. ppm	Ratios (Total Pb)											
		²⁰⁶ Pb/ ²³⁸ U	1 σ			²⁰⁶ Pb	U conc. ppm	²³⁸ U/ ²⁰⁶ Pb	1 σ %	²⁰⁷ Pb/ ²⁰⁶ Pb	1 σ %	²⁰⁴ Pb/ ²⁰⁶ Pb	1 σ %	U O2/ U	U O/ U	Th/ U	
BL-73-M260 - 61.385387 N 74.649948 W																	
10b	20x7	782	(46)	98.5	1353	7.634	(6)	0.078	(2)	0.0008	(19)	8.544				0.12	
31b*	19x5	688	(36)	97.8	1011	8.688	(5)	0.080	(2)	0.0011	(14)	8.828				0.08	
20b	15x3	785	(27)	97.3	1420	7.513	(4)	0.091	(3)	0.0014	(21)	13.41				0.11	
7b	10x3	877	(50)	96.8	808	6.640	(6)	0.084	(3)	0.0017	(27)	8.859				0.25	
68b@1*	39x7	678	(54)	96.0	659	8.651	(8)	0.091	(3)	0.0021	(13)	6.534				0.14	
67b	20x5	1002	(120)	95.4	3776	5.675	(13)	0.102	(7)	0.0024	(18)	5.111				0.38	
23,24b	36x5	959	(82)	94.9	1212	5.917	(9)	0.108	(4)	0.0027	(10)	6.127				0.55	
27b	16x6	930	(50)	92.1	1014	5.931	(6)	0.125	(2)	0.0042	(11)	12.3				0.67	
6b*	14x6	1306	(356)	91.7	1662	4.083	(30)	0.116	(8)	0.0044	(21)	3.518				0.17	
13b	17x9	698	(69)	89.9	1797	7.862	(10)	0.143	(6)	0.0054	(12)	5.782				0.17	
2b	25x5	845	(43)	89.6	1726	6.394	(5)	0.140	(1)	0.0055	(10)	9.271				0.45	
19b	9x5	765	(39)	87.4	1557	6.935	(5)	0.152	(3)	0.0067	(9)	9.303				0.79	
21b	11x2	982	(117)	85.3	2370	5.181	(12)	0.153	(11)	0.0078	(30)	6.157				0.14	
5b	12x11	834	(65)	84.8	1108	6.139	(7)	0.160	(4)	0.0081	(27)	8.192				1.23	
16b	7x4	736	(57)	84.3	471	6.969	(6)	0.176	(4)	0.0083	(31)	11.63				0.36	
1b@1 ^Y	60x5	952	(38)	84.2	535	5.288	(4)	0.164	(6)	0.0084	(9)	13.08				0.20	
69b	23x6	901	(73)	82.8	651	5.516	(8)	0.198	(6)	0.0091	(16)	6.79				0.43	
66b	20x7	833	(34)	80.9	630	5.865	(4)	0.203	(3)	0.0101	(8)	13.76				0.32	
68b@2	39x7	726	(79)	77.2	576	6.477	(11)	0.220	(5)	0.0121	(14)	5.881				0.06	
30b	5x3	871	(185)	74.4	718	5.147	(22)	0.218	(3)	0.0136	(21)	3.909				0.19	
29b	13x6	819	(89)	72.4	1637	5.350	(11)	0.264	(2)	0.0146	(10)	5.471				0.11	
25b	9x5	772	(75)	69.6	972	5.476	(9)	0.292	(3)	0.0161	(12)	6.507				0.29	
26b	20x4	769	(39)	68.5	155	5.408	(3)	0.267	(3)	0.0167	(9)	27.73				3.69	
17b* ^Y	14x3	491	(69)	62.4	151	7.880	(4)	0.297	(2)	0.0199	(23)	17.36				0.38	
1b@2	60x5	868	(118)	49.9	1251	3.458	(13)	0.412	(4)	0.0266	(7)	5.385				0.39	
18b	5x3	516	(149)	32.1	555	3.857	(19)	0.415	(2)	0.0360	(11)	4.264				0.45	
15z	6x6	388	(39)	92.3	1938	16.044	(6)	0.101	(4)	0.0041	(24)			9.5		3.25	
14z	13x7	460	(73)	75.3	8278	11.186	(7)	0.247	(2)	0.0131	(14)			8.0		6.13	
8z	33x18	308	(61)	96.2	6604	21.858	(9)	0.088	(2)	0.0020	(7)			7.5		2.20	

values in parentheses are absolute errors at one sigma level for ages, percent for ratios.

b = baddeleyite, z = zircon, @_ = multiple spots on same grain, * = excluded from weighted mean age calculations,

^Y = excluded from concordia calculations.

$r^{206}\text{Pb}$ = radiogenic ^{206}Pb in percent.

$^{206}\text{Pb}/^{238}\text{U}$ ages corrected for common Pb using ^{204}Pb method.

ratios = total Pb including radiogenic and common.

3.4 U-Pb LA-ICP-MS on magmatic baddeleyite and zircon

3.4.1 Samples

3.4.1.1 BL-73-M326

Sample BL-73-M326 is a medium grained gabbroic sill with an ophitic texture, located in the southern part of the Beauparlant Formation at 61.408167 N 74.624238 W. The majority of the sample is made up of 5 mm large clinopyroxene grains that are variably recrystallized to amphibole. The small plagioclase crystals (0.5 mm) enclosed by pyroxene crystals show only minor alteration. The matrix, making up 30 % of the sample is completely recrystallized and altered and is composed of saussurite, chlorite and actinolite aggregates, as well as titanite.

This sample is relatively poor in baddeleyite and zircon, with only 6 baddeleyite grains and 12 zircon grains found. A close examination revealed that the large grain sizes identified by the SEM (Table 3.1) were zircon rims around small baddeleyite cores. Four baddeleyite cores were identified, that were large enough to fit the 20 μm beam of the LA-ICP-MS (Fig. 3.12, B1-B4). Grain B3 exhibited two baddeleyite cores that were analyzed separately.

Figure 3.12. Photomicrographs in (a) plane polarized light and (b) cross polarized light of sample BL-73-M326, as well as back-scattered electron (BSE) images of the analyzed baddeleyite grains (B1-B4) of sample BL-73-M326. Grain numbers correspond to Table 3.5 and Figure 3.14. Note in the BSE images the brighter colored baddeleyite interior and the darker colored zircon rims.

3.4.1.2 SAB-87-D273A

Sample SAB-87-D273A is a fine-grained gabbro sill, intruding into Dumas Formation semi-pelite at 61.401368 N 74.396422 W. The sample shows remnant ophitic texture where amphibole has completely replaced pyroxene. The ophitic amphiboles enclose up to 1 mm plagioclase that have been completely altered to saussurite. Matrix makes up roughly 20 % of the sample and is saussurite rich, and also contains fibrous chlorite and actinolite.

Three separate thin sections for sample SAB-87-273 were analyzed using the SEM. Analyses (Table 3.1) did not identify any baddeleyite, and only a total of 22 zircon grains. Back-scattered electron (BSE) images of these zircon grains showed that the grains were composed of dark disturbed areas that are interpreted to represent metamorphosed zircon as well as undisturbed brighter areas that were interpreted to represent magmatic zircon. Three zircon grains were identified from sample SAB-87-273A that showed areas of undisturbed zircon larger than 20 μm (Fig. 3.13, Z1-Z3) and for this reason were selected for LA-ICP-MS analysis.

Figure 3.13. Photomicrographs in (a) plane polarized light and (b) cross polarized light of sample SAB-87-273A, as well as back-scattered electron (BSE) images of the analyzed zircon grains (Z1-Z3) of sample SAB-87-D273A. Multiple damaged parts can be seen and care was taken to only target the more undisturbed, brighter areas.

3.4.2 Analytical procedure

Thin sections for both samples (BL-73-M326 and SAB-87-D273A) were imaged using the JSM-7100F field emission SEM from the TERRA facility – CREAT (Memorial University, Earth Science Department) to identify Zr-bearing phases present and to image them and their internal morphologies. Grains chosen for U-Pb analysis were then selected for LA-ICP-MS. The U-Pb age measurement with LA-ICP-MS was done on a GeoLas ArF 193 nm excimer laser ablation system (Coherent, Göttingen, Germany) coupled to an element XR (Thermo Fisher Scientific, Bremen, Germany). An in-house custom teardrop-shaped ablation chamber was used. The chamber fits a regular sized thin section and a mount containing standards. Zircon 91500 (Wiedenbeck et al., 1995) was used as primary standard and zircon grains OG 1 (3465.4 ± 0.6 Ma, Stern et al., 2009), Plešovice (337.13 ± 0.37 Ma, Sláma et al., 2008) and 02123 (295 ± 1 Ma, Ketchum et al., 2001) were used as secondary standards.

The ICP-MS was tuned to high sensitivity and a low oxide ratio ($\text{ThO}^+/\text{Th}^+ < 0.3\%$). Data evaluation was done with Lolite applying an exponential-linear downhole U/Pb fractionation correction model. Although ^{235}U was measured it was recalculated from the ^{238}U signal during data evaluation using a $^{238}\text{U}/^{235}\text{U}$ ratio of 137.818. The laser ablation settings were: crater size of 20 μm , repetition rate of 5 Hz, fluence of 5 J/cm^2 , 300 pulses and 0.95 L/min He for the carrier gas flow.

The ICP-MS settings were: 0.92-0.94 L/min for the sample gas flow, 1.00 L/min for the auxiliary gas flow, 16 L/min for the plasma gas flow, the plasma

power was 1200 W. The measured isotopes and dwell times were ^{202}Hg (10 ms), $^{204,206,207,208}\text{Pb}$ (10, 50, 50, 10 ms, respectively), ^{232}Th (10 ms) and $^{235,238}\text{U}$ (50, 10 ms, respectively).

The determined weighted average $^{206}\text{Pb}/^{238}\text{U}$ ages for the secondary standards were 3467 ± 7.1 Ma (MSWD=1.3, n=09), 343 ± 14 Ma (MSWD=1.7, n=15) and 298 ± 28 Ma (MSWD=1.3, n=09) for OG1, Plešovice and 02123, respectively. Because only zircon standards have been used, additional samples from the 2053 ± 12 Ma Phalaborwa baddeleyite standard (see Sylvester et al., 2007; 2009) were analyzed to confirm the accuracy of the baddeleyite U-Pb results. These measurements on the Phalaborwa standard yielded a weighted average of 2055.3 ± 8.9 Ma (MSWD=1.4, n=09). All U-Pb zircon data are reported with 2σ uncertainties and all calculated ages using ISOPLOT are reported at the 95% confidence interval.

3.4.3 Results

3.4.3.1 BL-73-M326

All identified baddeleyite samples have significant zircon rims. Four grains with large baddeleyite cores were identified for analysis (Fig. 3.13). A total of five points has been analysed from these 4 baddeleyite grains (Fig. 3.14, B1 to B4, red filled ellipses). The calculations represent the average of the whole peak, measured during analyses. The signal for sample B2 changed abruptly during the measuring period and we suspect the laser drilled through the grain when the signal changed and therefore only use the first half of the signal as our whole peak analysis. In an additional effort to provide a more detailed and precise age, the smooth parts at the beginning of the individual signals (usually the first third) have been calculated separately as well (red open ellipses). Details can be found in Table 3.5. A regression age was calculated, omitting ellipse B3.1 and yielded an age of 2093 ± 86 Ma, with lower intercept at 214 ± 72 Ma and an MSWD of 15. If the lower intercept is anchored to 0 Ma the age changes to 2062 ± 72 Ma (MSWD = 14). The weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age for all analyses is 2021 ± 100 Ma (MSWD = 62) (Fig. 3.14) and this age (with uncertainties) spans the full range of previously obtained ages for the Povungnituk Group (see Section 2.1.2 or Fig. 3.1). The whole peak analysis of sample B3.1 has a significantly older age than the rest of the sample set. If this individual analysis is omitted from the calculation, the age calculated using a weighted average changes to 2011 ± 83 Ma (MSWD = 38). Using the whole signal analyses, the weighted average is 2060 ± 270 Ma (MSWD = 129). The weighed age using only the first third of each

signal averages 1972 ± 88 Ma (MSWD = 12). The last two ages with their error not only cover the whole range of Povungnituk Group magmatism but also include the overlying Chukotat Group (see Section 2.1.2 or Fig. 3.1).

Figure 3.14. (top) Concordia plot of in-situ LA-ICP-MS zircon analyses from BL-73-M326. (bottom) Weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date from in-situ LA-ICP-MS zircon analyses from BL-73-M326. Analysis B3.1 was considered an outlier and was rejected from the calculations (brown).

3.4.3.2 SAB-87-D273A

Three zircon grains for sample SAB-87-D273A (Z1 to Z3, blue filled ellipses) have been analysed (Table 3.5). All analyses are shown on a concordia diagram in Figure 3.15 and age calculations include the average of the whole peak, measured during analyses. In an additional effort to provide a more detailed and precise age, the smooth parts at the beginning or the end of the individual signals have been calculated separately (blue open ellipses). A regression line through the data points results in an uncalculatable lower intercept (far below 0 Ma); therefore, the weighted average has been used to calculate the age for this sample. All data points shown form a weighed $^{207}\text{Pb}/^{206}\text{Pb}$ mean age of 2079 ± 62 Ma (MSWD = 5.9) (Fig. 3.15). If only the whole peak analyses are used for calculation, the resulting age is 2096 ± 110 Ma (MSWD = 3.1). The complete error range of this analysis does not completely overlap with the previously obtained ages for the Povungnituk Group. Only the reported age for the Korak Sill (Machado et al., 1993) lies within this margin of error.

Figure 3.15. (top) Concordia plot of in-situ LA-ICP-MS zircon analyses from SAB-87-D273A. (bottom) Weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date from in-situ LA-ICP-MS zircon analyses from SAB-87-D273A.

Table 3.5. U-Pb LA-ICP-MS results for magmatic baddeleyite and zircon

No.	U (ppm)	Th (ppm)	Th/U	Ratio						Age						% Disc.
				$^{207}\text{Pb}/^{235}\text{U}$	2 σ	$^{206}\text{Pb}/^{238}\text{U}$	2 σ	$^{207}\text{Pb}/^{206}\text{Pb}$	2 σ	$^{207}\text{Pb}/^{235}\text{U}$ Age (Ma)	2 σ	$^{206}\text{Pb}/^{238}\text{U}$ Age (Ma)	2 σ	$^{207}\text{Pb}/^{206}\text{Pb}$ Age (Ma)	2 σ	
BL-73-M326 - 61.408167 N 74.624238 W																
B1 1/3	887	960	0.77	4.970	0.150	0.3023	0.0110	0.1164	0.0022	1816	26	1699	53	1902	34	11
B1 1/3	546	617	0.80	4.611	0.130	0.2819	0.0092	0.1173	0.0016	1748	24	1601	46	1915	24	16
B2, 1/2	934	1246	0.95	4.680	0.170	0.2480	0.0130	0.1356	0.0041	1758	31	1423	69	2172	53	31
B2, 1/3	1712	1464	0.61	5.670	0.190	0.3374	0.0130	0.1187	0.0026	1927	29	1871	62	1937	39	3
B3.1 1/3	253	11	0.03	8.950	0.360	0.4980	0.0220	0.1295	0.0043	2327	37	2615	91	2091	58	-26
B3.1 1/3	113	35	0.22	10.670	0.690	0.3900	0.0160	0.2150	0.0150	2401	51	2121	71	2944	113	10
B3.2	412	1120	1.93	7.220	0.250	0.3566	0.0120	0.1470	0.0041	2125	29	1965	57	2311	48	12
B3.2, 1/3	289	630	1.56	7.360	0.350	0.4140	0.0200	0.1291	0.0062	2157	44	2229	92	2086	84	-9
B3.2, 1/3	510	1600	2.23	6.040	0.180	0.3475	0.0110	0.1232	0.0021	1976	27	1919	55	2003	30	4
B3.3	354	1040	2.09	5.890	0.190	0.3227	0.0110	0.1292	0.0026	1955	27	1799	53	2087	35	12
B4	136	12	0.06	8.280	0.310	0.4690	0.0180	0.1261	0.0025	2254	34	2467	79	2044	35	-29
SAB-87-273A - 61.401368 N 74.396422 W																
Z1	1197	2690	1.59	6.390	0.120	0.3612	0.0110	0.1267	0.0037	1990	51	2022	17	2053	52	1
Z1, 1st 1/2	1963	5660	2.03	6.010	0.110	0.3525	0.0110	0.1240	0.0036	1952	55	1975	16	2015	51	3
Z2	833	4500	3.80	6.416	0.100	0.3423	0.0100	0.1333	0.0038	1902	50	2032	14	2142	50	10
Z2, 2nd 1/2	138	677	3.45	6.790	0.190	0.3350	0.0140	0.1409	0.0061	1865	66	2082	24	2238	75	17
Z2, 1st 3/4	1105	5790	3.68	6.506	0.095	0.3720	0.0110	0.1258	0.0036	2039	53	2042	13	2040	51	-2
Z2, 1st 1/3	1933	10780	3.93	6.400	0.150	0.3734	0.0120	0.1262	0.0045	2043	58	2033	21	2046	63	-2
Z3	950	4920	3.64	6.517	0.098	0.3642	0.0110	0.1294	0.0039	2001	53	2049	13	2090	53	2

3.5 U-Pb LA-ICP-MS on detrital zircon

3.5.1 Samples

3.5.1.1 BNB13-066

Sample BNB13-066 was collected from a drill core at the Raglan mine site and includes distal turbidite greywacke and mudstone of the Nuvilik Formation, below the Kikialik deposit (Fig. 3.16). The location of the drill core is 61.641949 N 73.909456 W.

Figure 3.16. (top) Representative images of the core sample BNB-13-066. Graded bedding in distal turbiditic greywackes and mudstones of the Nuvilik Formation, below the Kikialik deposit. Photograph taken by W. Bleeker and shown in an unpublished field report for the GSC. (bottom) Images of separated zircon grains in front of two different backgrounds.

3.5.1.2 NK-13-2316

Sample NK-13-2316 is from a quartzite located within the Roberts Lake Syncline. The quartzite layer is located (60.172868 N 70.298081 W), at the northern end of the Qarqasiaq sill and underlies the dunite base of the sill. The sample contains more than 95 % quartz with minor amounts of biotite (Fig. 3.17). Individual quartz grains are less than 0.4 mm in size. No rounded grains can be identified and the sample appears to have been recrystallized based on the texture of interlocking grain boundaries.

Figure 3.17. Representative thin section photomicrograph for sample NK-13-2316 in plane polarized light (top-left) and cross polarized light (top-right). (bottom) Images of separated zircon grains in front of two different backgrounds.

3.5.1.3 NK-13-2321

The sample NK-13-2321 is from a quartzite located within the Roberts Lake Syncline and located at 60.035794 N 70.069172 W. The quartzite layer is overlying a gabbroic layer from the southern end of the Qarqasiaq sill that was previously dated by Wodicka (2002) and has an age of 1882 ± 4 . The quartzite sample contains more than 90 % quartz with minor feldspar and biotite (Fig. 3.18). Individual quartz grains are up to 0.8 mm in size. No rounded grains can be identified and the sample has been recrystallized.

Figure 3.18. Representative thin section photomicrograph for sample NK-13-2321 in plane polarized light (top-left) and cross polarized light (top-right). (bottom) Images of separated zircon in front of two different backgrounds.

3.5.2 Analytical procedure

The quartzite samples were processed at the Jack Satterly Geochronology Laboratory at the University of Toronto. Following crushing and pulverization, initial separation of heavy minerals was carried out on a Wilfley table. This was followed by paramagnetic separations with the Frantz isodynamic separator and density separation using bromoform and methylene iodide. Final sample selection for geochronology was done by hand picking of zircon grains under a binocular microscope. To increase the chances of identifying the youngest grain population, the zircon grains that showed the least amount of rounding were selected (Figs. 3.16-18) and mounted in epoxy.

Grains were partially ablated with a 193 nm New Wave excimer laser and an Agilent 7900 ICP-MS. The laser was operated at 5 Hz and about 5 J/cm² fluence with typical beam diameter of 25 microns, depending on the sample. Data were collected on ⁸⁸Sr (10 ms), ²⁰⁶Pb (30 ms), ²⁰⁷Pb (70 ms), ²³²Th (10 ms) and ²³⁸U (20 ms). Prior to analyses, spots were pre-ablated with a larger beam diameter for 1 sec (5 pulses) to clean the surface. Following a 10 sec period of baseline accumulation, the laser sampling beam was engaged and data were collected for 25 seconds followed by a washout period. About 150 measurement cycles per sample were collected and ablation pits are about 15 microns deep.

Data were edited and reduced using custom VBA software (UtilLAZ program) written by Davis (1982). ²⁰⁶Pb/²³⁸U ratios show increasing fractionation throughout zircon runs caused by loss of refractory U with increasing penetration depth while the ²⁰⁷Pb/²⁰⁶Pb profile is usually flat. No corrections were made for

common Pb, since the ^{204}Pb peak is too small to be measured precisely and common Pb is usually insignificant for unaltered Precambrian zircon. If present, common Pb would have the effect of pushing data to the right, away from the concordia curve, along a shallow mixing line with slope determined by the isotopic composition of the common Pb contaminant. ^{88}Sr was monitored in order to detect intersection of the beam with zones of alteration or inclusions. Data showing high Sr or irregular time resolved profiles were either averaged over restricted Sr-free time windows or rejected (marked with * out in Table 3.6). The Th/U ratio of zircon can be a useful petrogenetic indicator and was also measured, although it is only a rough estimate because the ratio is not constant in the standard. Low Th/U (< 0.1) is characteristic of metamorphic and hydrothermal zircon, whereas most zircon crystallized from felsic melts has Th/U in the range 0.1-1.0. The zircon standards used were DD85-17, a quartz diorite from the Marmion batholith in northwest Ontario previously dated at 3002 ± 2 Ma by ID-TIMS (Tomlinson et al., 2003) and DD91-1, a monzodiorite from the Pontiac province of Quebec dated at 2682 ± 1 Ma (Davis, 2002). Standards were analyzed after every 4 measurements.

3.5.3 Results

Results of U-Pb isotopic analyses by LA-ICP-MS are given in Table 3.6. Data are plotted on Figures 3.19-3.20 below. Average age errors in the text and error ellipses on figures are given at 2 sigma (twice the sigma errors in Table 3.6). These were calculated using the Isoplot program of Ludwig (2003). MSWD values would be expected to be around 1 or slightly higher with correctly chosen analytical errors for $^{207}\text{Pb}/^{206}\text{Pb}$ ages if the age population is unimodal. Since Pb/U errors do not include possible biases from compositional differences between samples and standard, scatter above and below concordia may be more pronounced.

U decay constants are taken from Jaffey et al. (1971). For Paleoproterozoic and Archean samples, $^{207}\text{Pb}/^{206}\text{Pb}$ ages are more precise than $^{206}\text{Pb}/^{238}\text{U}$ ages and are also much less susceptible to fractionation biases between samples and standards. Therefore, ages are based on $^{207}\text{Pb}/^{206}\text{Pb}$ ratios.

3.5.3.1 BNB-13-066

Sample BNB-13-066 contains abundant zircon as colourless prisms with varying degrees of rounding. Fourteen zircon grains that showed the least amount of rounding have been measured. All measurements have individual ages above 2700 Ma (Table 3.6). The majority (11 of 14) of the samples fall on a regression line that intersects the concordia at 2705 ± 14 Ma with a lower intercept of 598 ± 460 Ma (Fig. 3.19).

Figure 3.19. Concordia diagram of in-situ LA-ICP-MS zircon analyses from sample BNB13-066. Linear regression omits three grains (empty ellipses) and has an upper intercept of 2705 ± 14 .

Table 3.6. U-Pb LA-ICP-MS results for detrital zircons of sample BNB-13-066

No.	U (ppm)	Th/U	Ratio				Age								% Disc.
			$^{207}\text{Pb}/^{235}\text{U}$	1 σ	$^{206}\text{Pb}/^{238}\text{U}$	1 σ	$^{207}\text{Pb}/^{206}\text{Pb}$	1 σ	$^{207}\text{Pb}/^{235}\text{U}$ Age (Ma)	1 σ	$^{206}\text{Pb}/^{238}\text{U}$ Age (Ma)	1 σ	$^{207}\text{Pb}/^{206}\text{Pb}$ Age (Ma)	1 σ	
BNB-13-066 - 61.641949 N 73.909456 W															
BNB066-2*	59	0.31	12.506	0.134	0.4703	0.0039	0.1954	0.0013	2643	10	2485	17	2767	11	12
BNB066-5	42	1.13	13.352	0.151	0.5132	0.0049	0.1914	0.0011	2705	11	2670	21	2731	10	3
BNB066-6	45	0.76	13.212	0.133	0.5140	0.0043	0.1893	0.0010	2695	9	2674	18	2711	9	2
BNB066-7*	57	0.81	15.677	0.137	0.5896	0.0043	0.1960	0.0008	2857	8	2988	18	2767	8	-10
BNB066-9	46	0.59	14.188	0.136	0.5520	0.0045	0.1893	0.0008	2762	9	2834	19	2711	8	-6
BNB066-10	73	0.79	14.132	0.161	0.5539	0.0057	0.1876	0.0008	2759	11	2841	24	2699	8	-7
BNB066-11	102	0.45	12.201	0.098	0.4851	0.0032	0.1846	0.0007	2620	8	2550	14	2675	8	6
BNB066-12	24	0.98	14.881	0.148	0.5705	0.0048	0.1910	0.0009	2808	9	2910	20	2735	9	-8
BNB066-13	19	0.83	14.003	0.207	0.5434	0.0068	0	0.0014	2750	14	2797	28	2715	13	-4
BNB066-14*	152	0.13	17.200	0.160	0.6087	0.0051	0	0.0007	2946	9	3065	21	2866	6	-9
BNB066-15	26	0.84	14.381	0.165	0.5540	0.0054	0.1900	0.0011	2775	11	2842	22	2727	10	-5
BNB066-16*	116	0.41	14.627	0.151	0.5487	0.0048	0.1952	0.0009	2791	10	2820	20	2771	9	-2
BNB066-18	56	0.24	12.930	0.123	0.5061	0.0040	0.1871	0.0009	2675	9	2640	17	2701	9	3
BNB066-19*	86	1.42	9.753	0.079	0.4094	0.0028	0.1745	0.0006	2412	7	2212	13	2585	7	17

* = omitted from age calculation

3.5.3.2 *NK-13-2316*

Sample NK-13-2316 contains abundant zircon as colourless prisms with varying degrees of rounding. Thirty two zircon grains that showed the least amount of rounding have been measured (Table 3.7). All measured zircon grains have Archean ages and do not show input from Paleoproterozoic magmatism. Multiple suits of ages can be seen on the diagram and no linear regression has been calculated. On a probability density diagram three peaks can be seen. These peaks correlate to ages between 2700 and 2750 Ma, ~2800 Ma and between 2850 and 2950 (Fig. 3.20).

Figure 3.20. Concordia plot of in-situ LA-ICP-MS zircon analyses from NK-13-2316. Multiple suites of ages can be seen on the diagram and no linear regression has been calculated. On a probability density diagram three peaks can be seen. These peaks correlate to ages between 2700 and 2750 Ma, ~2800 Ma as well as between 2850 and 2950 Ma.

Table 3.7. U-Pb LA-ICP-MS results for detrital zircons of sample NK-13-2316

No.	Ratio						Age						% Disc.
	U (ppm)	Th/U	$^{207}\text{Pb}/^{235}\text{U}$	1 σ	$^{206}\text{Pb}/^{238}\text{U}$	1 σ	$^{207}\text{Pb}/^{206}\text{Pb}$	1 σ	$^{207}\text{Pb}/^{235}\text{U}$	1 σ	$^{206}\text{Pb}/^{238}\text{U}$	1 σ	
									Age (Ma)		Age (Ma)		
NK-15-2316 - 60.172868 N 70.298081 W													
NK15-2316-2	87	0.06	15.974	0.121	0.5529	0.0034	0.2123	0.0007	2875	7	2837	14	3
NK15-2316-3	111	0.65	18.533	0.134	0.6217	0.0037	0.2188	0.0008	3018	7	3117	15	-7
NK15-2316-4	248	0.52	12.972	0.082	0.5017	0.0027	0.1896	0.0005	2678	6	2621	11	4
NK15-2316-5	171	0.73	13.524	0.101	0.5268	0.0035	0.1880	0.0005	2717	7	2728	15	-1
NK15-2316-7	63	0.33	17.018	0.135	0.5907	0.0039	0.2109	0.0007	2936	8	2992	16	-4
NK15-2316-8	79	0.37	15.428	0.116	0.5499	0.0034	0.2055	0.0007	2842	7	2825	14	1
NK15-2316-9	177	0.45	13.457	0.113	0.5186	0.0037	0.1902	0.0006	2712	8	2693	16	1
NK15-2316-10	292	0.84	14.549	0.158	0.5275	0.0050	0.2023	0.0008	2786	10	2731	21	4
NK15-2316-11	282	0.98	16.864	0.161	0.5849	0.0050	0.2115	0.0006	2927	9	2969	20	-3
NK15-2316-13	374	0.50	15.998	0.123	0.5625	0.0038	0.2086	0.0005	2877	7	2877	16	0
NK15-2316-17	247	0.49	13.475	0.116	0.5268	0.0039	0.1876	0.0006	2714	8	2728	17	-1
NK15-2316-18	201	0.83	14.725	0.115	0.5397	0.0037	0.2000	0.0005	2798	7	2782	15	1
NK15-2316-19*	104	0.28	14.915	0.121	0.5322	0.0036	0.2051	0.0007	2810	8	2751	15	4
NK15-2316-21	104	0.25	15.565	0.126	0.5488	0.0038	0.2073	0.0007	2851	8	2820	16	2
NK15-2316-23	65	0.65	14.032	0.115	0.5359	0.0036	0.1911	0.0008	2752	8	2766	15	-1
NK15-2316-24	88	0.18	17.777	0.148	0.5771	0.0041	0.2245	0.0007	2978	8	2937	17	3
NK15-2316-25	179	1.30	15.334	0.117	0.5678	0.0035	0.1965	0.0006	2836	7	2899	15	-5
NK15-2316-26	96	0.71	15.661	0.102	0.6033	0.0031	0.1889	0.0006	2856	6	3043	12	-15
NK15-2316-27	103	0.46	14.106	0.102	0.5448	0.0034	0.1883	0.0005	2757	7	2804	14	-4
NK15-2316-28	108	0.34	13.098	0.121	0.5044	0.0035	0.1889	0.0010	2687	9	2633	15	4
NK15-2316-29	163	0.29	15.943	0.128	0.5610	0.0039	0.2068	0.0006	2873	8	2871	16	0
NK15-2316-30	62	0.76	17.184	0.125	0.5896	0.0036	0.2121	0.0006	2945	7	2988	15	-3
NK15-2316-31	95	1.20	16.063	0.118	0.5767	0.0036	0.2028	0.0006	2881	7	2935	15	-4
NK15-2316-32	138	1.12	14.769	0.118	0.5395	0.0035	0.1994	0.0008	2800	8	2781	15	1
NK15-2316-33	117	0.18	15.424	0.101	0.5681	0.0031	0.1980	0.0005	2842	6	2900	13	-4
NK15-2316-34	65	1.14	13.686	0.117	0.5327	0.0040	0.1874	0.0007	2728	8	2753	17	-2
NK15-2316-35	111	0.22	14.848	0.094	0.5271	0.0028	0.2056	0.0006	2806	6	2729	12	6
NK15-2316-36	166	0.59	16.351	0.174	0.6062	0.0061	0.1970	0.0005	2898	10	3055	24	-12
NK15-2316-37	100	0.52	15.338	0.101	0.5432	0.0031	0.2062	0.0005	2836	6	2797	13	3
NK15-2316-38	66	0.21	15.364	0.113	0.5427	0.0034	0.2066	0.0006	2838	7	2795	14	3
NK15-2316-39	128	0.28	14.842	0.126	0.5439	0.0039	0.1991	0.0008	2805	8	2800	16	0
NK15-2316-40	157	0.48	16.950	0.117	0.5830	0.0034	0.2120	0.0005	2932	7	2961	14	-2

* = omitted from age calculation

3.5.3.3 *NK-13-2321*

Sample NK-13-2321 contains abundant zircon grains as colourless prisms with varying degrees of rounding. Forty zircon grains that showed the least amount of rounding have been measured (Table 3.8).

Two main accumulation of ages can be seen. The majority of the samples show Archean ages that fall between 2600 and 3000 Ma. The younger group falls on a regression line, anchored to 0 Ma that has an upper intercept of 1858 ± 70 Ma (Fig. 3.21). This age corresponds with the Hellencourt Formation of Cycle 2.

Figure 3.21. Concordia plot of in-situ LA-ICP-MS zircon analyses from NK-13-2321. Two main accumulations of ages can be seen. The majority of the samples show Archean ages that fall between 2600 and 3000 Ma. The younger group falls on a regression line that, anchored to 0 Ma, has an upper intercept of 1858 ± 70 Ma.

Table 3.8. U-Pb LA-ICP-MS results for detrital zircons of sample NK-13-2321

No.	U (ppm)	Ratio						Age						% Disc.	
		Th/U	$^{207}\text{Pb}/^{235}\text{U}$	1σ	$^{206}\text{Pb}/^{238}\text{U}$	1σ	$^{207}\text{Pb}/^{206}\text{Pb}$	1σ	$^{207}\text{Pb}/^{235}\text{U}$	1σ	$^{206}\text{Pb}/^{238}\text{U}$	1σ	$^{207}\text{Pb}/^{206}\text{Pb}$		1σ
		Age (Ma)			Age (Ma)			Age (Ma)							
NK-15-2321 - 60.035794 N 70.069172 W															
NK2321-1	255	0.40	5.128	0.090	0.3123	0.0041	0.1189	0.0012	1841	15	1752	20	1943	21	11
NK2321-2	357	0.19	10.519	0.144	0.4418	0.0043	0.1724	0.0013	2482	13	2359	19	2584	16	10
NK2321-3	49	0.32	5.380	0.223	0.3431	0.0105	0.1136	0.0031	1882	35	1901	50	1860	50	-3
NK2321-4	138	0.77	15.931	0.200	0.5674	0.0052	0.2034	0.0010	2873	12	2897	21	2856	14	-2
NK2321-5	78	1.29	14.299	0.207	0.5487	0.0053	0.1888	0.0013	2770	14	2820	22	2734	18	-4
NK2321-6	60	0.37	4.978	0.152	0.3201	0.0044	0.1154	0.0030	1816	26	1790	22	1845	49	3
NK2321-7	60	0.25	14.186	0.169	0.5312	0.0042	0.1974	0.0013	2762	11	2746	18	2774	15	1
NK2321-8	43	0.87	12.987	0.184	0.5093	0.0047	0.1878	0.0017	2679	13	2654	20	2698	18	2
NK2321-9	103	0.83	14.223	0.177	0.5272	0.0047	0.1980	0.0011	2765	12	2730	20	2790	14	3
NK2321-10	256	0.35	14.483	0.188	0.5335	0.0052	0.1993	0.0011	2782	12	2756	22	2800	14	2
NK2321-11	35	0.65	15.607	0.238	0.5532	0.0060	0.2080	0.0018	2853	15	2839	25	2863	18	1
NK2321-12	117	0.73	14.096	0.178	0.5294	0.0051	0.1971	0.0009	2756	12	2739	21	2769	14	1
NK2321-13	135	0.55	14.432	0.225	0.5422	0.0065	0.1978	0.0012	2779	15	2793	27	2768	17	-1
NK2321-14	29	0.32	38.589	0.611	0.8012	0.0095	0.3573	0.0024	3735	16	3793	34	3704	16	-3
NK2321-15	586	0.19	5.613	0.097	0.3546	0.0048	0.1167	0.0010	1918	15	1956	23	1877	20	-5
NK2321-16	61	0.59	17.802	0.256	0.5897	0.0062	0.2212	0.0016	2979	14	2988	25	2973	16	-1
NK2321-17	75	0.75	15.164	0.229	0.5512	0.0057	0.2004	0.0016	2826	14	2830	24	2822	18	0
NK2321-18	43	0.47	18.100	0.310	0.5818	0.0071	0.2256	0.0021	2995	16	2956	29	3021	19	3
NK2321-19	37	0.67	17.455	0.287	0.6595	0.0079	0.1923	0.0017	2960	16	3265	31	2759	18	-23
NK2321-20	289	0.16	7.135	0.100	0.3693	0.0038	0.1406	0.0009	2128	12	2026	18	2229	17	11
NK2321-21	56	0.59	18.983	0.305	0.6038	0.0066	0.2293	0.0019	3041	15	3045	27	3038	19	0
NK2321-22	39	0.51	14.795	0.256	0.5397	0.0062	0.2007	0.0020	2802	16	2782	26	2817	21	1
NK2321-23	62	0.35	5.062	0.185	0.3427	0.0079	0.1084	0.0030	1830	31	1899	38	1752	51	-10
NK2321-24	149	1.01	14.620	0.162	0.5458	0.0042	0.1970	0.0010	2791	11	2808	18	2779	13	-1
NK2321-25	94	1.16	16.335	0.227	0.6065	0.0061	0.1984	0.0014	2897	13	3056	25	2788	16	-12
NK2321-26	36	0.78	13.414	0.190	0.5117	0.0045	0.1934	0.0018	2709	13	2664	19	2743	18	4
NK2321-27	344	0.52	13.108	0.150	0.5065	0.0042	0.1908	0.0010	2687	11	2642	18	2722	13	4
NK2321-28	212	0.91	14.491	0.202	0.5455	0.0056	0.1958	0.0013	2782	13	2806	23	2765	15	-2
NK2321-29	116	0.14	15.114	0.215	0.5301	0.0054	0.2100	0.0012	2822	14	2742	23	2881	16	6
NK2321-30	222	0.88	13.902	0.204	0.5441	0.0057	0.1872	0.0011	2743	14	2800	24	2701	17	-5
NK2321-31	40	0.53	16.117	0.335	0.5698	0.0076	0.2063	0.0029	2884	20	2907	31	2867	26	-2
NK2321-32	121	0.64	13.538	0.190	0.5177	0.0050	0.1898	0.0014	2718	13	2690	21	2739	17	2
NK2321-33	81	0.70	15.872	0.299	0.5649	0.0069	0.2031	0.0024	2869	18	2887	28	2857	23	-1
NK2321-34	87	1.12	13.886	0.227	0.5400	0.0055	0.1852	0.0019	2742	15	2784	23	2711	21	-3
NK2321-35	92	0.62	13.396	0.190	0.5029	0.0052	0.1922	0.0014	2708	13	2626	22	2770	16	6
NK2321-36	87	0.37	13.789	0.214	0.5398	0.0059	0.1845	0.0017	2735	15	2783	25	2701	18	-4
NK2321-37	121	0.72	17.614	0.225	0.6083	0.0056	0.2095	0.0012	2969	12	3063	23	2906	14	-7
NK2321-38	178	0.42	15.910	0.179	0.5916	0.0049	0.1948	0.0009	2871	11	2996	20	2785	13	-9
NK2321-39	83	0.85	15.753	0.189	0.5922	0.0047	0.1927	0.0014	2862	11	2999	19	2767	15	-10
NK2321-40	42	0.53	13.770	0.259	0.5168	0.0067	0.1930	0.0023	2734	18	2685	29	2770	22	4

* = omitted from age calculation

3.6 Discussion

In the scope of this thesis, 7 mafic igneous samples have been analysed using different U-Pb geochronological techniques and 3 sedimentary samples have been analysed for detrital zircon grains.

3.6.1 Age of the Povungnituk Group

3.6.1.1 The role of the Korak sill

An age of 2038 ± 3 Ma was previously obtained for the gabbroic Korak sill (Machado et al, 1993) that intrudes siltstone in the upper Dumas Formation, located at the base of the Povungnituk volcanic sequence in the west of the belt. Picard et al. (1989a, 1989b) correlates this and other diabase sills in the lower Povungnituk sediments to the main flood basalt package of the Beuparlant Formation, implying that the Povungnituk volcanics could have started at 2038 Ma. In an effort to test this claim, we selected two samples from gabbroic sills intruding the Dumas Formation in the east of the belt. Both samples were not successful in yielding baddeleyite or zircon grains for U-Pb TIMS analyses and instead ages were obtained using LA-ICP-MS. The reported ages are 2021 ± 100 Ma and 2079 ± 62 Ma. The large margin of error on both analyses prohibit correlation to the previously obtained ages of the Korak sill. The second age (sample SAB-87-D273A) has an age of 2079 ± 62 Ma, where error margins are too large to effectively correlate the age of the sample to previously obtained dates. The error of the analyses only overlaps only with the age of the Korak sill

and therefore might be an indicator that a ca. 2040 Ma magmatic event is widespread over the whole Cape Smith belt.

Although included in the mapping of Picard et al. (1989a/b), the Korak sill has not been analysed for geochemistry and can therefore not be magmatically linked with certainty to any event.

Nilsson et al. (2010) noted that the 2038 ± 3 Ma Korak sills could be instead linked to the Kangamiut-MD3 LIP with dolerite ages of 2050 ± 2 Ma, 2041 ± 3 Ma and 2029 ± 3 Ma in southern West Greenland and 2045 ± 2 Ma, 2051 ± 6 Ma, $2051, \pm 1$ Ma, and 2050 ± 2 Ma for dolerite dykes in the Nain Province (Nilsson et al., 2010; Sahin and Hamilton, 2019). These have been interpreted to represent components of a single, areally extensive, radiating dyke swarm that signaled the arrival of a mantle plume centred on what is presently the western margin of the North Atlantic craton (Nilsson et al., 2010; Sahin and Hamilton, 2019). At 2040 Ma the North Atlantic craton is interpreted to have been located further north adjacent to the NE margin of Superior craton, at a minimum distance of approximately 250 km from the Korak sill (see Fig. 5 in Nilsson et al. (2010)). More detailed work to constrain of the location of the North Atlantic craton is required to access the distance between plume origin and possible dyke emplacement and link with the Korak sill.

3.6.1.2 Age of the Povungnituk Group flood basalts

Two U-Pb TIMS baddeleyite ages have been obtained from the middle and the top of the Beuparlant Formation. These ages are the first ages on mafic rocks of the stratigraphy of the Beuparlant Formation. The best estimate for the timing of the main flood basalt package is the ca. 1998 ± 6 Ma maximum age that we have obtained for a sill intruding the upper Povungnituk Group. This sample matches the geochemistry of the associated volcanic rocks (see Section 4.4.2). The age is obtained from a dolerite sill at the top of the exposed Povungnituk Group succession and therefore provides a new minimum age for the Povungnituk magmatism.

The new age of 1998 ± 6 Ma reported herein overlaps with the 1991 ± 2 Ma age reported by Machado et al. (1993) from a small granodiorite intrusion which cross-cuts Povungnituk Group pillow basalts and represented the minimum age for the formation of the Povungnituk Group until now.

Only rare dolomite and semi-pelite occurrences, described as "volumetrically subsidiary intercalated" (Mungall, 2007) or "very minor quantities of interflow sediment" (St-Onge et al., 1992), are present within the Povungnituk flood basalts and led to the interpretation by St-Onge et al. (1992) that the main part of the Beuparlant Formation was emplaced in a relatively short time. This interpretation is also consistent with the Korak sills belonging to a different event (as discussed above). The short duration of emplacement of the Beuparlant Formation is also supported by Mungall (2007), who interprets the Beuparlant

Formation as a single sequence, ~3 km in original thickness. This is comparable to other LIP events, like the Siberian Trap (Burgess and Bowring, 2015) and Deccan Trap (Schoene et al., 2015) which also lack significant interflow sediments and for which precise U-Pb dating indicates a duration of less than a few million years.

Given the regional chemical homogeneity of the Beauparlant Formation (Modeland et al., 2003; this thesis), and the lack of significant sedimentation or alteration horizons (St-Onge et al., 1992; Mungall, 2007), it is possible that large components of the Beauparlant Formation may have formed over a short duration. Therefore, we provisionally interpret our age of 1998 ± 6 Ma as representative of the main phase of Beauparlant Formation volcanism.

The second dated sample (BLS-73-197) was obtained from a sill intruding the middle of the Beauparlant Formation and shows a younger age of 1967 ± 7 Ma. The older age (1998 ± 6 Ma), which is located at the top of the stratigraphic column, means that it cannot be representative for the majority of the Beauparlant Formation. A more reasonable interpretation for the 1967 ± 7 Ma age is that it is a later stage pulse that intruded into the previously formed flood basalt package. The sample shows geochemical similarities to the enriched mid-ocean ridge basalt (E-MORB) endmember of the Beauparlant Formation. This endmember is interpreted to represent melting of the ambient mantle (see Section 4.5.3.2 of thesis). In the ca. 30 myr between the emplacement of the Beauparlant Formation and the emplacement of sample BLS-73-197, it is likely

that another melting event tapped into the same ambient mantle source as before, although the cause of the secondary magmatic pulse might differ.

In addition, an age of ca. 820 Ma has been obtained from a gabbro at the base of the Beuparlant Formation which probably represents a Neoproterozoic sill.

3.6.1.3 Correlation with other magmatic units

The observed age of 1998 ± 6 Ma for sample BLS-73-31 links with a number of mafic magmatic units in the Quebec promontory of the Superior craton. An identical age was obtained from the Watts Group (Purtiniq ophiolite) in the northern Cape Smith Belt (Parrish, 1989). The Watts Group comprises layered ultramafic and mafic rocks, clinopyroxenite intrusions, sheeted mafic dykes and gabbros, and pillowed and massive mafic volcanic rocks intruded by rare felsic sills and dykes (St-Onge et al., 1992; Scott et al., 1991; 1999). The rocks that comprise the ophiolite are separated from the underlying sedimentary rocks by a south-verging thrust fault. Zircon grains obtained from a metagabbro yield a U-Pb age of 1998 ± 2 Ma (Parrish, 1989). The Minto dykes in the interior of the craton have a U-Pb baddeleyite-zircon age of 1998 ± 2 Ma (Buchan et al., 1998). The Lac Shpogan dykes of the James Bay area of the eastern Superior craton have been recently correlated with the Minto dykes based on a U-Pb baddeleyite age of 1999 ± 2 Ma (Hamilton et al., 2016).

Multiple ages obtained within the Cape Smith belt overlap within the uncertainty of the 1967 ± 7 Ma age obtained from sample BLS-73-197. Within the Povungnituk Group, a rhyolite sample obtained from the alkaline rocks at the Cecilia and Kenty lakes localities has been dated at 1958 ± 3 Ma (Parrish, 1989). The localities are interpreted to represent an association of an alkaline volcanic series with the continental tholeiites and sediments of the Povungnituk Group via a volcanic island on a rifted continental margin (Gaonac'h et al., 1992) and are not directly linked to the main flood basalt pulse. An initial $\epsilon_{\text{Nd}}^{2.0\text{Ga}}$ value of + 0.4 on the dated rhyolite sample further indicates minor contamination with crust, which probably occurred during ponding of mafic magma at the crust-mantle interface during rifting (Francis et al., 1983; Hegner and Bevier, 1991). A rhyodacite from the Parent Group, which separates the Watts Group and the Chukotat Group, ranges in age between 1917 to 2423 Ma. The Parent Group consists of mafic to felsic lavas and pyroclastites that are intercalated with semipelite, quartzite, and greywacke beds (Picard et al., 1990). The volcanic rocks comprise massive and pillowed basalts, andesitic pyroclastites, and rhyolitic crystal tuffs, forming a calc-alkaline suite (Picard et al., 1990). The dated rhyodacite contains a wide variety of zircon morphologies suggesting the presence of a large quantity of inherited zircon (Machado et al., 1993). Together with petrographic indicators, Machado et al. (1993) suggest that the rock may either be a terrigenous metasediment or that it contains an inherited component. One of the U-Pb zircon analyses yielded a minimum age of 1968 Ma, correlating with the magmatic age obtained on sample BLS-73-197. The wide range of dates

(between ca. 1917 and 2423 Ma) within the rhyodacite led Machado et al. (1993) to the interpretation that the rhyodacite generally is younger than 1917 Ma and contains inherited zircon grains older than 2.4 Ga. It may be possible that the reported age of 1968 Ma represents inherited zircon grains from a magmatic event that correlates to the newly dated sill (BLS-73-197) within the Beuparlant Formation, but large uncertainties make a correlation and comparison to our dated sample unfeasible. A metamorphosed anorthositic layered gabbro within the Watts Group in the northern part of the Cape Smith belt includes zircon grains with cloudy cores that are 1995-2000 Ma in age, as well as clear overgrowths that are dated at 1977 ± 3 (Parrish, 1989), which overlaps with the observed age of sample BLS-73-197 of 1967 ± 7 Ma. This age has been interpreted as being related to seafloor metasomatism.

No clear correlation can be established between the observed age of 1967 ± 7 Ma and other units of similar age within the Cape Smith belt. The geochemical similarity of the sample to E-MORB, with no indication of the plume component that is observed in the older sample BLS-73-31 (see Sections 4.4.2, 4.4.3), might suggest that this sample just represents a smaller pulse, maybe related to the ongoing rifting along the Superior craton margin.

3.6.2 Age of the Chukotat Group

3.6.2.1 Age of the Chukotat Group

Chukotat volcanism was previously thought to have occurred between 1887 and 1870 Ma and has recently been narrowed to between 1883 and 1870 Ma (Bleeker, 2014; Bleeker and Kamo, 2018). Our new baddeleyite ages of 1874 ± 3 and 1861 ± 28 from intrusions give new insight and provide an additional minimum age for the formation of the Chukotat group. While a detailed location or precise stratigraphic position is not known for the previously reported age of 1870 ± 4 (Parrish, unpublished data; mentioned in St-Onge et al., 1992), our new age at the very top of the Chukotat succession supports the Chukotat Group as a single magmatic event that has formed over 9 to 20 myr.

3.6.2.2 *Correlation with other regional magmatic units*

Multiple magmatic events with similar ages are recognized around the Superior craton margin. These events include the well-studied Circum-Superior LIP (Baragar and Scoates, 1987; Ernst and Buchan, 2001; Heaman et al., 2009; Ernst and Bell, 2010; Minifie et al., 2013; Ciborowski et al., 2016) of which the Chukotat Group forms one part. The different ages of the Circum-Superior LIP are summarized here.

In Cycle 2 of the Labrador Trough, the Willbob Formation was dated at 1885 ± 67 Ma (Rohon et al., 1993). Two glomeroporphyritic Montagnais gabbro sills within the Labrador Trough have ages of 1874 ± 3 (Machado et al., 1997) and 1878.5 ± 0.8 (Bleeker and Kamo, 2018; re-dated from Findlay et al., 1995). Two more analyses obtained on a rhyodacite and syenite within the Labrador Trough yielded ages of 1870 ± 4 (Machado et al., 1997) and 1877.8 ± 1.3 (Findlay et al., 1995), respectively. The sulphide-bearing Qarqasiaq sill in the Roberts Lake Lake Syncline was dated at 1874 ± 4 (Wodicka et al., 2002). In the Lake Superior region the Hemlock formation has been dated at 1874 ± 9 Ma (Schneider et al., 2002) and the Gunflint formation has been dated at 1878 ± 1.3 Ma (Fralick et al., 2002). A lamprophyric dyke on Little Presque Isle has been dated at 1877 ± 5 Ma (Craddock et al., 2007). In the Thompson Nickel belt, two pyroxenites with the ages of 1880 ± 5 (Hulbert et al., 2005) and 1876.7 ± 5.1 (Heaman et al., 2009) overlap with the age of 1885 ± 49 (Hulbert et al., 2005) obtained from the sulphide ores. In the Winnipegosis Belt a basalt with associated komatiites yielded an age of 1870 ± 7.1 (Waterton et al., 2017). Two sills have been dated

in the Fox River belt, showing ages of $1882.9 \pm 1.5/-1.4$ Ma (Heaman et al., 1986) and 1900 ± 14 (Heaman et al., 2009). In the Hudson Bay region, a Haig sill is dated at 1870 ± 4 Ma (Hamilton et al., 2009) and a Sutton Inlier sill is dated at 1870 ± 2 Ma (Hamilton et al., 2009).

In addition to the units at the margins of the Superior craton, a radiating dyke swarm has been identified that comprises the 1876 ± 8 Ma Pickle Crow dykes (Buchan et al., 2003), the 1870.7 ± 1.1 Ma Fort Albany dykes (Hamilton and Stott, 2008) and the Molson dykes, that have four reported ages between 1882 and 1885 Ma (Heaman et al., 1986; 2009) and one younger dyke dated at $1877 \pm 7/-4$ Ma (Halls and Heaman, 2000)

The age range observed within the Chukotat Group is also observed within several carbonatite complexes. Within the Kapuskasing Structural Zone several ages between 1882 and 1886 Ma have been observed (Ruhklov and Bell, 2010; Sage 1988a,b), as well as 1880 ± 2 for a carbonatite dyke within the Labrador Trough (Chevé and Machado, 1988) and a 1865 ± 22 carbonatite complex in the NW Superior craton (Ruhklov and Bell, 2010).

3.6.3 Age of detrital sedimentary horizons

Sample BNB-13-066 from the Nuvilik Formation of the Cape Smith belt was selected for detrital zircon work to assess the nature of the mapped fault belt. The presumably youngest fraction of zircon grains was selected (based on having the minimum degree of mechanical rounding) and all the analysed grains fall on a regression line that intersects the concordia curve at 2705 ± 14 Ma. If this age represents the youngest age of source rocks, no input from any (1998 – 1874 Ma) Circum-Ungava magmatism can be found. Therefore, it is not possible to temporally separate the sedimentary horizon between Povungnituk and the Chukotat Group from the sediments of the Dumas Formation, and the presence of a major thrust fault at the base of the Nuvilik Formation remains a matter of debate.

Within the Roberts Lake Syncline, two quartzite units have been sampled. They are located on the top and the bottom of the Qarqasiaq sill, which has been dated by Wodicka et al. (2002) at 1874 ± 4 Ma. The quartzite horizon that lies stratigraphically beneath the Qarqasiaq sill (NK-13-2316) shows multiple generations of detrital zircon grains that are all older than 2700 Ma. These observed zircon ages correlate with the results obtained from the Cape Smith belt (2705 ± 14 Ma). The second sample (NK-13-2321) lies stratigraphically above the Qarqasiaq sill. Here, a small population of grains have an age of 1858 ± 70 Ma and overlap with the age for the 1882 ± 4 Ma Qarqasiaq sill and the inferred age magmatism in the Roberts Lake Syncline (Wodicka et al., 2002). More detailed work is necessary to conclude the precise depositional age of the

quartzite horizon, as secondary rims could have formed due to the intrusion of the Qarqasiaq sill.

3.7 Conclusions

Seven ages on magmatic baddeleyite and zircon have been determined throughout the Cape Smith belt. The ages were obtained using multiple U-Pb analytical techniques. The most precise ages were obtained via ID-TIMS and yielded two new ages of 1998 ± 6 Ma and 1967 ± 7 Ma for the Povungnituk Group magmatism and one new age of 1874 ± 3 Ma for an interpreted feeder to the Chukotat Group. The older age of the Povungnituk Group is interpreted to represent the main pulse for the Beuparlant Formation volcanism and correlates with ages obtained from the Minto and Lac Shpogan dykes in the interior of the Superior craton and the Watts Group of the Cape Smith belt. The younger magmatic age correlates partly with zircon grains found within the upper Beuparlant Formation that show detrital origin as well as metamorphic zircon rims measured within the Watts Group. The TIMS age for a gabbroic dyke within the Nuvilik Formation yielded an age of 1874 ± 3 Ma. Together with previously published ages that correlate with Ni-Cu-(PGE) deposit forming processes related to Chukotat magmatism, the age suggests that Chukotat magmatism was active for at least 5 myr and might have lasted as long as 12 myr. Magmatic ages obtained via SIMS and LA-ICP-MS show larger errors. Two ages obtained for sills intruding into the Dumas Formation show ages that, depending on the calculation method, either correlate to the 2038 Ma Korak sill, or show ages that overlap with the originally interpreted age range (2040 – 1958 Ma) for the Povungnituk Group. A SIMS age obtained from the top of the Chukotat Formation yielded an age of 1861 ± 28 Ma. It correlates with the ages obtained

on feeder dykes at the bottom of the succession (1883 – 1874 Ma) and shows that the entire package of the Chukotat Group belongs to the same temporally constrained large magmatic event.

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**4 GEOCHEMISTRY OF THE POVUNGNITUK GROUP OF
THE CAPE SMITH BELT: PART OF A CRATON-SCALE
CIRCA 2.0 GA MINTO-POVUNGNITUK LARGE
IGNEOUS PROVINCE, NORTHERN SUPERIOR
CRATON**

4.1 Introduction

Large igneous provinces (LIPs) have become an important focus for research in recent years due to their use in paleocontinental reconstructions (e.g., Bleeker and Ernst, 2006; Ernst et al., 2013), in exploration targeting (e.g., Ernst and Jowitt, 2013, 2017), and as a result of their links to dramatic climate change (e.g., Wignall, 2001, 2005; Ernst and Youbi, 2017). LIPs represent large volume ($>0.1 \text{ Mkm}^3$; frequently above 1 Mkm^3), mainly mafic (-ultramafic) magmatic events of intraplate affinity (based on tectonic setting and/or geochemistry) that occur in both continental and oceanic settings, and are typically either of short duration ($<5 \text{ Ma}$) or consist of multiple short pulses over a maximum of a few 10s of Myr. LIPs comprise volcanic packages (flood basalts) and a plumbing system of dyke swarms, sill complexes, layered intrusions, and a magmatic underplate (Coffin and Eldholm, 1992; Bryan and Ernst, 2008; Ernst, 2014).

Numerous diabase dyke swarms and associated magmatism, mostly of Proterozoic age, intrude the Superior craton, and are thought to record periodic LIP magmatism (e.g., Ernst and Bleeker, 2010). One of the more dramatic LIP remnants occurs in the Cape Smith Belt in the northernmost part of the Quebec promontory of the Superior craton. The Cape Smith Belt is dominated by the Povungnituk basalts (Hynes and Francis, 1982; Francis et al., 1983), interpreted as continental flood basalts, and associated dolerite sills which are broadly constrained in age between 2.04 and 1.91 Ga (Machado et al., 1993), and the younger ca. 1880 Ma Chukotat event (Wodicka et al., 2002; Randall, 2005; Bleeker and Kamo, 2018) (Fig. 4.1).

The volcanics of the Povungnituk Group of the Cape Smith Belt, in particular in the western part, have been linked to various other magmatic units in the Superior craton to the south based on geochronological and geochemical evidence (Schmidt, 1980; Schwarz and Fujiwara, 1981; Parrish 1989; Legault et al., 1994; Buchan et al., 1998, 2007). These include the Minto dykes, Eskimo volcanics, Nastapoka basalts and Persillon volcanic rocks in the Richmond Gulf area, and Inukjuak dykes. Halls and Davis (2004) viewed these units as the product of a 2.0 Ga rifting event, located beneath modern day Hudson Bay, that caused a relative rotation of the eastern and western Superior craton. The magmatic event associated with that rifting may correspond to what we are describing herein as the Minto-Povungnituk LIP, which includes the Minto and Inukjuak dyke swarms east of Hudson Bay, Eskimo volcanics of the Belcher Islands, Nastapoka basalts and Persillon volcanic rocks of the Richmond Gulf, Povungnituk volcanics and the Watts Group of the Cape Smith Belt. Scott et al. (1999) and Hamilton et al. (2016) have suggested a mantle plume as the context for this 2 Ga magmatism. Geochemistry will help assess whether a mantle plume was involved and whether the magmatism is associated with thinned lithosphere (which could also support the role of rifting), and to identify any spatial variations in chemistry that can be related to different mantle sources.

To provide this geochemistry assessment, we have reanalyzed material previously collected from the Povungnituk Group. We have utilized high quality geochemistry data from the literature for the Watts Group, Minto dykes, and Eskimo Formation (Scott et al., 1991, 1999; Hegner and Bevier, 1991; Ernst and

Buchan, 2010) and have obtained Nd isotopes for the Minto dykes that have not been available before.

In summary, this research provides the precise age described in Section 2.3.1 and new geochemical data for the Povungnituk Group, links it with additional coeval units into a regional LIP, and then integrates the chemistry of all the units in order to develop a petrogenetic and geodynamic model for the emplacement of the LIP.

4.2 Samples

Samples for geochemistry and petrography from the mafic Povungnituk Group (MPG) were collected by W.R.A. Baragar during a 1973 field season with the Geological Survey of Canada. The traverses are shown in Figure 4.1 and certain geochemical indicators are shown superimposed on a stratigraphic column for all three traverses in Figure 4.2. Petrography and major element chemistry of the three traverses (Figs. 4.1, 4.2) were published in Baragar (2015, 2017). The samples were collected over three transects through the MPG, spanning an inferred volcanic thickness of 3 km; although fault duplication has been proposed (St-Onge et al., 1992), it has been considered unlikely by Mungall (2007) and Bleeker and Ames (2017) (see discussion in Section 3.5.1.2). Baragar's 84 samples (reanalyzed in our study) were collected as chips over a distance of roughly one meter in homogenous units for each geochemistry sample and as selected large hand samples for geochronology. In order to

address the above research objectives, new Nd isotopes and a new U-Pb age were produced for a selected suite of the MPG.

In addition, geochemical data for seven samples from four WNW-NW trending Minto dykes were included. These data were from Ernst and Buchan (2010), who analyzed samples from Buchan et al. (1998). These Minto dyke samples were then further analyzed for Nd-isotopes as part of this research (see below).

Several samples were processed for U-Pb dating but only one sample yielded baddeleyite. This dated sample (BLS-73-31) is from the lower part of a dolerite sill at the very top of exposed western Povungnituk Group rocks. It is separated from the Chukotat lavas to the north by a highly contorted shale unit that presumably marks the presence of a thrust fault somewhere in close proximity.

Figure 4.1. Geological map of the Cape Smith belt, northern Quebec, Canada (after St-Onge et al., 2004). Stars represent the location of the Lac Leclair alkaline complex from Baragar et al. (2001) and the Kenty Lake alkaline complex from Gaonac'h et al. (1992). Small map shows the Superior craton in grey after Goodfellow (2007) and a red box indicated the map location. Also shown is the location of sample BLS-73-31 used to date the Beuparlant Formation (section 2.3.2.1). Traverses are from Baragar (2015, 2017).

4.3 Methodology

4.3.1 Major and trace element analyses

All 84 sample powders from the Povungnituk suite, including an internal standard (10–LT–05, a basaltic andesite from Lake Tahoe, California) were sent to ALS Geochemistry laboratories in North Vancouver, British Columbia for major element analysis via Inductively Coupled Plasma Atomic Emissions Spectrometry (ICP–AES) as well as trace and rare earth element (REE) analysis via Inductively Coupled Plasma Mass Spectrometry (ICP–MS). Major- and trace-element analytical uncertainties (2σ) for ALS Geochemistry laboratories based on repeat analysis of the Lake Tahoe internal standard are reported in the Supplementary Data.

4.3.2 Sm–Nd isotopes analysis

Rock powders were spiked with a mixed ^{148}Nd – ^{149}Sm spike before being dissolved in a mixed solution of ~29 M HF and ~16 M HNO_3 . The samples were then dried down on a hotplate before being redissolved with 8 M HNO_3 and 6 M HCl sequentially. The dried residue of each sample was finally dissolved in 2.5 M HCl prior to being loaded onto 14 ml Bio-Rad borosilicate glass chromatography columns containing 3.0 ml of Dowex AG50W–X8 cation resin. REE were eluted with 6 M HCl. REE fractions were then dissolved in 0.26 M HCl and loaded onto Eichrom Ln Resin chromatographic columns containing Teflon powder coated with HDEHP [di(2-ethylhexyl) orthophosphoric acid] (Richard et al., 1976).

Neodymium was eluted with 0.26 M HCl, followed by Sm with 0.5 M HCl. The isotope ratios were measured using a Thermo Finnigan Triton thermal ionization mass spectrometer (TIMS) housed at the Isotope Geochemistry and Geochronology Research Centre (IGGRC), Carleton University, Ottawa, Canada. Neodymium and Sm fractions were loaded with H₃PO₄ onto one side of a double rhenium filament assembly. The isotope ratios were measured at temperatures of 1700–1800°C and are normalised to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$. An IGGRC's in-house Nd standard was routinely measured and an average value of the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio was 0.511826 ± 0.000007 (1σ) over a period of three years; this value is equivalent to $^{143}\text{Nd}/^{144}\text{Nd} = 0.511855$ reported for the La Jolla Nd standard. Sm and Nd concentrations were measured precisely within 1%, whereas $^{147}\text{Sm}/^{144}\text{Nd}$ ratios are reproducible to 0.5%. Analyses of the USGS standard BCR-2 yield Nd = 28.53 ppm, Sm = 6.618 ppm, and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512643 \pm 0.000011$ ($n = 13, 2\sigma$). Total procedural blanks for Nd are less than 50 picograms.

4.4 Results

4.4.1 *Field observations*

The MPG consists of mainly massive, basaltic flows with rare pillows and sills interlayered with black shale and greywacke, tuff, pyroclastic rocks and, in places, siliceous material (St-Onge et al., 1992; Mungall, 2007). Northward (up section), the MPG is increasingly dominated by pillowed and massive tholeiitic basalt with minor interflow black shale, cherty sediment, tuff and agglomeritic tuff (Fig. 4.2). This study focuses on the basalts and associated dolerites.

The general petrography of the three traverses across the Cape Smith Belt are summarized here. Both volcanic rocks and dolerites are present, the latter probably reflecting associated sills or thick flows. The dolerite samples commonly preserve a vestige of igneous texture mainly in the distribution of the mafic minerals and feldspar, and most samples have little or no foliation. All doleritic rocks are at greenschist facies and the mineralogy is typical of this low metamorphic grade. Sodic plagioclase is intergrown with actinolite and chlorite, in roughly equal proportions, epidote is subordinate, and titanite, commonly enclosing relics of opaque minerals, is accessory. Sparsely distributed blebs of chlorite observed in some dolerite sills are interpreted as pseudomorphs after olivine. The volcanic rocks are of similar mineralogy, but finer grained, and are commonly sparsely plagioclase phyric. They generally show some degree of foliation development.

Figure 4.2. Selected element ratios along the eastern and central traverses shown in Fig. 4.1. Stratigraphic columns modified from Baragar (2017). Larger symbols indicate samples identified as flows in the field. Rare earth elements normalized to chondrite. Normalizing values from Sun and McDonough (1989). Owing to the presence of anticlines and synclines (see symbols) the vertical scale is not linear.

Figure 4.2 (continued). Selected element ratios along the Western traverse shown in Fig. 4.1. Stratigraphic column modified from Baragar (2015). Larger symbols indicate samples identified as flows in the field. No structural units were mapped, so approximated locations of folds and faults are overlain from St-Onge et al. (2004). Rare earth elements normalized to chondrite. Normalizing values from Sun and McDonough (1989). Owing to the presence of anticlines and synclines (see symbols) the vertical scale is not linear.

4.4.2 Major and trace elements

The results for major and trace elements are shown for representative samples in Table 4.1. Element mobility during metamorphism needs to be considered for the MPG. The entire belt has undergone greenschist to lower amphibolite facies metamorphism (St-Onge et al., 1992) and the rocks commonly have LOI >2 wt. %. Most element concentrations therefore need to be viewed with caution. Other elements such as the high field-strength elements (e.g., Nb and Zr) and REE (e.g., La and Sm), particularly their ratios, have been shown to be relatively immobile and thus are believed to preserve a record of the magmatic processes (e.g. Pearce, 2008). In comparison with these immobile elements, some major and trace elements in the MPG display varying degrees of scatter and have been avoided for geochemical interpretations in this study. These include Si, Ca, K, Na, Rb, Sr, Cs, Ba and U. Figure 4.3 shows selected elements against Zr to display the varying degrees of scatter.

Figure 4.3. Representative bivariate diagrams to show the difference between degrees of scatter. Elements not affected by alteration are shown on the left and elements affected by alteration are shown on the right. R^2 values above 0.2 have been used as a marker for major element concentrations. Possible effects of alteration on trace elements have been determined through behaviour against zirconium. R^2 values for the complete dataset can be found in Appendix 4 and 5.

The chemical compositions of the MPG lavas are quartz-normative, Fe and Ti-rich basalts. Their relatively evolved nature is indicated by their low MgO contents (range ~6 wt. %, with maximum of ~10 wt. %) and elevated Fe₂O₃ contents (range ~17 wt. %). Al₂O₃ content is positively correlated with Mg# (Fig. 4.4a). Zirconium and Sc concentrations are negatively correlated with Mg# (Fig. 4.4c, 4.4d). On the basis of Zr/Ti vs. Nb/Y plots (Pearce, 1996) the MPG varies in composition from basalt to alkaline basalt (Fig. 4.5a). Rare earth element diagrams display a large variation of trends, with samples showing E-MORB-like La/Yb_C trends (La/Yb_C between 1 and 2) and fanning towards OIB-like trends (La/Yb_C between 2 and 10) (Fig. 4.5b). This is supported by primitive mantle normalized incompatible trace element patterns, where MPG samples vary between E-MORB and OIB as characterized by Sun and McDonough (1989) (Fig. 4.5c). In a Th/Yb vs. Nb/Yb plot (Pearce, 2008) (Fig. 4.5d), all MPG samples fall along the mantle array, forming a trend between E-MORB and OIB end member compositions. The absence of elevated Th/Yb ratios (Fig. 4.5d) indicates no or only minor interaction with crustal material. Th/Nb_{PM} ratios are also a good monitor of crustal input (e.g., Wang et al. 2007) and Th/Nb_{PM} data for MPG sample (not shown) also show no correlation with Mg# and form a cloud between 0.6 and 1. Thus, based on major and trace element behaviour two compositional end members have therefore been involved in the formation of the MPG and can be identified as a E-MORB-like group showing patterns that are indicative of having formed from melting of depleted upper Proterozoic mantle and a more enriched OIB-like group.

Figure 4.4. Selected elements for the mafic Povungnituk Group (MPG) versus Mg-number (Mg#: molar MgO/(MgO + FeO^T)*100). Samples fall into two fields, representing the depleted and enriched endmembers of the MPG. This separation can be seen in flows as well as sills. Samples for which the mode of emplacement could not be identified are shown as clouds in the background. The dated sample refers to sample BLS-73-31.

Figure 4.5. Trace element geochemical diagrams for the mafic Povungnituk Group (MPG). (a) Zr/Ti vs. Nb/Y (after Pearce, 1996) used for rock type identification where samples vary from basalt to alkaline basalt, (b) Chondrite normalized Rare Earth element diagram, colors correspond to the labels in the legend. Samples form a wide array from mild slopes of the depleted endmember to more prominent slopes for the enriched endmember. (c) Multielement diagram for incompatible elements. Samples show a wide array of compositions with varying slopes and no significant anomalies, colors correspond to the labels in the legend. Primitive Mantle values from Sun and McDonough (1989), Chondrite values of Sun and McDonough (1989). Dated sample corresponds to sample BLS-73-31.

Figure 4.5 (continued). Trace element geochemical diagrams for the mafic Povungnituk Group (MPG). (d) Th/Yb vs. Nb/Yb (after Pearce, 2008), where all samples fall within the defined mantle array ranging from E-MORB compositions towards OIB compositions. Average lower crust (LC), upper crust (UC) middle lower crust (MC) are from Rudnick and fountain (1995). (e) Tb/Yb_{PM} vs. Zr (after Wang et al., 2002), indicating the transition zone between garnet- and spinel-lherzolite. Samples fall along a wide array spanning both stability fields indicating gradual shallowing of the melt source or mixing of a shallow and a deeper mantle melt. Primitive Mantle values from Sun and McDonough (1989). Dated sample corresponds to sample BLS-73-31.

As shown in Figures 4.4 and 4.5, samples that can be clearly defined as sills overlap geochemically with the samples that can be clearly identified in the field as flows, as well as with the samples for which no clear emplacement method can be determined. This overlap occurs in the depleted as well as the enriched samples. The dated sample (BLS-73-31), corresponds to the enriched type of MPG, and indicates that the dated unit is part of the MPG magmatic event.

The ratio Tb/Yb_{PM} , normalised to primitive mantle values ("PM"; Sun and McDonough, 1989), can be used to show the depth of partial melt formation (Wang et al., 2002). Primitive mantle-normalized Tb/Yb ratios above 1.8 are interpreted to show melting of the mantle within the garnet stability field, whereas Tb/Yb ratios below 1.8 signify melting in shallower (<75 km) spinel lherzolite, although values close to 1.8 likely melted both garnet and spinel lherzolitic components (Wang et al., 2002) (Fig. 4.5e). The E-MORB-like group of the MPG falls uniformly into the spinel lherzolite field whereas the OIB-like samples form a continuous trend between the spinel stability field and the garnet stability field, with normalized Tb/Yb ratios ranging between 1.2 and 2.2. This shows that melt generation for the OIB group occurred over a range of depths or represents mixing of melts from different depths.

Table 4.1. Representative geochemical data from the mafic Povungnituk Group.

Sample	<u>BLS 57</u>	<u>BLS 55</u>	<u>BLS 34</u>	<u>BLS 53</u>	<u>BLS 183</u>	<u>BLS 31</u>	<u>BLS 264</u>	<u>BLS 182</u>
	<u>73</u>	<u>73</u>	<u>73</u>	<u>73</u>	<u>73</u>	<u>73</u>	<u>73</u>	<u>73</u>
Location	K1-28A	K1-30	K1-07	K1-32	C1-41	K1-04	N4-02	C1-40
Traverse	West	West	West	West	Central	West	East	Central
Group	Depleted	Depleted	Depleted	Enriched	Enriched	Enriched	Enriched	Enriched
Easting	401889	402205	396393	402874	462536	395889	519628	462608
Northing	6779706	6779411	6781526	6779219	6801174	6781752	6815745	6801726
SiO ₂	47.40	48.50	48.00	46.60	46.40	44.80	43.00	40.50
Al ₂ O ₃	13.50	13.10	13.30	12.50	13.95	12.85	14.25	14.40
Fe ₂ O ₃	13.45	14.85	14.65	15.70	13.30	16.85	16.75	13.60
CaO	12.30	10.45	10.50	10.05	9.01	8.23	8.46	8.51
MgO	6.30	6.50	7.02	6.21	8.06	5.34	6.82	3.69
Na ₂ O	1.23	2.56	2.44	2.34	2.99	2.83	2.41	4.76
K ₂ O	0.02	0.14	0.12	0.23	0.17	0.57	0.61	0.53
TiO ₂	1.03	1.37	1.52	1.77	1.84	3.03	3.63	3.07
MnO	0.22	0.21	0.20	0.23	0.25	0.24	0.27	0.34
P ₂ O ₅	0.07	0.12	0.11	0.16	0.12	0.37	0.57	0.42
LOI	4.12	2.06	2.84	2.38	2.94	3.01	2.93	8.31
Total	99.68	99.9	100.76	98.24	99.12	98.18	99.78	98.18
Mg#	28.77	27.40	29.24	25.43	34.32	21.46	25.98	18.96
Ba	5.3	59.4	22.2	58.7	262	199.5	164	213
Ce	10.4	14.4	16.1	22.5	23.6	50.5	67.4	50.6
Dy	3.79	4.18	4.31	5.46	3.41	5.93	7.52	4.93
Er	2.25	2.78	2.25	3.36	2.09	3.26	4.16	2.36
Eu	0.93	1.24	1.38	1.6	1.25	2.55	2.69	2.5
Ga	16.5	21.5	18.8	21.4	18.6	25.8	26	20.9
Gd	3.4	4.27	4.24	5.43	3.91	7.23	8.74	6.29
Hf	1.8	2.3	2.6	3.3	2.1	5.4	6	4
Ho	0.87	0.97	0.89	1.19	0.7	1.3	1.51	0.93
La	4.2	5.8	6.3	9.4	10.3	20.9	28.4	19.8
Lu	0.31	0.42	0.32	0.45	0.26	0.45	0.65	0.31
Nb	3.1	4.9	5.5	8.8	8.7	22.8	33.9	31.3
Ta	0.2	0.3	0.3	0.5	0.5	1.3	1.5	1
Nd	7.9	11	12.3	15.9	15	31.9	42.5	32.3
Pb					2		4	2
Pr	1.59	2.22	2.5	3.36	3.19	6.98	9.42	7.09
Rb		2.9	1.4	3.6	2.3	12.9	18.2	7.5
Sc	41	40	40	39	39	37	34	19
Sm	2.67	3.56	3.54	4.73	3.76	7.11	9.58	7.27
Sr	222	194	325	321	368	216	353	244
Tb	0.57	0.75	0.67	0.86	0.63	1.12	1.27	0.9
Th	0.36	0.51	0.41	0.86	0.7	1.79	1.81	2.45
Tm	0.36	0.4	0.35	0.49	0.29	0.46	0.58	0.33
U	0.13	0.15	0.12	0.24	0.15	0.42	0.41	0.74
V	329	420	399	458	333	548	361	245
Y	20.2	25.3	21	29.7	18	31.1	38.2	23.8
Yb	2.43	2.51	1.94	2.96	1.63	3	3.82	2.14
Zr	63	85	85	118	72	220	249	159
Cr	80	70	100	140	140	80	120	
Co	53	51	57	49	52	52	40	32
Cu	164	148	156	117	117	75	17	34
Ni	74	62	72	71	90	32	49	16
Zn	100	93	103	136	99	139	135	115

Coordinates in NAD27, Zone 18V

Major element compositions in wt.% and trace element compositions in ppm

Full dataset in Appendix 6

4.4.3 Sm-Nd isotopes

The Nd isotope analyses are calculated using the 1998 Ma U-Pb age reported for the MPG and Minto dykes. Results are shown in Table 4.2. The initial Nd ratios for MPG range from $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ of +2.1 to +3.8 and form a continuous trend between a depleted and an enriched endmember (Fig. 4.6). The dated sample (BLS-73-31) falls in the middle of this trend with an $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ value of +2.56. Neodymium Depleted Mantle model ages range between 2.2 Ga and 2.3 Ga. The initial ϵ_{Nd} values for Minto dykes range from -3.7 to -7.8.

Table 4.2. Summary of Sm-Nd results for the mafic Povungnituk Group and the Minto dykes

Formation	Sample	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\epsilon_{\text{Nd}}^{2.0 \text{ Ga}}$
MPG	BLS 31 73	7.67	32.98	0.1407	0.512031	2.56
MPG	BLS 41 73	6.08	25.15	0.1463	0.512111	2.67
MPG	BLS 55 73	3.33	10.85	0.1855	0.512685	3.8
MPG	BLS 180 73	5.92	26.63	0.1345	0.511925	2.06
MPG	BLS 183 73	3.71	15.11	0.1483	0.510216	3.24
MPG	BLS 197 73	3.40	11.44	0.1798	0.512590	3.42
MPG	BLS 264 73	8.89	40.44	0.1329	0.511902	2.02
MPG	SAB87-179	3.83	15.29	0.1513	0.512171	2.56
Minto	BXA91-7904	6.87	30.06	0.1381	0.511677	-3.73
Minto	BXA91-8103	6.91	31.57	0.1324	0.511559	-4.57
Minto	BXA91-8203	6.15	29.79	0.1248	0.511290	-7.87
Minto	BXA91-8406	4.81	20.38	0.1426	0.511668	-5.06

2 σ uncertainties of $^{147}\text{Sm}/^{144}\text{Nd}$ are 0.5 %

2 σ uncertainties of $^{143}\text{Nd}/^{144}\text{Nd}$ are 0.000011, given by the reproducibility of the standard
 ϵ_{Nd} values are calculated relative to a present day $^{143}\text{Nd}/^{144}\text{Nd}$ CHUR values of 0.512630

Figure 4.6. Sm-Nd isotopes of mafic Povungnituk Group (MPG) samples and Minto dykes. (a) $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{147}\text{Sm}/^{144}\text{Nd}$. Samples fall on a line with a slope age of 2.225 Ga that intersects the ordinate at an initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.50996 corresponding to a $\epsilon_{\text{Nd}}^{2.0 \text{ Ga}}$ of -1.78. These differences to the obtained U-Pb age of 1998 Ma and the $\epsilon_{\text{Nd}}^{2.0 \text{ Ga}}$ values obtained on single analyses calculation show that they values have been modified (b) $\epsilon_{\text{Nd}}^{2.0 \text{ Ga}}$ vs. $1/\text{Nd}$ for the mafic Povungnituk Group (MPG), where samples show radiogenic Nd-isotopic ratios and fall along a linear array indicating mixing of two endmembers. Dated sample corresponds to sample BLS-73-31. (c) $\epsilon_{\text{Nd}}^{2.0 \text{ Ga}}$ vs. $1/\text{Nd}$ for the Minto dykes, where samples show unradiogenic Nd-isotopic ratios.

4.5 Discussion

4.5.1 Link with coeval regional units and recognition of a 1998 Ma

Minto-Povungnituk LIP

The new age constraints suggest links with a number of mafic magmatic units in the Quebec promontory of the Superior craton (Fig. 4.7). The Minto dykes in the interior of the craton have a U–Pb baddeleyite-zircon age of 1998 ± 2 Ma (Buchan et al., 1998). The Lac Shpogan dyke of the James Bay area of the eastern Superior craton has been recently correlated with the Minto dykes based on a U-Pb baddeleyite age of 1999 ± 2 Ma (Hamilton et al., 2016).

An identical age was obtained from the Watts Group (Purtiniq) ophiolite in the northern Cape Smith Belt (Parrish, 1989; Scott et al., 1999). The Watts Group comprises layered ultramafic and mafic rocks, clinopyroxenite intrusions, sheeted mafic dykes and gabbros, and pillowed and massive mafic volcanic rocks intruded by rare felsic sills and dykes (St-Onge et al., 1992; Scott et al., 1991; 1999). The rocks that comprise the ophiolite are separated from the underlying sedimentary rocks by a south-verging thrust fault. Two samples of Watts Group mafic cumulates were dated using zircons. In a metagabbro, zircons yield a U-Pb age of 1998 ± 2 Ma, whereas in a metamorphosed anorthositic layered gabbro, two types of zircons are present. Igneous cloudy grains are 1995–2000 Ma whereas clearer overgrowths are 1977 ± 3 Ma, perhaps related to seafloor metamorphism (Scott et al., 1999).

In recent years, the use of precise U-Pb geochronology has led to the recognition of many new LIPs and revealed their wide distribution within continental blocks (e.g., Ernst et al. 2013; Samal et al. 2019; Baratoux et al. 2019). The standard approach has become to correlate intraplate units in a region on the basis of their U-Pb age matches and then to subsequently use geochemistry (mainly trace elements and isotopes) to interpret the distribution of distinct geochemical types. Geochemistry on its own has been shown to be an unreliable guide to identifying units that belong to the same LIPs, given that many LIPs exhibit multiple geochemical types (Ernst, 2014).

Therefore, the overlap in age determinations as discussed above suggests that a 1998 Ma magmatic event was widespread in and around the Quebec promontory, the eastern part of the Superior craton, covering an area over 400,000 km² (Fig. 4.7). An additional unit, the Eskimo volcanics, are correlated on the basis of paleomagnetism with the dated Minto dykes (Buchan et al., 1998), and together all these linked units are referred to as the 1998 Ma Minto-Povungnituk LIP. Additionally, the Nastapoka basalts in the Richmond Gulf area have been paleomagnetically linked to the Minto and Lac Shpogan dykes (Buchan et al., 1998), and the Persillon Formation in the Richmond Gulf area has been geochemically linked to the Eskimo Formation and the undated Inukjuak dykes (Legault et al., 1994).

Figure 4.7. Map showing coeval magmatic units of the ca. 1998 Ma Minto-Povungnituk large igneous province as red fields or solid red lines. Also shown are the Inukjuak dykes as red dotted lines, which have been provisionally correlated with the Eskimo Formation, Nastapoka basalts and Persillon volcanic rocks based on their geochemical signatures (Legault et al. 1994) but otherwise remain undated. Small map shows the outline of North America and the location of the Superior craton with the map position highlighted in red. Compiled from Chandler (1984); Buchan et al. (1998); St. Onge et al. (2004); Baragar (2007); Goodfellow (2007); Maurice et al. (2009).

4.5.2 Differences between chemistry of northern and southern portions of the Minto-Povungnituk LIP

The geochemistry of the different units within this Minto-Povungnituk LIP reveals regional compositional and isotopic differences that split the proposed LIP into “northern” and “southern” groups that were emplaced at approximately the same time.

4.5.2.1 Northern magmatism (Povungnituk – Watts Group connection)

Trace element concentrations for the MPG vary between two distinct endmembers (Fig. 4.5d). Neodymium isotopic ratios show correlations with element concentrations (Fig. 4.6b) and indicate that the two endmember compositions represent individual mantle compositions.

Stratigraphy along the three traverses sampled for the MPG vs. diagnostic element ratios show no correlation with increasing stratigraphic level. Instead, they vary between the two geochemical signatures (Fig. 4.2). The clearest distinction is seen in the westernmost traverse (Fig. 4.2, western traverse). A pattern like this across the stratigraphy would be in accordance with magma being derived alternately from two staging chambers (Lightfoot and Hawkesworth, 1997). Overlain mapping by St-Onge et al. (2004) shows possible locations of a syncline and anticline pattern that roughly follows the observed geochemical pattern. If these structural units are accurate, no assessment about the possibility of mixing can be made using the stratigraphy of the western

traverse. A constant variation between the two chemical compositions can also be observed in the lower section of the eastern traverse. This area represents a mixture of basaltic flows and sills. Only two samples have been clearly identified as flows and belong both the E-MORB-like endmember. The exact emplacement method for each individual sample is not known. The possibility of mixing can therefore not be accurately assessed using chemo-stratigraphic observations of the Povungnituk Group.

Two geochemically distinct magmatic suites have been identified within the layered mafic and ultramafic rocks of the Watts Group (Fig. 4.8) on the basis of major and trace element abundances and Nd isotopic composition (Scott et al., 1991, 1999; Hegner and Bevier, 1991). An E-MORB-like suite is tholeiitic and characterized by low abundances of elements such as Ti, Zr, Y, and the light REE, has $\text{La/Sm}_{\text{PM}} \sim 1.1$, and has $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ values that range from +3.0 to +4.7 (Hegner and Bevier, 1991). Both of the aforementioned gabbro samples for which U-Pb ages were determined belong to this suite.

The second suite is referred to as tholeiitic OIB-like (Scott et al., 1991, 1999). It has incompatible element abundances that are enriched relative to MORB (Scott et al., 1991), normalized incompatible element patterns that peak at Nb and Ta as do modern OIB (Stracke et al., 2005), and $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ values that range from +2.7 to +3.1 (Hegner and Bevier, 1991). The absolute age of this suite has not been directly determined but has been assumed to also be ca. 2000 Ma. The sheeted dykes of the two geochemical groups are indistinguishable from each other in outcrop and do not show consistent cross-cutting relationships,

suggesting that their emplacement was coeval with the cumulate rocks (Scott et al., 1999).

Two distinct mantle sources (E-MORB- and OIB-like) with different isotopic compositions were involved in the generation of the Watts Group (Purtunig) ophiolite. Each source produced a suite of cumulate rocks and sheeted mafic dykes. The first one is a trace element depleted source similar to E-MORB. Scott et al. (1991) interpreted this source to have been long-lived and to have been enriched metasomatically immediately prior to melt generation to form the more enriched OIB-like suite.

The two distinct geochemical groups outlined by the Watts Group (Scott et al., 1991, 1999) correlate well with the two groups in the MPG. All samples can be classified as basalts. In a diagram of Th/Yb vs. Nb/Yb (Pearce et al., 2008) the Watts sheeted dykes groups overlap with the two endmembers (E-MORB and OIB-like) of the MPG, whereas the volcanic rocks of the Watts Group overlap with the E-MORB endmember of the MPG (Fig. 4.8). This behaviour is apparent for both major and trace elements. An exception is the Gd/Yb ratios. The low ϵ_{Nd} sheeted dyke samples of the Watts Group generally show the highest Gd/Yb ratios. This indicates that melting of the low ϵ_{Nd} member of the Watts Group might have occurred at an even greater depth than for the majority of the OIB-like samples of the MPG.

Figure 4.8. Geochemical comparison of Depleted and Enriched suites for the mafic Povungnituk Group (MPG) and the low and high ϵ_{Nd} Watts Group. (a) Th/Yb vs. Nb/Yb (Pearce, 2008). All samples from the Watts Group fall within the mantle array and overlap with the observed compositions for the MPG ranging from E-MORB towards OIB. (b) ϵ_{Nd} 2.0 Ga vs. 1/Nd, where the Watts Group samples fall along the array defined by the MPG, indicative of mantle mixing. (c) Primitive mantle normalized trace element diagram. Watts Group samples separate into two distinct compositions overlapping with the enriched and depleted endmember compositions of the MPG. Primitive mantle values from Sun and McDonough (1989).

Another option would be that slightly different material has been melted to produce the Watts Group compared to the MPG magmas. Several lines of evidence indicate that the modern mantle contains a significant fraction of pyroxenites (Schulze, 1989). These may play an important role in controlling the chemical variability of mantle-derived melts (Sobolev, 2007) and have been shown to be incorporated into OIB (Kogiso et al., 2003; Peterman and Hirschmann, 2003; Kogiso and Hirschmann, 2006) and MORB (Lambart et al., 2009). In a diagram of Ce/Yb vs. Yb it is evident that, instead of different melting depths, pyroxenite melting could have been involved in the generation of the Watts Group, resulting in the observed HREE signature (Fig. 4.9). This would open the possibility that the northern domain has been melted at the same depth but definitive constraints on the depth of melting cannot be made.

Broadly, the Watts Group Nd isotope values overlap with those of MPG. The low ϵ_{Nd} sheeted dykes and volcanic rocks of the Watts Group have values that overlap with but tend to be more depleted than MPG (+3 to +6.1). They form no trend towards less depleted $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ values but correlate well with the trend shown by MPG in a diagram of $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ vs. $1/\text{Nd}$. The low ϵ_{Nd} sheeted dykes of the Watts Group roughly overlap with the OIB-like samples of the MPG. Overall, the Watts Group displays the same mixing trend between two distinct mantle endmembers as the MPG.

Figure 4.9. Ce/Yb vs. Yb concentration showing non-modal, batch partial melting models (Shaw, 1970) for spinel peridotite, garnet peridotite (black lines), and garnet pyroxenite (blue line). Tic marks represent 1 % steps between 1 and 10 % melting and 10 % steps above 10% melting. For peridotite models, starting compositions assume primitive mantle (PM) concentrations; modal abundances, and D values from Salters and Stracke (2004). For the garnet pyroxenite model, starting composition is N-(Normal) MORB with modal abundances, and D values from Petermann et al. (2004). Shown as fields are the most primitive samples of each part of the Minto-Povungnituk large igneous province (LIP).

4.5.2.2 Southern magmatism (Minto – Eskimo connection)

The geochemistry of the Minto dykes does not show either of the two distinct groups observed in the MPG and Watts Group samples (Fig. 4.10). The Minto dykes all plot in the basalt field shown in a Zr/Ti vs. Nb/Y (Fig. 4.10a) diagram and all plot in the field described for plate margin basalt in a diagram of Zr/Y vs. Ti/Y (Fig. 4.10b).

In a Th/Yb vs. Nb/Yb (Fig. 4.10c) diagram half of the samples plot within the mantle array, overlapping with the E-MORB field and the transitional samples described for the MPG. Although showing similar La/Yb ratios, Minto dyke samples have slightly higher La/Sm_c (c = normalized to chondritic abundances) ratios against La/Yb_c than MPG (Fig. 4.10d). Two samples plot slightly outside of the mantle array showing that interaction with crustal material might have been involved in their formation. These samples also display a negative Nb-anomaly in primitive mantle-normalized multi-element diagrams (Fig. 4.10e). Minto dyke REE compositions are slightly different from compositions observed in the MPG.

Figure 4.10. Trace element geochemical diagrams for the Minto dykes compared with the mafic Povungnituk Group (MPG). (a) Zr/Ti vs. Nb/Y (after Pearce, 1996), used to identify the rock type. All samples fall within the basalts field. (b) Zr/Y vs. Ti/Y (after Pearce and Gale, 1977) used to distinguish between plate margin basalts and within-plate basalts. All samples fall into the field defined for plate margin basalts. (c) Th/Yb vs. Nb/Yb (after Pearce, 2008), where samples form an array within the mantle array at E-MORB compositions outside of the mantle array, indicating contamination. Lower crust (LC), middle cst (MC) and upper crust (UC) from Rudnick and Fountain (1995) (d) La/Sm_c vs. La/Yb_c, chondrite values from Sun and McDonough (1989), showing different trends defined by the MPG and the Minto dykes. (e) Multi-element diagram for immobile elements, showing different patterns, with Nb-Ta anomalies, steeper La to Sm ratios and shallower Ti to Lu ratios compared to MPG samples. Chondrite values from McDonough and Sun (1995).

Figure 4.10 (continued). (e) Multi-element diagram for immobile elements, showing different patterns, with Nb-Ta anomalies, steeper La to Sm ratios and shallower Ti to Lu ratios compared to MPG samples. Primitive Mantle values from Sun and McDonough (1989).

The most significant geochemical difference between Minto dykes and MPG can be seen in their Nd isotopes. Minto dykes have initial $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ values that are highly negative (-3.7 to -7.9) (Fig. 4.11a) in contrast to the positive $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ values (+2.0 to +3.8) for the MPG. Therefore, although matching in age and interpreted to belong to the same LIP (as discussed above), MPG and Minto dykes must have been fed from vastly different sources or have undergone different magmatic evolution processes since magma segregation from the mantle. To assess if this spread in Nd-isotope geochemistry is source-related or related to magmatic processes such as mixing or assimilation, we have to observe how these Nd-isotopic differences correlate with differences in major and trace element ratios. As illustrated in Figure 4.11a the spread in Nd-isotopic composition is negatively correlated with Mg#, meaning the lower $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ values of the Minto dykes are correlated with higher Mg#. If the samples would have been contaminated by felsic rocks during fractionation, the lower ϵ_{Nd} values would be correlated with lower Mg#.

Figure 4.11. Geochemical diagrams for Minto dykes and Eskimo basalts. (a) $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ vs. Mg-number (Mg#: molar $\text{MgO}/(\text{MgO} + \text{FeO}^T) \times 100$). Minto dyke samples form a trend. Possible plot locations for the Eskimo Formation are indicated. (b) Nb/Zr vs. Zr, where higher Nb/Zr ratios correlate with higher Zr ratios, opposite to expected behaviour during assimilation and fractional crystallization (AFC). (c) Primitive mantle normalized Th/Nb vs. Mg#. Th/Nb ratios serve as a proxy for contamination. Samples fall on a trend where the most primitive samples show the highest degree of contamination, opposite to the expected behaviour during AFC. (d) $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ vs. 1/Nd. Samples show a linear trend associated with contamination. Primitive Mantle values from Sun and McDonough (1989).

Explanations for such observations have been listed by Devey and Cox (1987) in their study of the 66 Ma Deccan Traps LIP event (India). They concluded that either coincidental causes (contaminants become less available over time, or changes of flow rate in dykes) need to be constrained or that magmatic temperature is the controlling influence for the contamination signature. The correlation between Mg-number and Nd isotopes is then directly linked to the magma source before dyke emplacement, as crustal contamination fades out because the magmas become too cool. This is essentially an assimilation-equilibrium crystallization (AEC) process, which demands the suppression of fractional crystallization (Devey and Cox, 1987). Hence, even if a fractionation chamber existed, it cannot be the main site of contamination. Two plausible explanations for this have been proposed.

The existence of magma chambers with high surface-to-volume ratios, which allow a large surface area of magma to come into contact with the country rock, and so permit partial fusion of the more acidic or volatile-rich portion of the crust over relatively short time scales (Huppert and Sparks, 1985) or a chamber located within the upper part of the mantle rather than the crust, where any interaction with wall rocks did not result in identifiable isotopic or trace element signatures (Devey and Cox, 1987). In both cases any measurable contamination would occur during magma ascent in the crust rather than in a magma chamber. Contamination is then dependent on the type of flow in the magma conduit. Under laminar flow the initial formation of a chilled margin around the conduit would shield later magmas from contamination. Huppert and Sparks (1985) and

Campbell (1985) have shown however, that basaltic magma can ascend turbulently if the flow rate is sufficiently high ($>25 \text{ m}^2/\text{s}$) and the conduit width great enough ($>3 \text{ m}$ in primitive magmas). Under such circumstances a chilled margin is less likely to form initially, or is subsequently remelted. This brings turbulently ascending magma into continual contact with surrounding country rock and so the most fusible wall rock components can be melted and incorporated into the magma. Such a process will lead to the most primitive (i.e., hottest magmas) becoming the most contaminated.

Three factors are vital to explain the unusual anticorrelation between contamination and magma evolution. (1) The magma reservoir must fractionate without assimilating crustal material and gradually inject dykes over time causing the observed evolution of the different basalts comprising the Minto dyes. (2) More primitive, early dykes, are hotter and form wider dykes while propagating through the crust. (3) Low viscosity magmas ascent and propagate turbulently in wide enough dykes, causing constant abrasion of an initial chilled margin and resulting in thermal erosion of the country rock.

The basaltic magmas of the Minto dykes have a low enough viscosity, so that high flow rates along wide dykes would create turbulence during motion (Huppert and Sparks, 1985). With widths between 13 and 60 m, most of the Minto dykes greatly exceed the width requirements for turbulent flow of tholeiitic basalts given by Huppert and Sparks (1985). Although thinner dykes can be simply offshoots of thicker dykes, the thinner Minto dykes (samples 7904 and 8003; Ernst and Buchan, 2010) would have ascended under laminar flow conditions and show the

least contaminated Nd-isotopic signature ($\epsilon_{\text{Nd}2.0 \text{ Ga}} = -3.73$) in support of the proposed model.

Primary paleomagnetic directions have been obtained from the Minto and Lac Shpogan dykes (Buchan et al., 1998). Their paleopole is similar to pre-fold paleopoles reported from the Eskimo volcanic rocks of the Belcher Islands and the Nastapoka basalts of Richmond Gulf (Schmidt, 1980; Schwarz and Fujiwara, 1981). The Persillon volcanic rocks of Richmond Gulf and the Inukjuak dykes have also been correlated on the basis of geochemistry with the Eskimo Formation (Legault et al., 1994), suggesting that they may also belong to the ca. 1998 Ma Minto-Povungnituk event. Given the limited data available and the absence of available ages, a correlation with the Persillon volcanic rocks and the Inukjuak dykes remains tentative.

The Eskimo volcanic rocks have geochemical signatures that are distinct from those of the MPG. Previous work on the Eskimo Formation by Chauvel et al. (1987) and Legault et al. (1994) provided isotopic compositions and detailed major and trace element chemistry, respectively. Both studies concluded that Eskimo Formation basalts are evolved and unlikely to represent direct partial melts of the upper mantle. Both studies argued that assimilation-fractional crystallization (AFC) is the controlling process during the formation of the Eskimo Formation due to the negative correlation of Nb/Zr with Zr.

Reanalyzed major and trace element data on the Eskimo Formation have been published by Ernst and Buchan (2010). The data differ drastically in trace element values from the results published by Legault et al. (1994). The same

element ratio diagrams as previously mentioned (Nb/Zr vs. Zr) now show the exact opposite and show a positive correlation between Nb/Zr and Zr (Fig. 4.11b). Data from Ernst and Buchan (2010) were determined by fused disk wavelength dispersive X-ray Fluorescence (XRF) for major elements and by Inductively Coupled Plasma – Emission Spectrometry (ICP-ES) as well as Inductively Coupled Plasma – Mass Spectrometry for trace element concentrations. In contrast, the trace element data in Legault et al. (1994) were obtained by XRF and INAA. Niobium concentrations below 10 ppm for the Eskimo volcanic rocks are below the detection limit of XRF, casting doubt on the Nb/Zr ratios of Legault et al. (1994).

The precision of the Ernst and Buchan (2010) data was supported by 26 duplicate measurements and 18 measurements of an internal CANMET standard (TDB-1) that were interspersed in the sample stream, and all samples were analyzed for the same elements at the same laboratories, resulting in an internally consistent dataset. For the present manuscript the Ernst and Buchan (2010) data on the Eskimo volcanic rocks are used for comparison with the Minto dykes. Using these data, indicators of crustal assimilation are now correlating with more primitive samples and fall on linear trends with the Minto dykes. To avoid the effects of element mobility we used a ratio of Th/Nb to show the higher amount of crustal assimilation (high Th/Nb) in the more primitive rocks (Fig 4.11c). $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ values for the Eskimo Formation range from -5.7 to -7.0 (Chauvel et al., 1987). Unfortunately, published data are only available with Nd and Sm concentrations and prohibit a detailed geochemical comparison with the Minto

dykes to determine their relationship. Nevertheless, a plot of $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ vs. $1/\text{Nd}$ shows a negative correlation in agreement with the proposed AEC processes (Fig. 4.11d). The chemistry of the two widest Minto dykes (60 m and 17 m) directly overlap with the Eskimo volcanic rocks (Fig. 4.11b/c) and suggest that these might represent feeders for the Eskimo volcanic suite.

Assessing the extent of crustal contamination proves complicated. Huppert and Sparks (1985) showed that turbulently-ascending tholeiitic dykes can only assimilate around 5 % crustal rocks. Using average Nd-isotopic values for Superior craton basement (Chauvel et al., 1987) as a contaminant does not yield the observed geochemical signatures. Either contaminants with more depleted Nd-isotopic signatures were involved or higher Nd concentrations are necessary to explain the observed data. Neodymium concentrations exceeding 100 ppm have been documented on several analyses of felsic basement in the Superior craton (Énergie et Ressources Naturelles Québec, 2018). These high Nd concentrations would be sufficient to contaminate Minto dykes with Archean lower crust, with $\epsilon_{\text{Nd}2.0 \text{ Ga}}$ of -12 (Chauvel et al., 1987), to produce Eskimo basalt compositions with <5 % contamination.

4.5.3 Bilateral asymmetry of Minto-Povungnituk LIP: possible models

An interesting feature of the Minto-Povungnituk LIP recognized in the present research is the distinct “northern” and “southern” domains. The northern domain consists of the MPG and the Watts Group ophiolite. The southern domain consists of the Minto dykes and the Eskimo basalts. The Minto dykes therefore do not seem to be feeders for the MPG, but instead seem to be feeders only for the Eskimo volcanic suite.

A continental rift zone located beneath northern Hudson Bay (present location) between 2040 Ma and 1998 Ma associated with uplift in the Superior craton has been proposed by Halls and Davis (2004), parallel to the orientation of the Minto dykes (Fig. 4.7). One possible model is that this rifting event could have been induced by the arrival of a mantle plume (cf. Midcontinent rift volcanism (Nicholson and Shirey, 1990; Hollings et al., 2007)). Plume-related magmatism in the northern domain of the LIP is supported by the OIB component of the MPG and Watts Group. A potential plume centre could be associated with the uplift determined by Halls and Davis (2004) and would be located beneath Hudson Bay (Fig. 4.7).

It can be inferred that the Minto dykes trend away from this plume centre and were potentially fed from a source in the southern domain (Fig. 4.7), while the MPG (and Watts Group) were linked to sources in the northern domain. Figure 4.12 shows hypothetical scenarios that could explain how the different magmatic

domains of the Minto-Povungnituk LIP were fed. Two possible models for generating this bilateral compositional asymmetry are considered below.

Figure 4.12. Schematic sections showing the two possible models of the 1998 Ma Minto-Povungnituk LIP. (top) Bilateral plume transports two distinct deep mantle sources within the plume conduit that stay separated during ascent and impinge the lithosphere, causing the emplacement of geochemically different units on the opposite sides of the plume. (bottom) Rise of a plume along the lithospheric root of the Superior craton induced melting of the subcontinental lithospheric mantle (SCLM), which gave rise to geochemically distinct melts, opposed to the melts created by plume head melting and mixing with melts from the ambient asthenospheric mantle. For further details see text (modified after Maurice et al., 2009).

4.5.3.1 *Lithospheric effect.*

Different lithospheric domains have been hypothesised by Maurice et al. (2009) to underlie the northern vs. southern intrusions of the plume centre region. The “northern” domain would be underlain by thinned Superior craton lithosphere. The “southern” domain would be underlain by older and thicker lithosphere. In an extensive comparative study, Maurice et al. (2009) showed that a change in the chemistry of dyke swarms occurred after 2030 Ma in the northeastern part of the Superior craton. They concluded that part of the lithospheric mantle delaminated from the northeastern Superior craton at 2030 Ma, and upwelling asthenospheric mantle underwent partial melting to produce the different geochemical signatures observable in dykes emplaced after 2030 Ma in that region. This could explain the geochemical and isotopic differences observed between the northern and the southern domains of the Minto-Povungnituk event, with the northern domain lacking signatures of the Superior craton lithosphere, whereas the southern domain still ascended through thick lithosphere creating the observed enriched signatures. In their study, Maurice et al. (2009) implied that all dyke swarms ascended vertically through the crust and were created by melting of the underlying lithospheric mantle. However, lateral dyke injection from the plume centre region into the craton may be more common, especially in areas with thick lithospheric roots. Dykes can only propagate upwards under conditions of positive buoyancy and will tend to propagate laterally once a level of neutral buoyancy is approached (Lister and Kerr, 1990; Lister, 1991; Wilson and Head, 2002). Lateral injection of dyke

swarms away from a mantle plume centre has been shown for the Mackenzie and other swarms (Baragar et al. 1996; Ernst, 2014; Sandeman et al., 2014; Mackinder et al., 2019).

The lithospheric thickness of the Superior craton is believed to exceed 140 km south of the Cape Smith Belt and to be up to 170 km at the location of the Minto dykes (Fig. 2 in Maurice et al., 2009). The magmatic inactivity of the Ungava Peninsula after the emplacement of the Paleoproterozoic dyke swarms and Circum-Ungava belts implies that the present structure of its underlying mantle has not been affected by younger events (Maurice et al., 2009). This coincides with observations that the northeastern Superior Province represents an Archean terrain with a thick cratonic keel older than 2.6 Ga (Boily et al., 2009; Bédard et al., 2013; Harris and Bédard, 2014), a notion that is further supported by the inferred occurrence of Archean diamondiferous kimberlites in the western Superior craton (Kopylova et al., 2011). Basaltic melt generation at depths >100 km is unlikely and even if melt was produced it would reflect melting in the garnet peridotite stability field (e.g., higher Gd/Yb ratios resulting from residual garnet) (White and McKenzie, 1995). Therefore, it is unlikely that melting occurred beneath the thick root of the craton. Instead melting likely occurred closer to the plume centre on the side of the craton, and that dykes propagate laterally into the craton from this plume centre. Interaction with Superior craton lithosphere would then occur at the cratonic margin beneath Hudson Bay, near the inferred plume centre along the projection of the Minto dyke trend.

A possible explanation for the melt generation, still involving a lithospheric signature, would include the plume rising up beneath the Superior craton and gradually spreading along the lithospheric keel (e.g., Sleep, 1996). On its ascent, the plume can then gradually melt parts of the lithosphere and flow towards a shallow magma chambers surrounding the plume centre. From there a sub-swarm of dykes could intrude into the Superior craton creating the Minto dykes (Fig. 4.12, bottom panel). The injection of individual sub-swarms surrounding plume centres has been suggested by Baragar et al. (1996) and Blanchard et al. (2017).

4.5.3.2 Distinct or zoned plumes

Another explanation for the difference in compositions between the northern domain (MPG and Watts Group) and southern domain (Minto dykes and Eskimo basalts) is, they originate from two different neighboring mantle plumes. Numerical modelling for determining coeval plume spacing is strongly affected by boundary layer heterogeneities and has not been successful (Sleep, 1992; Wüllner and Davies, 1999). However, several examples throughout Earth's history show that roughly coeval neighboring plumes exist. These include the 2.48-2.45 Ga Matachewan and 2.51 Ga Mistassini plume centers along the eastern margin of the Superior craton, which are 800 km apart and the 2.23 Ga Malley and 2.21 Ga MacKay plume centers along the eastern margin of the Slave craton, which are >500 km (e.g., Ernst and Bleeker, 2010; Buchan et al. 2012).

Based on these examples, coeval plume centers can be as closely spaced as 500-800 km.

Alternatively, we consider the possibility of two distinct types of deep mantle material rising in a single plume representing chemical zonation of the plume source. The deep mantle currently hosts one minor and two major regions of low shear wave velocity (LLSVP) that are inferred to represent slab graveyards. The edges of the LLSVPs have been associated with the formation of mantle plumes for the past 300 myr (Torsvik et al., 2010, 2014). Weis et al. (2011) and Harrison et al. (2017) proposed the Hawaiian plume rising along the edge of the LLVSP and mechanically removing fragments of the LLSVP and carrying them upward. They attribute the Kea compositional component to the plume and the Loa compositional component to the LLSVP (Harrison et al., 2017; Harrison and Weis, 2018). These components ascended on opposite sides of the plume and reflect the structure of the underlying deep mantle. This process has been shown to be possible, if thermal heterogeneity outweighs chemical heterogeneity in the source region (Jones et al., 2016). Utilizing the Hawaiian model, we consider the idea that the northern vs. southern compositional domains of the Povungnituk-Minto LIP are related to ascent of different mantle material on opposite sides of the plume. Upon arrival at the base of the lithosphere the plume would spread out in opposite directions from the plume centre (Fig. 4.12, top panel). Partial melting of these distinct mantle sources would lead to mafic compositions with the distinct features of the Povungnituk-Watts (northern) vs. Minto-Eskimo (southern) domains.

If we consider the northern vs. southern domain chemistry in the Minto-Povungnituk LIP to have originated in the deep mantle, the Povungnituk-Watts component (northern domain) would be linked to ambient deep mantle, whereas the Minto-Eskimo (southern domain) composition has isotopic characteristics that represent an enriched source (e.g., slab graveyard). It would be difficult to achieve upwelling of a plume from a slab graveyard causing complications for the two-plume model, but can be satisfied in a single zoned plume model.

The potential implications of the Minto-Povungnituk LIP being related to ambient lower mantle and LLSVP are intriguing. A key issue is the potential for sufficiently enriched ϵ_{Nd} compositions ($\epsilon_{\text{Nd}2.0 \text{ Ga}} = -4$) in a slab graveyard at 2 Ga, since modern ocean island basalts rarely have ϵ_{Nd} values that are so negative. Differentiated material needs to be subducted early in Earth's history to give sufficient time to develop the observed isotopic signatures.

In summary, various assumptions need to be made that cannot be further tested with the currently available dataset, namely that the Minto-Povungnituk LIP represents a) a single plume, b) is sourced from different lower mantle sources and has not been influenced more by upper mantle or lithospheric processes and c) is ultimately derived from the LLSVP region.

4.5.3.3 Constraining the existence and location of LLSVPs in the Precambrian

Two models have been proposed for the origin of the different geochemistries between the northern and the southern domains of the Minto-Povungnituk LIP. Melting of Superior craton continental lithosphere created magmas (southern domain) that are geochemically different from the melts produced by mixing of the melting plume head with adjacent depleted mantle (northern domain). Alternatively, the bilateral chemistry of the Minto-Povungnituk LIP can be explained by melting of different plume head material that had been sourced from different deep mantle domains and has been brought up on opposite sides of a plume conduit.

One implication of the single bilateral plume model for the 1998 Ma Povungnituk-Minto LIP is the southern domain would imply a LLSVP existed at this time. Based on published paleomagnetic data on the Minto dykes (Buchan et al., 1998), the inferred paleo-latitude and orientation of the boundary between LLSVP (southern domain) and ambient mantle (northern domain) at this time are shown in Figure 4.13. Note both northern and southern hemisphere options are shown with uncertainty as to which paleomagnetic polarity option applies. Accordingly, this suggests a new technique for locating LLSVPs in the Paleozoic and Precambrian:

- 1) Search for LIP events that have a bilateral chemistry and characterize their chemistries, and assess which would be comparable to a LLSVP.
- 2) Use paleomagnetism to determine the paleo-latitude and paleo-orientation of the edge of the LLSVP implied by the LIP bilateral chemistry.
- 3) Assuming a number of LIPs exhibit such bilateral chemistry, then build up a dataset on the location of LLSVP boundaries and the evolved LLSVP compositions for a variety of plume centre locations, and through time.

Such a dataset could test whether these Paleozoic and Precambrian LLSVPs are consistent with the current pattern of two major LLSVPs (Torsvik et al., 2010, 2014) allowing the testing of models in which new LLSVPs are created during each supercontinent cycle (Li and Zhong, 2009) or whether the two main LLSVPs have been present for at least 300 Myr and potentially through much of Earth history (Torsvik et al., 2010, 2014). Supporting persistent ancient mantle heterogeneities (proto-LLSVPs?), Rizo et al. (2016) argued for Hadean reservoirs displaying W isotopic anomalies as a source for modern LIPs (the 55 Ma North Atlantic Igneous Province and the 122 Ma Ontong Java Plateau).

Figure 4.13. Globe showing the paleo-latitude of the Superior craton at the time of Minto dyke emplacement (ca. 1998 Ma) referenced to present day geography (after Buchan et al., 2007). Minto dykes location is shown as squares. Open squares represent the paleomagnetic reconstruction and the closed square represents its current day location. Coloured fields indicate the deep mantle structures inferred to explain the bilateral symmetry of the Minto-Povungnituk LIP proposed in Figure 4.12 (top). For further details see discussion in text.

4.6 Conclusions

A new U-Pb baddeleyite maximum age of 1998 ± 6 Ma has been obtained for a dolerite sill belonging to the mafic Povungnituk Group (MPG) of the Cape Smith Belt and matched to the flood basalts on the basis of geochemistry. This represents the first direct U-Pb age on mafic magmatism of the Povungnituk Group (Beauparlant Fm.). The previously correlated 2038 Ma Korak sills likely belong to a separate LIP event.

The similarity in age among Povungnituk, Watts and Minto units suggests a widespread event in the northern Superior craton covering over 400,000 km². The orientation of the Minto dykes would suggest a magmatic center (plume center) in Hudson Bay situated along the extrapolated trend of the Minto swarm which we can intersect with a westward extension of the interpreted rift-related Povungnituk flood basalts.

Distinct geochemical differences can be observed between the northern and southern domains of the Minto-Povungnituk LIP. The northern domain, consisting of the MPG and the Watts Group, is characterized by uncontaminated rocks created by mixing of depleted mantle and an enriched plume component ($\epsilon_{\text{Nd}2.0 \text{ Ga}} = +4$ and $+1$, respectively). The southern domain, comprising the Minto dykes and the Eskimo Formation basalts, was fed by an even more enriched source ($\epsilon_{\text{Nd}2.0 \text{ Ga}} = -4$) with the basaltic magmas undergoing additional crustal contamination.

Two different processes can cause this geochemical difference. Plume-induced melting of subcontinental lithospheric mantle beneath the Superior craton could have formed magmas that differ sufficiently in composition and isotopic signature from melts generated by mixing of a plume head and depleted ambient mantle. Another possible cause could be the direct transport of different primary compositions in two adjacent plumes, or within a single plume, from deep mantle sources. In the latter interpretation the different plume compositions would remain on separate sides of the plume during ascent and would imply that the two spatially separated compositions in the Minto-Povungnituk LIP directly represent the composition of the deep mantle on different sides of the plume at the time of emplacement. The alternate interpretation of two nearby but separate coeval plumes would also imply two distinct compositions of the deep mantle. Given this example, bilateral asymmetry in LIP events could then be used as a tool to map and locate deep mantle anomalies (like the modern LLSVP) back through Earth's history.

4.7 References

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**5 GEOCHEMISTRY OF THE MAFIC LAVAS OF THE
ROBERTS LAKE SYNCLINE (NE SUPERIOR
CRATON): COMPARISON WITH
PALEOPROTEROZOIC VOLCANIC SEQUENCES OF
THE CAPE SMITH BELT AND THE LABRADOR
TROUGH**

5.1 Introduction

Massive volumes of basaltic lava flows erupted between ca. 2.2 and 1.8 Ga onto the rifted margins of the Superior craton. This volcanism represents the extrusive component of several large igneous provinces (LIPs) and in some cases the flows are linked to sills and dykes in the interior of the craton (Ernst and Bleeker, 2010; Minifie et al., 2013; Hamilton et al., 2017; Kastek et al., 2018; Section 4.5.1). Earlier LIPs are regarded as manifestations of the predominant mechanism driving the breakup of Superia, a late Archean ancestral landmass from which the modern Superior Province was derived. However, despite early suggestions that they might have recorded the transition from rifting to passive margin formation (e.g., Hynes and Francis, 1982), the LIPs erupted along the northern and eastern margins of the Superior Province (ca. 2.0 and 1.8 Ga) clearly postdate Superia rifting by a considerable span of time (Mungall, 2007). While the regional distribution of these volcanic packages has been mapped by a number of studies (Dimroth et al., 1970; Baragar and Scoates, 1981, Clark and Wares, 2006, St. Onge et al., 2006), their association to specific events remains poorly constrained or unknown altogether. This is especially true for the volcanic units located within the Roberts Lake Syncline located on the northeastern margin of the Superior craton between the Cape Smith belt and Labrador Trough (New Quebec orogen). Geochemical studies of mafic units in the Labrador Trough have concentrated on its central part (Machado et al., 1997; Findlay et al., 1995; Rohon et al., 1993; Skulski et al., 1993), whereas analyses from the northern part are scarce (Ciborowski et al., 2016) and have not included the

Roberts Lake Syncline. Based on regional geology, the basalt units of the Roberts Lake Syncline have been previously mapped as part of the ca. 1880 Ma Circum-Superior LIP unit of the New Quebec orogen (e.g., Hardy, 1976; Clark and Wares, 2006), but these correlations have not been confirmed by geochronology or geochemistry. In the past, the entire Roberts Lake Syncline with its basal sedimentary units and upper mafic flows, intruded by ca. 1880 Ma Circum-Superior LIP aged sills, was suggested to represent a continuation of the Cape Smith stratigraphy, rather than the cyclic Labrador Trough stratigraphy farther south (e.g., Ferron et al., 2000; Madore and Larbi, 2001; Bleeker and Kamo, 2018; citing work by Rohon et al., 1993; Findlay et al., 1995; Clark and Wares, 2006; Henrique-Pinto et al., 2017).

Detailed sampling, including a traverse along the western limb of the Roberts Lake Syncline, enables us to present a geochemical analysis of the units and identify unique signatures that might be linked to the Cape Smith units (with 1998 Ma Minto-Povungnituk and ca. 1880 Ma Chukotat Group of the Circum-Superior LIP) or linked to the Labrador Trough units (with ca. 2170 Ma Biscotasing and ca. 1880 Ma Cycle 2 volcanics of the Circum-Superior LIP).

5.2 Methodology

5.2.1 Sample collection

During field seasons in 1998, 1999 and 2000, James Mungall collected 49 samples from lavas on a traverse along the western limb in the central part of the Roberts Lake Syncline, 11 samples from the basaltic successions above and below the Kyak Bay intrusion, and 24 samples of peridotites, gabbros, and massive sulfides from the Qarqasiaq sill.

An additional 10 samples were collected during field work in 2013 by Nico Kastek and Wouter Bleeker. Three of these sample are from the southern part of the syncline near the Arnaud River and the remaining 7 samples from a drill core located at the northern extremity of the Qarqasiaq complex, where peridotites of the sill, as well as underlying basalts, have been sampled. The sample information is summarized for representative samples in Tables 5.1 and 5.2 and the distribution of samples is shown in Figure 5.1. The full dataset is available in Appendixes 7-9.

Figure 5.1. Geological map of the Roberts Lake Syncline. Modified from SNRC 25D, 25C, 24M, 24N. Sample locations include the field season from 2013, locations from the collection of J.E. Mungall and samples from Bunting (2000).

Table 5.1. Representative major element geochemical data from the Roberts Lake Syncline.

Sample No.		JMPAR99-54	JMPAR99-77	NQ0-13-03	JMPAR99-15	JMPAR99-17	NQ0-13-08
description		basalt	basalt	basalt	basalt	basalt	peridotite
Group		Group 1	Group 1	Group 1	Group 2	Group 2	Qarqasiaq
UTMX		430874	426672	444318	434418	434418	428647
UTMY		6672197	6671985	6653397	6672550	6672550	6670242
SiO2	%	48.74	46.33	49.40	47.90	46.84	37.40
TiO2	%	1.16	1.48	2.20	0.75	0.63	0.18
Al2O3	%	14.79	14.59	12.25	11.37	8.68	3.15
Fe2O3	%	12.61	14.53	17.40	12.00	11.94	12.30
MnO	%	0.21	0.22	0.25	0.17	0.18	0.18
MgO	%	7.50	7.26	5.68	13.32	17.74	33.40
CaO	%	10.42	12.33	9.58	8.82	9.75	1.71
Na2O	%	2.92	1.77	0.97	2.38	1.07	<0.01
K2O	%	0.19	0.11	0.08	0.03	0.02	0.01
P2O5	%	0.08	0.10	0.14	0.05	0.04	0.01
LOI	%	2.23	2.04	2.88	3.32	3.81	10.00
Total		100.85	100.74	100.85	100.11	100.71	98.86
Mg#		34	30	22	49	56	70

Coordinates in NAD83, zone 19V
Full dataset in Appendix 7-9.

Table 5.2. Representative trace element geochemical data from the Roberts Lake Syncline

Sample No.		JMPAR99-54	JMPAR99-77	NQ0-13-03	JMPAR99-15	JMPAR99-17	NQ0-13-08
description		basalt	basalt	basalt	basalt	basalt	peridotite
Group		Group 1	Group 1	Group 1	Group 2	Group 2	Qarqasiaq
UTMX		430874	426672	444318	434418	434418	428647
UTMY		6672197	6671985	6653397	6672550	6672550	6670242
Ba	ppm	47	42.2	24.8	12.8	6.6	2.3
Ce	ppm	10.3	13.3	19.5	5.1	4.9	2.0
Cr	ppm	370	170	20	1500	1810	3850
Cs	ppm	0.04	0.02	0.19	0.1	0.14	0.09
Dy	ppm	3.55	3.81	5.46	2.77	2.26	0.6
Er	ppm	2.2	2.41	3.05	1.63	1.23	0.37
Eu	ppm	0.76	1.11	1.57	0.58	0.47	0.16
Ga	ppm	18.1	21.8	20.1	11.9	8.8	3.8
Gd	ppm	3.18	3.71	5.02	2.35	1.98	0.45
Hf	ppm	1.7	2.0	3.0	1.1	0.9	0.3
Ho	ppm	0.72	0.79	1.14	0.55	0.47	0.14
La	ppm	3.8	5.0	7.5	1.9	2.1	0.6
Lu	ppm	0.3	0.31	0.45	0.25	0.18	0.1
Nb	ppm	4.0	4.7	7.1	1.8	1.4	0.5
Nd	ppm	8.2	10.4	13.7	4.8	4.2	1.0
Pr	ppm	1.62	1.99	2.83	0.85	0.77	0.22
Rb	ppm	6.0	1.5	1.5	0.5	0.3	1.3
Sm	ppm	2.71	3.07	4.07	1.7	1.44	0.4
Sn	ppm	1.0	2.0	1.0	1.0	<1	<1
Sr	ppm	105.0	200.0	152.5	105.0	28.7	1.9
Ta	ppm	0.2	0.2	0.4	0.2	0.2	<0.1
Tb	ppm	0.55	0.62	0.88	0.46	0.34	0.08
Th	ppm	0.27	0.41	0.57	0.21	0.15	0.07
Tm	ppm	0.31	0.33	0.45	0.27	0.19	0.06
U	ppm	0	0	0.2	0	<0.05	0.09
V	ppm	328	373	613	292	229	93
W	ppm	1.0	1.0	<1	<1	<1	<1
Y	ppm	18.2	20.6	27.4	16	12.9	3.4
Yb	ppm	2.0	2.0	3.0	2.0	1.0	0
Zr	ppm	59	75	105	42	34	8.0
Ag	ppm	<0.5	2.5	<0.5	<0.5	<0.5	<0.5
As	ppm	<5	<5	<5	<5	<5	<5
Cd	ppm	<0.5	<0.5	1.1	0.9	0.7	<0.5
Co	ppm	48	49	61	65	66	136
Cu	ppm	89	662	71	138	68	89
Li	ppm	10	10	<10	20	20	<10
Mo	ppm	<1	<1	<1	<1	<1	<1
Ni	ppm	100	94	43	451	561	1720
Pb	ppm	6.0	40	4.0	<2	5.0	<2
Sc	ppm	44	43	48	35	29	7.0
Tl	ppm	<10	<10	10	<10	<10	<10
Zn	ppm	91	112	155	75	68	83
Au	ppb	14	87		<1	<1	
Pt	ppb	2.1	2.6		9.4	8.3	
Pd	ppb	9.3	22		9.7	8.7	

Coordinates in NAD83, zone 19V
Full dataset in Appendix 7-9

5.2.2 Major and Trace elements

Major element concentrations in samples from the Roberts Lake Syncline in 1998, 1999 and 2000 were determined by borate fusion X-ray fluorescence analysis at the Geoscience Laboratories (Geo Labs) of the Ontario Geological Survey in Sudbury, Ontario. Powders were calcined to measure loss on ignition (LOI) followed by complete fusion using a lithium borate flux. Analytical uncertainties for Geo Labs have been reported in Mungall and Martin (1994). Samples collected in 2013 and a subset of the 1999 sample suite were sent to ALS Geochemistry laboratories in North Vancouver, British Columbia. Samples from the combined sample suite which had not already been analyzed for major elements, were analyzed via Inductively Coupled Plasma Atomic Emission Spectrometry (ICP-AES).

Thirty six sample powders from the combined suite, including an internal standard (10-LT-05, a basaltic andesite from Lake Tahoe, California) were analyzed at ALS Geochemistry laboratories for trace and rare-earth elements (REE) via Inductively Coupled Plasma Mass Spectrometry (ICP-MS) and 26 were analyzed for Pt, Pd and Au by fire assay and ICP-MS finish. Trace-element analytical uncertainties (1σ) for ALS Geochemistry laboratories, based on repeat analysis of the Lake Tahoe internal standard, are less than 1 ppm with the exception of Ba (15.8 ppm), Ce (2.0 ppm), Cr (12 pm), Sr (16.0 ppm), V (8 ppm), Zr (5 ppm), Co (1.6 ppm), Cu (2.71 ppm), Ni (4.8 ppm), Pb (1.6 ppm) and Zn (3 ppm). Major element uncertainties are <0.1 wt. %. Since no internal standard

was included for the Pd, Pt and Au data, the analytical uncertainties are less constrained.

5.3 Results

5.3.1 Petrography

The lavas from the Roberts Lake Syncline analysed for this study are all metamorphic rocks in which the primary magmatic minerals (olivine, pyroxene, plagioclase, and accessory minerals) have been replaced by a greenschist facies assemblage. Minor calcite veining is observed, but the rocks generally appear pristine apart from the pervasive hydrous alteration. Primary textures are still clearly preserved in most cases despite the imposition of a mild mineral foliation. Olivine phenocrysts, typically about 0.5 to 1 mm in diameter, have been replaced by serpentine, whereas pyroxenes have largely been replaced by tremolite-actinolite. Primary plagioclase crystals are largely replaced by albite-oligoclase and epidote mixtures; the growth of abundant chlorite throughout the matrix was presumably fed by the destruction of both pyroxene and plagioclase. Primary Fe-Ti oxide minerals have been replaced by titanite or rutile.

5.3.2 Major elements

All major element concentrations have been recalculated on an anhydrous basis. The complete series of basalt samples displays a wide range with MgO contents ranging from 3.63 to 18.31 wt. %. $\text{Fe}_2\text{O}_3^{\text{T}}$ (total iron expressed as Fe_2O_3) varies from 11.04 to 17.91 wt. % (Fig. 5.2a). Mg# (molar $\text{Mg}/(\text{Mg}+\text{Fe}^{\text{T}}) \times 100$) varies between 18 and 56. Al_2O_3 varies between 8.96 and 15.49 (Fig. 5.2b). TiO_2 varies from 0.65 to 3.12 (Fig. 5.2c). P_2O_5 varies between 0.04 and 0.48 (Fig. 5.2d). The silica content of all samples ranges from 44.29 to 54.04 wt. %. CaO varies from 4.58 to 8.31. The total alkali content ($\text{Na}_2\text{O} + \text{K}_2\text{O}$) varies from 0.71 to 6.18 wt. %. SiO_2 , CaO and the alkali elements show scatter against MgO, which is attributed to rock-fluid interactions during secondary alteration processes such as weathering and metamorphism.

On the basis of stratigraphy and chemical composition, the rocks can be subdivided into two groups. On the sample traverse they are divided by a major layered sill complex east of Lac Chaunet, which marks a large-scale fault structure. Current mapping does therefore not indicate that the upper unit lies conformably on the other. Group 1 exhibits MgO from 3.63 to 8.65 wt. %, $\text{Fe}_2\text{O}_3^{\text{T}}$ from 11.04 to 17.91 wt. %, TiO_2 from 0.83 to 3.12 wt. %, P_2O_5 concentrations between 0.06 to 0.48 wt. % and Al_2O_3 from 10.88 to 17.41 wt. %. $\text{Fe}_2\text{O}_3^{\text{T}}$, TiO_2 and P_2O_5 concentrations increase with decreasing MgO and Al_2O_3 concentrations decrease with decreasing MgO.

Group 2 has the highest MgO concentrations (12.99 to 18.31 wt. %). If, as their textures suggest, these compositions represent liquids, then they span the

range from komatiite to picritic high-MgO basalt. $\text{Fe}_2\text{O}_3^{\text{T}}$ (11.98 to 12.88 wt. %), TiO_2 (0.65 to 0.77 wt. %) and P_2O_5 (0.04 to 0.05 wt. %) show no significant variation with decreasing MgO concentration. Al_2O_3 concentrations (8.96 to 11.75 wt. %) increase with decreasing MgO. All samples of Group 2 are located in the center of the syncline and therefore lie stratigraphically at the top of the succession.

In addition to the 5 peridotites of the Qarqasiaq complex, we have compiled analyses from peridotite samples from the bottom and gabbroic portions from the top of the complex from Mungall (1998) and Bunting (2000) that include limited major and trace element data as well as some PGE analyses. The peridotites have MgO contents from 23.99 to 34.22 wt. %, $\text{Fe}_2\text{O}_3^{\text{T}}$ from 11.80 to 15.00 wt. %, TiO_2 from 0.17 to 0.45 wt. %, P_2O_5 concentrations from 0.02 to 0.03 wt. % and Al_2O_3 from 3.07 to 7.93 wt. %. A positive correlation between Al_2O_3 , TiO_2 and P_2O_5 with MgO can be observed, while Fe_2O_3 concentrations do not follow a clear trend. The gabbros have MgO contents from 6.90 to 13.03 wt. %, $\text{Fe}_2\text{O}_3^{\text{T}}$ from 9.03 to 11.72 wt. %, TiO_2 from 0.63 to 0.74 wt. %, P_2O_5 concentrations from 0.05 to 0.06 wt. % and Al_2O_3 from 12.43 to 13.63 wt. %. $\text{Fe}_2\text{O}_3^{\text{T}}$, TiO_2 and P_2O_5 concentrations increase with decreasing MgO and Al_2O_3 concentrations decrease with decreasing MgO. One sample, mapped in the field as peridotite, forms an outlier with concentrations of MgO, $\text{Fe}_2\text{O}_3^{\text{T}}$, TiO_2 , P_2O_5 and Al_2O_3 of 15.13, 12.39, 0.61, 0.04, and 11.83, respectively. In the following descriptions, this outlier is omitted.

Figure 5.2. Major element composition of lavas of the Roberts Lake Syncline and Qarqasiaq complex. Small symbols compiled from Ernst and Buchan (2010); Barnes et al. (2015); Ciborowski et al. (2016); this thesis.

5.3.3 Trace elements

A representative selection of bivariate diagrams of both compatible and incompatible trace elements against MgO for Groups 1 and 2, as well as the samples from the Qarqasiaq complex, is shown in Figure 5.3. Also shown for comparison are representative compositions of basaltic lavas both from the 2170 Ma Biscotasing LIP, the 1998 Ma Beuparlant Formation (Cape Smith belt) and from the younger ca. 1880 Ma Circum-Superior suites (Thompson Nickel Belt, Chukotat Group of the Cape Smith belt, and Hellancourt basalts of the Labrador Trough as compiled by Ernst and Buchan, 2010; Barnes et al., 2015; Ciborowski et al., 2016) as well as the ultramafic feeders of the Circum-Superior suites (Barnes et al., 2015). For the lavas, the compatible elements Ni, Cr and Co all show tight positive correlations with MgO. Group 1 shows intermediate concentrations, ranging between 50 to 121 ppm for Ni, 80 to 370 ppm for Cr, and 45 to 54 ppm for Co. Group 2 shows the highest concentrations of Ni, Cr and Co with 490 to 712 ppm, 1340 to 1450, and 63 to 64 ppm, respectively. On a Ni vs. MgO diagram (Fig. 5.3a), all lavas plot mostly along a trend of within plate basalt compositions (Keays and Lightfoot, 2007). While some of the Group 1 samples fall slightly below this trend, most of the Group 2 samples fall slightly above it. The peridotites of the Qarqasiaq complex have high Ni concentrations from 897 to 2320 ppm and Cr concentrations 2614 to 3006 ppm, while the gabbros have concentrations from 124 to 347 ppm Ni and 394 to 1603 Cr. Ni concentrations are positively correlated with MgO. Copper concentrations range from 3 to 662 ppm (majority lie below 150 ppm) in the lavas, 74 to 1540 ppm in the peridotites

and 124 to 347 ppm in the gabbros and do not correlate with decreasing MgO (Fig. 5.3b). A weak correlation between decreasing MgO and decreasing Pt (0.7-2.1 ppb) and Pd (0.7-9.3 ppb) can be seen (Fig. 5.3c/d). The incompatible elements La, Sm, Yb and Nb all increase in concentration as MgO decreases (Fig. 5.3e/f).

Figure 5.3. Selected trace element composition for lavas of the Roberts Lake Syncline and Qarqasiaq complex. The solid line is an array of within plate basalts taken from Keays and Lightfoot (2007). Small symbols compiled from Ernst and Buchan (2010); Barnes et al. (2015); Ciborowski et al. (2016); this thesis.

All lavas are classified as basalts on a Nb/Y vs. Zr/Ti diagram (Fig. 5.4a; Pearce, 1996). On a Ti/V diagram Group 1 forms a range between 20 and 50, typical for continental flood basalts, whereas Group 2 is distinct with Ti/V ratios below 20 (Fig. 5.4b; Shervais, 1982). The peridotites and gabbros plot below Ti/V ratios of 20, where peridotites show the lowest concentrations and the gabbros overlap with the Group 2 lavas.

On a diagram of Th/Nb vs. Nb/Yb, all lavas plot inside the mantle array (Fig. 5.4c; Pearce, 2008) and in between N-MORB and E-MORB compositions. On a diagram of Nb/Y vs. Zr/Y, all samples plot inside the Icelandic array (Fig. 5.4d; Fitton et al., 1997), showing that all samples differ from modern N-MORB.

Figure 5.4. Trace element geochemical diagrams for lavas of the Roberts Lake Syncline and Qarqasiaq complex. (a) Zr/Ti vs. Nb/Y (after Pearce, 1996), to discriminate rock type. All samples fall within the basalt field (b) V vs. Ti/1000 diagram (after Shervais, 1982), where Group 1 falls to the right of a Ti/V ratio of 20 and Group 2 and the samples of the Qarqasiaq complex fall on the right of the line. (c) Th/Yb vs. Nb/Yb (after Pearce, 2008), where all samples fall within the defined mantle array between compositions of N-MORB and E-MORB. Lower crust (LC), middle crust (MC) and upper crust (UC) from Rudnick and Fountain (1995) (d) Nb/Y vs. Zr/Y with Iceland plume array (Fitton et al., 1997). All lavas from the Roberts Lake Syncline fall above the array separating E-MORB from N-MORB compositions, showing that the samples do not represent Proterozoic depleted mantle melts. The upwards trend for the Qarqasiaq complex peridotites indicates minor contamination. Small symbols compiled from Ernst and Buchan (2010); Barnes et al. (2015); Ciborowski et al. (2016); this thesis.

Group 1 has steep REE profiles with La/Yb_C ratios between 1.24 and 3.15 and displays enrichment in the LREEs relative to the MREEs (Fig. 5.5). This is reflected in La/Sm_C ratios of 0.88 to 1.34. Heavy-REEs are depleted relative to the MREEs with Gd/Yb_C ratios ranging from 1.30 to 1.87 (Fig. 5.5). Eu-anomalies ($\text{Eu/Eu}^* = \text{Eu}_N/\sqrt{\text{Sm}_N \cdot \text{Gd}_C}$) for Group 1 vary between 0.79 and 1.11. No clear correlation between elemental concentrations (e.g., MgO) and calculated Eu-anomaly can be observed. It should be noted that the two lowest Eu-anomalies are seen in the most primitive samples of Group 1. These samples have very low REE concentrations and analytical uncertainty might be a cause for the irregular reported Eu concentrations.

Group 2 exhibits relatively flat, near parallel REE profiles (La/Yb_C 0.98 to 1.02) with REE concentrations ~ 8 -10 times that of chondritic values (Fig. 5.5). Group 2 samples are slightly depleted in the LREEs relative to the middle-REEs (MREE) and possess La/Sm_C ratios of 0.85 to 0.92 (Fig. 5.5). The HREEs are also slightly depleted relative to the MREEs and this is reflected Gd/Yb_C ratios of 1.14 to 1.22 (Fig. 5.5). No sample displays a significant Eu anomaly.

The peridotites exhibit relatively flat, near parallel REE profiles (La/Yb_C 0.95 to 1.30) with REE concentrations of ~ 2 -4 times that of chondritic value (Fig. 5.5).

Group 1 has primitive-mantle-normalised multi-element profiles that are steeper than those of Group 2 (Fig. 5.5). Thorium is depleted relative to Nb in all samples. Minor Eu-anomalies can be observed in some of the samples and all samples exhibit minor Y-anomalies. The majority of the samples show patterns

similar to E-MORB with some of them showing a trend towards N-MORB compositions.

The multi-element profiles of Group 2 are near parallel and relatively flat with elemental concentrations ~2-4 times those of primitive mantle (Fig. 5.5). Both samples are slightly depleted in Th relative to the rest of the elements. Minor Ti- and Y-anomalies can be observed in both samples. Group 2 multi-element profiles have slopes that lie between N-MORB and E-MORB, but exhibit lower concentrations than either of the two endmembers.

The peridotites show slight enrichments in Th/Yb and Nb/Y ratios, attributed to minor crustal contamination.

In all plots in Figures 5.3, 5.4 and 5.5, there is a clear overlap of Group 1 compositions with those of the Beuparlant Formation of the Povungnituk Group and the samples from the Biscotasing LIP, of Group 2 compositions with those of the Circum-Superior suites, and of the peridotites with the ultramafic feeders of the Circum-Superior suites.

Figure 5.5. Chondrite normalized rare earth element diagram (top) and multielement diagram for incompatible elements (bottom) for lavas of the Roberts Lake Syncline and Qarqasiaq complex and selected rock types. Group 1 and 2 form different trends on both diagrams with Group 1 showing overall higher concentration of depicted elements as well as steeper trends on both diagrams. Chondrite values of McDonough and Sun (1995), primitive mantle, N-MORB, E-MORB, and OIB values from Sun and McDonough (1989).

5.4 Discussion

5.4.1 *Fractional crystallization and crustal assimilation*

To assess the effects of fractionation, alphaMELTS thermodynamic software has been used (Ghiorso and Sack, 1995; Asimow and Ghiorso, 1998; Smith and Asimow, 2005).

To assess the possibility that rocks of Group 1 are related to one another by fractional crystallization, samples JMPAR-99-88 and JMPAR-99-54 were chosen as two alternative parental magmas. They represent the most magnesian samples in Group 1 (8.50 and 7.50 MgO wt. %) that have an aphyric texture and therefore do not likely represent cumulates. In the alphaMELTS model it was assumed that the magma included 0.25 % H₂O and oxygen fugacity was fixed at the value of the fayalite-magnetite-quartz solid oxygen buffer (FMQ). Sample JMPAR-99-54 has a liquidus temperature of 1188 °C and crystallizes assemblages along a gabbroic cotectic, with minor quantities of olivine early in the sequence followed by only plagioclase and pyroxene. Major element concentrations during fractionation fit the observed trends of Group 1 samples in bivariate plots (Fig. 5.6a-d). The samples fall along a 60 % fractionation trend for Al₂O₃, TiO₂ and P₂O₅ and a 40 % fractionation trend for Fe₂O₃. Unfortunately, these two highly magnesian samples also exhibit the most marked Eu-anomalies (Eu/Eu* = 0.79) in multi-element profiles. The Eu depletions are either products of early feldspar fractionation or result from analytical uncertainty. Therefore, to examine the behaviour of trace elements during a similar crystallization path with better constrained initial trace element concentrations the model was repeated

using the somewhat less magnesian sample JMPAR-99-77 to represent the initial magma. The best fitting model includes 0.25 % H₂O but requires a fractionation pressure of 2 kbar. In the model this composition has a liquidus temperature of 1208 °C and is co-saturated immediately with plagioclase and clinopyroxene. Figures 5.6a-d show that the model for JMPAR-99-77 follows the major element concentrations of Group 1 for Al₂O₃, TiO₂ and P₂O₅ well. Observed Fe₂O₃ concentrations cannot be reproduced with sample JMPAR-99-77. Figure 5.6e shows the trace element results of the model overlaying the multi-element diagram for Group 1 in increments of 20 %. In combination, these diagrams show that Group 1 can be formed by the fractional crystallisation of 60 % of a mineral assemblage comprising 2 % olivine, 50 % pyroxene and 48 % plagioclase from an initial liquid with major element composition similar to that of sample JMPAR-99-54, and trace element abundances similar to those in sample JMPAR-99-77. In both cases magnetite joins the model crystallizing assemblage at MgO concentrations lower than those of any samples in the suite, indicating that none of the samples were likely to have been affected by the removal of Fe and Ti into oxide minerals.

For Group 2 the model used the most primitive (highest MgO sample) from the Roberts Lake Syncline (JMPAR-99-17: 18.31 wt. % MgO), which is technically a komatiite and shows textures suggesting it represents a liquid. Although unlikely to be a true primary magma, it represents the closest approximation available. In the alphaMELTS model it was assumed that the magma was anhydrous and oxygen fugacity was fixed at the value of the fayalite-

magnetite-quartz solid oxygen buffer (FMQ). The modelled liquidus temperature of the system is 1416 °C at 1 kbar pressure. The model magma crystallizes only olivine and minor quantities of chromite from 1416 to 1240 °C, upon which olivine is replaced by clinopyroxene in the liquidus assemblage, when the MgO concentration in the residual liquid is 10.03 wt%. Since all members of Group 2 in our suite have greater than 10.03 wt. % MgO, we conclude that all of our Group 2 lavas are picrites that crystallized only olivine. Major element ratios suggest that the two geochemical groups (1 and 2) could be related by fractional crystallisation, but comparison of modeled and measured trace element concentrations show that the two suites cannot be comagmatic. This would agree with the observation that Group 2 lies stratigraphically above Group 1 and should not represent the primary magmas of Group 1.

The model results show that all of the variation seen in Group 2 can be formed by the fractional crystallisation of less than 20 % of olivine and accessory chromite from a composition similar to sample JMPAR-99-17.

Figure 5.6. Calculated fractional crystallization paths overlaying the major and trace element chemistry diagrams of the Roberts Lake Syncline. Tic marks represent 10 % steps of fractionation. Red line indicates fractionation from the sample with the most primitive trace element pattern, although not having the most primitive major element compositions. Samples fall along a calculated fractionation trend up to 50 % fractionation. The same can be seen on calculated fractionation patterns on a multielement diagram using increments of 20 %. Blue line shows the calculated fractionation part of the most primitive sample of Group 2 indicating that the geochemical behaviour is dominated by olivine fractionation. For further details see text.

5.4.2 Partial melting and primary magma compositions

A division can be seen in Ti/V ratios, which are dependent on the oxygen fugacity of a magma and subsequent fractionation and differentiation processes that can be used to determine its environment of formation (Fig. 5.5b; Shervais, 1982). All samples from Group 1 have Ti/V ratios between 20 and 50, suggestive of continental flood basalt-type magmatism (Shervais, 1982). All samples have low Tb/Yb_{PM} ratios below 1.8, suggesting shallower, larger degree partial melting if their REE systematics remain unaffected by additions of continental crust during ascent. The generally high La/Sm_{PM} ratios of Group 1 are either indicative of an enriched mantle source, or through addition of crustal contaminants. The enriched mantle source can be generated through the addition of material with higher La/Sm ratios, e.g., recycled oceanic crust and/or subduction-related sediments, or metasomatism of the mantle by the addition of small-degree partial melts that can enrich the mantle in La relative to Sm (e.g., Sun and McDonough, 1989). Any melt generated from enriched mantle will have La/Sm_{PM} values that are uniformly higher than those of the primitive mantle, i.e., > 1. However, addition of even small quantities of continental crust to an olivine-saturated primitive melt could generate a similar trace element signature (e.g., Mungall, 2007). Group 2 samples all have Ti/V ratios below 20.

Although fractional crystallization has played a role in the evolution of the Group 1 samples, it is not in agreement with the linear trend along the Nb/Y vs Zr/Y diagram (Fig. 5.5d). This behaviour can either be the result of different degrees of partial melting (Fitton et al., 1997) or could be the result of two distinct

mantle sources (e.g., Kastek et al., 2018; Section 4.5.2.1). In the case of the Beuparlant Formation, the differing mantle sources were identified using Nd isotopes (Kastek et al., 2018; Section 4.4.3).

A Sm-Nd isotopic ratio has been obtained for Group 1 on sample NQO-13-03. This sample has an $\epsilon_{\text{Nd}2.0\text{Ga}}$ of +3.5. This value lies below the isotopic ratios for depleted mantle at 2.0 Ga ($\epsilon_{\text{Nd}2.0\text{Ga}}$ of +6), but is not enough to assess the possibility of mixing of different mantle endmembers.

Table 5.3. Results of Sm-Nd isotopes for sample NQO-13-03

Group	Sample	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\epsilon_{\text{Nd}2.0\text{Ga}}$
Group 1	NQO-13-03	4.52	16.06	0.17	0.512463	3.47

2 σ uncertainties of $^{147}\text{Sm}/^{144}\text{Nd}$ are 0.5 %

2 σ uncertainties of $^{143}\text{Nd}/^{144}\text{Nd}$ are 0.000011, given by the reproducibility of the standard

ϵ_{Nd} values are calculated relative to a present day $^{143}\text{Nd}/^{144}\text{Nd}$ CHUR value of 0.512630

The most primitive sample from Group 2 does not necessarily equate to the primary magma of that suite, given the potential for post-melting modification of the primary magma (e.g., fractionation). Herzberg and Asimow (2015) have developed the PRIMELT3 software which models both batch and accumulated fractional melting using hybrid forward and inverse models to incrementally add or subtract olivine from an evolved lava composition until a melt fraction is generated, which can be formed by partial melting of mantle peridotite and replicates the major element composition of the starting magma through fractionation or accumulation of olivine. Through this modelling, PRIMELT can also calculate the degree of partial melting and mantle potential temperature (T_P), required to generate a primary magma from a more-evolved magma composition. These techniques can only be applied to Group 2, as the Group 1 fractionation assemblages start with plagioclase.

For PRIMELT3 modeling of all samples in Group 2, $Fe^{2+}/\Sigma Fe$ has been set to 0.9. PRIMELT3 calculated primary magma compositions from peridotite melting for 4 samples from Group 2. In the remaining samples the addition of olivine never produced a suitable starting position, indicating that they cannot be modeled as products of melting of lherzolite mantle.

The calculated primary magmas for Group 2 samples are summarized in the Supplementary Data. They contain 18.42-19.48 wt. % MgO, only slightly more magnesian than the most magnesian members of the suite. These magmas are in equilibrium with olivine compositions of $\sim Fo_{92}$ and would require ~ 31 % melting

of mantle peridotite with ambient mantle temperatures (T_P) between 1523 and 1554 °C using the calculation method of Herzberg and Asimov (2015).

Estimates on depth of melting have been made for basaltic magmas (e.g., Putirka, 1999; Lee et al., 2009). Whole rock chemistry barometers utilise either Na concentrations (Putirka, 1999) or SiO₂ concentrations (Lee et al., 2009). Both of these elements show irregular behaviour when plotted against MgO concentrations, which has been interpreted as the result of element mobility. We therefore should use them with caution. SiO₂ concentrations show less scatter in bivariate diagrams. FractionatePT, the approach of Lee et al. (2009), has therefore been applied to the samples, which is applicable to melts derived from peridotitic sources. The results for FractionatePT can be found in Table 5.4.

FractionatePT (Lee et al., 2009) calculates that the primary magmas of Group 2 were generated at a range of pressures from 2.0 to 2.3 GPa. At the calculated ambient mantle temperature of ~1530 °C, these pressures correspond to the spinel lherzolite stability field (Klemme and O'Neill, 2000).

These results are supported by generally low Gd/Yb_C ratios for Group 2 that also indicate melting in the absence of garnet, which corresponds to shallow melting depths.

Table 5.4. Primary magma compositions for Group 2 of the Roberts Lake Syncline as calculated by PRIMELT3. T – eruption temperature (°C), T_P^P – mantle potential temperature (°C) calculated by PRIMELT3 using method of Herzberg and Asimov (2015); T_P^L – mantle potential temperature (°C) calculated by Fractionate PT using method of Lee et al. (2009), P – melting pressure calculated by FractionatePT, Fo – forsterite content of olivine in equilibrium with the melt, F – degree of melting, % Ol – percentage of olivine added to the sample composition in order to obtain primary magma composition, Residual Mineralogy – Mineralogy of the residual mantle after melting.

Sample Name	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	NiO	P ₂ O ₅
JMPAR-99-16	47.24	0.61	9.46	0.19	1.10	9.94	0.18	19.10	11.39	0.60	0.03	0.13	0.05
JMPAR-99-17	48.42	0.63	8.69	0.27	1.19	9.98	0.18	19.64	9.77	1.07	0.02	0.09	0.04
JMPAR-99-19	47.86	0.69	10.39	0.21	1.08	9.69	0.17	18.42	10.01	1.30	0.02	0.11	0.05
JMPAR-99-20	47.44	0.65	9.72	0.22	1.13	10.03	0.17	19.48	10.18	0.71	0.10	0.11	0.05

Sample Name	T	T_P^P	T_P^L	P	Fo	F	% Ol	Residual Mineralogy
JMPAR-99-16	1412	1541	1529	2.18	91.89	0.31	12.73	Harzburgite
JMPAR-99-17	1428	1554	1531	2.29	91.88	0.34	3.88	Harzburgite
JMPAR-99-19	1408	1523	1523	2.17	91.70	0.30	11.03	Harzburgite
JMPAR-99-20	1422	1550	1495	2.05	91.90	0.32	10.49	Harzburgite

5.4.3 Metallogenic potential

The association between Proterozoic mafic-ultramafic LIPs and magmatic Ni-Cu-PGE mineralisation is well established (e.g., Ernst and Jowitt 2013; Jowitt and Ernst, 2013). In the Cape Smith belt, the Raglan deposit (Leshner, 2007) and Expo Intrusive suite (Mungall, 2007) are economic Ni-Cu-PGE deposits and several areas have undergone extensive past exploration in the Labrador Trough and the Roberts Lake Syncline, including large gabbroic sills and the Qarqasiaq complex (see Clark and Wares, 2006).

The relative abundance of chalcophile elements compared to incompatible elements is displayed using $\text{Cu}/\text{Zr}_{\text{PM}}$ and $\text{Pd}/\text{Yb}_{\text{PM}}$ ratios compared to MgO (Fig. 5.7). In S-undersaturated mafic-ultramafic magmas all of these elements are incompatible (Keays, 1995). Group 1 samples are depleted in chalcophile elements relative to incompatible lithophile elements and show a weak trend away from primitive mantle composition with decreasing MgO (Fig. 5.7). Group 2 samples display chalcophile element ratios similar to primitive mantle (Fig. 5.7). Platinum group element concentrations are available from mineralized portions of the Quarqasiaq complex (Mungall, 1998) and show overall chalcophile element enrichment (Fig. 5.7).

The abundance of chalcophile elements, especially the PGE and Cu, within a sample is indicative of the S-saturation history of that sample (e.g., Keays, 1995). The elements remain sequestered within sulphide melts in the upper mantle during partial melting, but will be released into ascending basaltic magmas when all of the sulphide in the source has dissolved in magma, hence creating fertile

magmas (Keays, 1995; Mungall and Brenan, 2014). LIPs whose compositions show variation in mantle sourcing and have uniformly low abundances of chalcophile elements are usually unprospective (Jowitt and Ernst, 2013), whereas LIPs whose compositions show wide variations in chalcophile element depletion might be considered to have been fertile at source but to have undergone variable amounts of removal of sulphide liquid during transit through the crust, hence showing potential for the existence of economic concentrations of sulphide melt in their feeder systems.

Given their generally low sulphide content, the chalcophile element chemistry of all the lavas from the Roberts Lake Syncline is likely to be controlled by the fertility of the magma and whether the magma has undergone an S-saturation event prior to emplacement or eruption. The fact that certain samples plot below the continental flood basalt array within the MgO vs. Ni diagram (Fig. 5.3a) suggests that the samples are S-saturated and have lost some Ni to an immiscible sulphide melt. This loss of chalcophile elements during S-saturation is also indicated by variations in Cu, Pd and Pt concentrations (Fig. 5.3b/c/d) and Cu/Zr and Pd/Yb ratios (Fig. 5.7), and the presence of samples with very low concentrations of Pd and Pt (e.g., Keays, 1995).

Figure 5.7. Variations in $(\text{Cu}/\text{Zr})_{\text{PM}}$ and $(\text{Pd}/\text{Yb})_{\text{PM}}$ ratios, indicating chalcophile element depletion, changing with MgO (a,b) concentrations and with increasing $(\text{Th}/\text{Yb})_{\text{PM}}$ (c,d) ratios as a proxy for increasing crustal contamination. Ratios are normalized to Primitive Mantle (PM) values of McDonough and Sun (1995). Small symbols compiled from Ernst and Buchan (2010); Barnes et al. (2015); Ciborowski et al. (2016); this thesis.

Group 1 includes both Cu-depleted and Cu-undepleted samples slightly below expected values for primary magmas. This suggests that the magmas may have undergone an S-saturation event, forming immiscible magmatic sulphides at very high silicate/sulfide mass ratios and therefore might suggest that the Roberts Lake Syncline is prospective for mineral exploration for Ni-Cu-PGE sulphide deposits.

This depletion can also be monitored using the Cu/Pd ratio (Fig. 5.8a; after Barnes et al., 1993). This diagram is a very powerful tool to assess whether or not the parental magma of a given sample became S-saturated during its evolution and experienced any sulphide segregation or sulphide addition. The diagram utilizes both the highly chalcophile nature of Pd and the marked difference in chalcophile character between Pd and Cu (Barnes et al., 1993). Both elements will preferentially partition into any sulphide liquid coexisting with a silicate magma, but Pd is much more chalcophile than Cu. This indicates that the segregation of immiscible sulphide from a silicate melt will rapidly decrease the Pd concentration of the silicate magma but will also increase its Cu/Pd ratio. The opposite will occur in silicate magmas that accumulated sulphides.

Samples, which have not lost or gained Pd preferentially to Cu will plot near the Primitive Mantle Cu/Pd ratio, indicating that the magmas that formed these samples did not undergo an S-saturation event, neither in their mantle source region nor during transit through the crust.

Group 1 data fall along a trend between Cu/Pd ratios of Primitive Mantle and higher Pd concentration towards higher Cu/Pd ratios (~200,000) and lower Pd

concentrations (0.7 ppb) (Fig. 5.8). This behaviour is indicative of the segregation of magmatic sulphides and is identical to the trend shown by suites known to be underlain by mineralized feeder intrusions elsewhere in the Circum-Superior LIP (Barnes et al., 2015). There is the possibility that this trend extends to even higher Cu/Pd ratios and lower Pd concentrations, as 6 samples have Pd concentrations below the detection limit.

The trend formed by Group 1 samples falls along the trend predicted for equilibrium fractionation and removal of immiscible sulphides from a primary magma containing 12 ppb Pd and 80 ppm Cu (Barnes et al., 2015; Yao et al., 2019) at silicate/sulphide mass ratios between 1,000 and 10,000, suggesting that immiscible sulphides have been removed from Group 1 magmas. This means that intrusions that acted as feeders to the Group 1 samples of the Roberts Lake Syncline may be prospective for discoveries of magmatic Ni-Cu-PGE sulphide deposits.

The question that would arise in that case is, what drove these magmas to S-saturation? Often, crustal contamination and the assimilation of crustal sulphur are described as a dominant mechanism during the formation of magmatic sulphide deposits (e.g., Keays and Lighfoot, 2010; Mungall, 2014). This is also the driving factor for mineralization in the Raglan horizon and the Expo intrusive suite of the Cape Smith belt further north (e.g., Leshner, 2007; Mungall, 2007). Whereas the Circum-Superior lava suites show clear indications of crustal assimilation in their Th/Yb ratios (Fig. 5.8b), no indications for crustal assimilation can be observed in the Group 1 magmas of the Roberts Lake Syncline. Crustal

contamination indicators (e.g., Th/Yb ratios) do not correlate with chalcophile element depletion, indicating that the S-saturation for these magmas was not driven by crustal assimilation of sulphur rich material. Only two of the three key factors outlined by Jowitt and Ernst (2013) (magma fertility, crustal contamination and chalcophile element segregation) are met, making Group 1 magmatism of the Roberts Lake Syncline different from conventional prospective suites. We suggest that this sort of signature could result from the saturation of the magma with sulphide liquid solely due to the high degrees of fractional crystallization shown by these highly evolved basaltic magmas in the absence of any significant degree of crustal assimilation (c.f., Mungall, 2014 Fig. 8). In this scenario, we would anticipate that silicate/sulfide mass ratios would be very high, generating small quantities of high PGE-tenor sulfides that could only be expected to form economic concentrations if they were able to settle in a quiescent, slowly crystallizing magma chamber. Since no layered intrusions has been observed that could be considered as feeders to the Group 1 magmas in the Roberts Lake Syncline, and it is considered to be a klippe transported an unknown distance from its original roots, we conclude that it is unlikely that economic mineralization will be associated with the Group 1 lavas.

Group 2 samples have high concentrations in chalcophile elements and plot at compositions slightly depleted relative to those of primitive mantle-derived melts, lying below the more extremely contaminated and depleted members of Circum-Superior suite compositions. They therefore represent fertile magmas that have undergone some limited degree of sulphide segregation in transit. The

samples form a trend predicted for equilibrium fractionation of removal of immiscible sulphides at silicate/sulphide mass ratios $>10,000$, suggesting that, although the Cu/Pd ratios are lower as for Group 1, the mineralized sulphides should have higher concentrations of chalcophile elements. Mineralized intrusions occur beneath Group 2 in the structurally lower parts of the Roberts Lake Syncline. One of these, the Quaqasiaq complex, includes mineralized zones where analyses show a negative correlation between Cu/Pd ratios and Pd concentrations. Some of these values fall along accumulation lines of immiscible sulphides with silicate/sulphide mass ratios between 1,000 and 10,000. Some of the samples plot above the Cu/Pd ratios of the primitive mantle, while still showing high Pd concentrations (> 10 ppb). This uncommon behaviour is likely the result of local redistribution of Pd during regional metamorphism (Liu et al., 2016). The geochemical similarities between the samples of the Quaqasiaq complex and the ultramafic and mafic samples of the Circum-Superior LIP suggest they are related, and probably represent feeders, to the Group 2 lavas of the Roberts Lake Syncline. The slight increase in Cu/Pd shown by the two Group 2 samples which were analyzed for Pd indicates that they have probably reached saturation with a small quantity of Pd-rich sulfide melt, possibly corresponding to the mineralized sills noted several km below them in the volcanic pile.

Figure 5.8. (a) $(Th/Yb)_{PM}$ vs. $(Nb/Yb)_{PM}$ ratios. N-MORB composition is from Hofmann (1988), upper continental crust (UCC) composition is from Taylor and McLennan (1985), and enriched mantle (EMI, EMII) and HIMU values are from Condie (2001). (b) Cu/Pd ratio compared to Pd concentration (after Barnes et al., 1993). The dashed lines toward the top left side correspond to modelled magmas undergoing equilibrium fractionation and removal of immiscible sulphides from a primary melt containing 12 ppb Pd and 80 ppm Cu (Barnes et al., 2015; Yao et al., 2019) with varying R-factors. The dashed line towards the bottom right side corresponds to sulphide removal with varying R-factors. Ratios are normalized to Primitive Mantle (PM) values of McDonough and Sun (1995). Small symbols compiled from Ernst and Buchan (2010); Barnes et al. (2015); Ciborowski et al. (2016); this thesis.

5.4.4 Testing correlation of the Roberts Lake Syncline magmatism with units to the north or south

5.4.4.1 Cape Smith belt magmatism

The Cape Smith belt is located to the northwest of the Roberts Lake Syncline. Magmatic activity in the southern portion of the Cape Smith belt is subdivided into two events. It is dominated by the Povungnituk basalts (Beauparlant Formation, Hynes and Francis, 1982; Francis et al., 1983), interpreted as continental flood basalts, and associated dolerite sills dated at 1998 ± 6 Ma (Kastek et al., 2018; Section 3.2.3.1) and the younger ca. 1880 Ma Chukotat event (Wodicka et al., 2002; Randall, 2005; Bleeker and Kamo, 2018).

The older 1998 Ma Beauparlant Formation consists of mainly massive, basaltic flows with rare pillows and sills interlayered with black shale and greywacke, tuff, pyroclastic rocks and, in places, siliceous material (St-Onge et al., 1992; Mungall et al., 2007). Northward (up section), the sequence is increasingly dominated by pillowed and massive tholeiitic basalt with minor interflow black shale, cherty sediment, tuff and agglomeritic tuff. The chemical compositions of the Beauparlant Formation are quartz-normative, Fe and Ti-rich basalts. Their relatively evolved nature is indicated by their low MgO contents (range ~6 wt. %, with maximum of ~10 wt. %) and elevated Fe_2O_3 contents (range ~17 wt. %) (Modeland et al., 2003; Kastek et al., 2018; Section 4.4.2). They display variable multielement patterns that lie between E-MORB

compositions and OIB-like trends. All of the samples from the Beuparlant Formation indicate no or only minor interaction with crustal material.

Based on the variation in trace element ratios and isotopic compositions, the Beuparlant Formation is interpreted to have been formed by mixing of two different mantle endmembers (Modeland et al., 2003; Kastek et al., 2018; Section 4.5.2.1). These include an E-MORB-like endmember, with flat multielement patterns and a more enriched OIB-like endmember, with steeper multielement diagram patterns. This is supported by varying Nd-isotopic ratios ($\epsilon_{\text{Nd}2.0 \text{ Ga}} = +2.1$ to $+3.8$) that correlate with trace element compositions and ratios (Kastek et al., 2018; Section 4.4.3). The melting of the OIB-like endmember occurred over a range of depths, indicated by varying Tb/Yb ratios (Fig. 4.9). The Beuparlant Formation forms part of the regional 1998 Ma Minto-Povungnituk LIP (Kastek et al., 2018; Section 4.5.1), that includes the 1998 ± 2 Watts Group ophiolite in the northern Cape Smith belt (Parrish, 1989), 1998 ± 2 Ma Minto dykes (Buchan et al., 1998) and 1999 ± 2 Ma Lac Shpogan dykes (Hamilton et al., 2016) in the interior of the Superior craton, as well as the paleomagnetically linked Eskimo Formation in the Belcher Islands (Buchan et al., 1998).

The ca. 1880 Ma products of magmatism of the Chukotat Group comprise mainly ultramafic to mafic volcanic rocks and intrusive rocks with a minor amount of fine grained siliciclastic sediment intercalated amongst the volcanic sequence. The thickness of the Chukotat Group varies between 4 and 7 km (Hynes and Francis, 1982; Baragar, 2008).

The volcanic rocks of the Chukotat Group are predominantly basaltic pillow lavas and subaqueous flows (Hynes and Francis, 1982; Francis et al., 1983) although massive flows with ropy surfaces and polyhedral jointing are not uncommon (Leshner, 2007). There are three distinct petrographic types of basalts in the Chukotat Group which are distinguished on the basis of phenocryst phases present (Hynes and Francis, 1982): i) olivine-phyric basalts with MgO ~12-19 wt. %; ii) clinopyroxene-phyric basalts with MgO ~8-12 wt. %; and iii) plagioclase-phyric basalts with MgO ~4-7 wt. % (Hynes and Francis, 1982; Francis et al., 1983; Picard et al., 1990; Baragar, 2007a,b, 2008). The lower portion of the Chukotat Group consists mostly of olivine-phyric basalts grading into clinopyroxene-phyric basalts, while plagioclase-phyric basalts become more common further north in the upper portion of the Chukotat Group (Hynes and Francis, 1982; Francis et al., 1983; Picard et al., 1989a, b).

The olivine-phyric and pyroxene-phyric lavas of the Chukotat Group can be linked to the plagioclase-phyric lavas via the process of fractionation (Hynes and Francis, 1982; Francis et al., 1983; Ciborowski et al., 2016). Ciborowski et al. (2016) have shown that the primitive lavas for the Chukotat Group can be formed by ~33 % melting of mantle peridotite with an average mantle temperature of 1530 °C and depths between 68 and 135 km (1.73-3.38 GPa). This would imply melting in the garnet stability field, and the authors attribute the lack of HREE depletions to high percentage of melting, which would not leave residual garnet in the mantle source.

5.4.4.2 Labrador Trough magmatism (New Quebec orogen)

The Labrador Trough is located to the south from the Roberts Lake Syncline. Magmatism during Cycle 1 of the Labrador Trough ranged from 2169 to 2142 Ma. The older age (Cramoilet Lake sills, 2169 ± 4 Ma, Rohon et al., 1993) overlaps with the 2172-2167 Ma Biscotasing dykes of the Superior craton to the west and 2170-2169 Ma Payne River dykes of the Superior craton to the north (Buchan et al., 1993; Halls and Davis, 2004; Hamilton et al., 2017) and the Otish sills in a Paleoproterozoic sedimentary basin overlying the Superior craton (Hamilton and Buchan, 2016). Collectively, they represent the Biscotasing LIP (Ernst and Bleeker, 2010). The Biscotasing dykes form a major linear dyke swarm that cuts across a large portion of the Superior craton. Together with the ca. 2169 Ma Payne River and the 2170 Ma Tasiataq dykes (Hamilton et al., 2017) they define a radiating swarm, which converges toward the Labrador Trough (Hamilton et al. 2017).

The mafic dykes and volcanic provinces have been attributed to (1) a plume centre approximately along the eastern margin of the Superior craton (Ernst and Bleeker, 2010; Ernst and Buchan, 2010; Hamilton et al. 2017), or (2) rift-related, focused ascent of magma derived from the immediately underlying fertile mantle (Maurice et al., 2009; Maurice and Francis, 2010; Milidragovic et al., 2016).

The ca. 2.17 Ga Biscotasing dyke swarm of the southern Superior Province comprises quartz-tholeiitic dykes with variable trace element and TiO_2 concentrations (Halls and Davis, 2004; Halls et al., 2005). Milidragovic et al.

(2016) showed three distinct magmatic lineages within the Biscotasing dykes. Type 1 Biscotasing dykes are characterized by fractionated REE profiles and pronounced relative depletions in Nb and Ta on primitive mantle normalized multielement diagrams, interpreted to reflect addition of a crustal component to a melt of a fertile mantle source. Compared to other Biscotasing dykes, they have relatively steep HREE profiles, indicating a deeper melting origin, near the spinel-garnet lherzolite transition. Type 2 Biscotasing dykes have lower absolute trace element abundances, relatively unfractionated REE ratios, with nearly flat HREE profiles and only minor Nb-Ta anomalies. These compositions are consistent with melts of a fertile mantle that did not experience significant crustal contamination. Type 3 dykes have trace element abundances that are intermediate between those of Type 1 and Type 2 Biscotasing dykes, and are characterized by fractionated LREE, but relatively flat HREE profiles, with prominent depletions in Nb and Ta relative to similarly compatible LREE, as well as negative Ti-anomalies. The ca. 2.17 Ga quartz-tholeiitic Payne River dykes of the north-eastern Superior Province have weakly fractionated primitive mantle normalized trace element profiles, and low abundances of HFSE and light and middle REE (Maurice et al., 2009). The major and trace element contents of the Payne River dyke swarm are similar to the Type 2 dykes of the Biscotasing dyke swarm (Milidragovic et al., 2016), representing melts of a fertile mantle. Neodymium isotopic data is available for the Payne River dykes and display $\epsilon_{\text{Nd}(2.17 \text{ Ga})}$ values between +3.5 and +4.2 (Maurice et al., 2009).

The younger end of the Cycle 1 magmatism is represented by the Rivière du Gué dykes, which have a younger age of 2149 ± 3 Ma (Maurice et al. 2009). Rivière du Gué dykes have similar HREE, but elevated La/Yb ratios compared to the Payne River dykes and the Group 1 Biscotasing dykes and show clear Nb-Ta depletions on primitive mantle normalized multielement diagrams (Maurice et al., 2009). This is also reflected in their $\varepsilon_{\text{Nd}(2.15 \text{ Ga})}$ values that vary between -1.2 and +2.1, with one exception of a dyke with an $\varepsilon_{\text{Nd}(2.15 \text{ Ga})}$ value of +4.8 (Maurice et al., 2009). In their plume related interpretation Ernst and Bleeker (2010) speculated that the 2170 Ma pulse corresponds to plume arrival and that the 2140 Ma pulse corresponds to the onset of rifting on the eastern margin of the craton. Maurice et al., 2009 also attributed the occurrence of the Rivière du Gué dykes to a possible fracture set associated with a failed rift in a thick Archean lithosphere.

Two main formations of basalts are observed in the Labrador Trough that belong to the younger Cycle 2 event (ca. 1880 Ma) and can be correlated with similar-aged Circum-Superior LIP activity around most of the margin of the Superior Province. These are the Hellancourt Formation in the northern part (Machado et al., 1997; Skulski et al., 1993) and the Willbob Formation in the central part of the trough (Findlay et al., 1995; Rohon et al., 1993), which are geochemically equivalent and most probably belong to the same magmatic event (Rohon et al., 1993). Both formations are intruded by mafic and ultramafic Montagnais sills (Machado et al., 1997; Findlay et al., 1995; Rohon et al., 1993; Skulski et al., 1993).

The Hellancourt Formation consists of plagioclase glomeroporphyritic pillowed and massive basalt flows overlain by a thick pile of aphyric massive and pillowed basalts. The aphyric basalts have major-element compositions and REE patterns that are similar to Chukotat plagioclase-phyric basalts in the Cape Smith belt (Skulski et al., 1993). The plagioclase glomeroporphyritic (GMP) gabbro sills and basalts described by Skulski et al. (1993) have relatively flat chondrite-normalized trace-element profiles, and the basalts have slightly higher REE and incompatible trace-element (e.g., Nb) abundances than aphyric basalts (Skulski et al., 1993). The $\epsilon_{\text{Nd}}(1875 \text{ Ma})$ values of GMP basalts range from +3.4 to +4.8, similar to the range of values found in both GMP gabbro chilled margins (+2.8 to +4.0) and aphyric basalts. These data indicate that the mantle source of GMP and aphyric magmas had a time-integrated depletion in Sm relative to Nd (Skulski et al., 1993). The samples from Ciborowski et al. (2016) used for comparison were divided based on these petrographic descriptions by Minifie (2010). Re-evaluation of the data has shown that they are mislabelled from the mapping done by J.E. Mungall, who provided the samples. According to his field notes, the samples labelled as Pl-phyric in Minifie (2010) represent the aphyric Hellencourt Formation and the samples labelled aphyric represent the plagioclase glomeroporphyritic samples (Mungall, pers. comm., 2019). The labels should therefore be switched, which has been applied in the diagrams shown.

The majority of sills and basalts show limited chemical variations. The aphyric basalts have Al and Si content that lie at the projected end of an olivine-accumulation trend defined by peridotite sills. The basalts overlap in Al and Si

content with olivine gabbros and ferrogabbros in equigranular sills. The basalts have major-element compositions that are similar to Chukotat plagioclase-phyric basalts in the Cape Smith belt, and both trend to higher Fe contents than low-Mg MORB basalts and glasses (Francis et al., 1983). The low Ni and Cr abundances, and high Fe/Mg values of the basalts reflect their evolved nature. Primitive (high-Mg) basalts are absent in the north but are found in the central part of the orogen (Willbob Formation, Doublet zone).

The Willbob Formation is a thick accumulation of tholeiitic basalt. Most of the columnar basalts are pyroxene basalts; massive basalts are commonly plagioclase-phyric (Rohon et al., 1993). The basalts show chemical features characteristic of extensional volcanism similar to present-day MORB and are impoverished in HFSE and enriched in LILE and have chemical compositions similar to olivine-phyric Chukotat lavas of the Cape Smith belt.

The Montagnais sills range in composition from olivine gabbro to gabbroic anorthosite. The sills are zoned with gabbro at the top and bottom and ultramafic cumulates in the centre. Rohon et al. (1993) divided the sills into three groups, from which only the uppermost sills intruded into the Hellancourt and Willbob Formations.

5.4.4.3 Comparisons with Roberts Lake Syncline magmatism

The complete magmatic succession of the Roberts Lake Syncline has been previously mapped as part of the same magmatic unit and labelled as belonging to the ca. 1880 Ma Hellencourt Formation (Hardy, 1976).

The two observed geochemically different and unrelated groups within the Roberts Lake Syncline are therefore compared with the magmatic rocks of the Labrador Trough, including both Cycle 1 and Cycle 2. Additionally the units are compared with the magmatic units of the Cape Smith belt, with which a correlation has been suggested based on stratigraphic similarities (Ferron et al., 2000; Madore and Larbi, 2001; Bleeker and Kamo, 2018; Rohon et al., 1993; Findlay et al., 1995; Clark and Wares, 2006; Henrique-Pinto et al., 2017), as well as with several suites from the Circum-Superior LIP. All of the aforementioned units for comparison are included in Figures 5.2-5.8.

The multielement patterns of Group 1 differ from the multielement patterns of the Hellancourt Formation collected from south of the Arnaud River (Ciborowski et al., 2016). Group 1 magmas exhibit higher overall HFSE and LREE concentrations (6-11 times primitive mantle) than the Hellancourt rocks (2-8 times primitive mantle). They only overlap in HREE concentrations (2-8 times primitive mantle). While higher overall concentrations in incompatible immobile elements could be caused by fractional crystallisation, their ratios should remain uniform. Trace element ratios along the multielement diagram are higher for Group 1 magmas than for Hellancourt basalts, showing that these two magmatic

suites are not correlated to each other (Fig. 5.9). This indicates that the majority of the Roberts Lake Syncline magmatism (Group 1) is not related to the Cycle 2 of the Labrador Trough.

Group 2 samples have low concentrations of immobile trace elements and have an overall flat pattern on a multielement plot. This pattern is similar to those of the Hellancourt Formation (Labrador Trough) and Chukotat Group (Cape Smith belt). The matching flat trace element patterns support previous correlations made between the Hellancourt Formation and the plagioclase-phyric rocks of the Chukotat Group as part of the ca. 1880 Ma Circum-Superior LIP by Ciborowski et al (2016). While sharing similar trace element patterns, Group 2 and Hellancourt Formation can be distinguished by overall trace element concentrations. Group 2 has trace element concentrations consistent with the olivine-phyric and pyroxene-phyric lavas of the Chukotat Group. This kind of primitive magma is not observed in the Hellancourt Formation.

The peridotites of the Qarqasiaq complex and their associated gabbros uniformly overlap with the combined ultramafic portions of the Circum-Superior LIP. This suggests that the Qarqasiaq sill represents a feeder to the stratigraphically higher lavas of Group 2.

Figure 5.9. Multi-element diagram for incompatible elements with selected geological units from the Labrador Trough and Cape Smith belt for comparison. Comparisons show that Group 1 samples are not related to Circum-Superior LIP magmatism in either orogens. Group 2 shows similar patterns to both displayed geological units but concentrations only overlap with the data from the primitive Chukotat Group. Data for Hellancourt Formation and Chukotat Group from Ciborowski et al. (2016).

Figure 5.10. Multielement diagram for incompatible elements with selected analyses from the radiating dyke swarm associated with the Biscotasing LIP and the Beauparlant Formation of the Cape Smith belt for comparison. Group 1 samples overlap partially with the data from the Payne River dykes and fully fall into the range defined by E-MORB and mixing data from the Beauparlant Formation. Data for Beauparlant Formation from this thesis, Biscotasing dykes from Ernst and Buchan (2010), and Payne River dykes from Maurice (2009).

Ciborowski et al. (2016) assessed the primary magmas and potential mantle temperatures for suitable lavas of the Circum-Superior-LIP. Their results showed that the mantle temperatures for the LIP ranges from 1438 to 1686 °C, with an average at 1562 °C. The calculated ambient mantle temperatures of 1523 – 1554 °C calculated for Group 2 lies within these values, but are slightly higher than the ambient mantle temperatures calculated for the Hellencourt Formation (Ciborowski et al., 2016). The values for Group 2 overlap exceptionally well with the calculated values reported for the Chukotat Group, where primary magma compositions with 18.7 wt. % MgO and ambient mantle temperatures of 1530 °C are reported (Ciborowski et al., 2016).

The lower stratigraphic position of Group 1 indicates that it is relatively older than Group 2. Although Group 1 and Group 2 are considered as a single volcanic sequence, they may not have necessarily been sourced by the same magmatic event. Older events, which should be considered as possible sources for Group 1 include the 1998 Ma Beuparlant Formation of the Minto-Povungnituk LIP (in the Cape Smith belt to the north) and the 2169 Ma Payne River dykes of the Biscotasing LIP (to the south in the Labrador Trough).

Trace element concentrations and ratios of Group 1 lavas overlap with the samples of the Beuparlant Formation that display E-MORB-like geochemical behaviour and form a trend that overlaps with the samples of the Beuparlant Formation that display a mixture of the two mantle endmembers (E-MORB and OIB) that were involved in its formation (Beuparlant MIX in Figs. 5.10, 5.11) (Modeland et al., 2003; Kastek et al., 2018; Section 4.5.2.1). This overlap is very

well developed in nearly all diagrams. Only in a diagram of La/Yb vs Zr do the mixed samples of the Beuparlant Formation display higher La/Yb ratios as well as ratios overlapping with the Group 1 data, therefore resulting in only partial overlap.

In all diagrams, the trend formed by Group 1 only partly overlaps with the compositions of the Biscotasing LIP.

Because of the similarities in geochemical composition, a link to the 2.17 Ga Biscotasing LIP cannot be ruled out. However, we provisionally prefer a link with the Beuparlant Formation data since those data more fully overlap with the observed data from the Roberts Lake Syncline. A geochemical similarity with the Beuparlant Formation would also agree with the preceding correlation of Group 2 samples and the Chukotat Group of the Cape Smith belt, showing that the volcanic part of the Roberts Lake Syncline more likely represents a continuation of Cape Smith Belt stratigraphy.

If true, the variations along the mantle array and Iceland array (Fig. 5.4c/d), would indicate interaction with the plume source involved in the formation of the Minto-Povungnituk LIP. The plume source is represented by an OIB-like endmember in Kastek et al. (2018) (Figs. 5.10, 5.11). The inferred plume centre is currently constrained along the projection of the Minto dykes and lies at the present day location of Hudson Bay. The position of the Roberts Lake Syncline at the other side of the Superior craton could explain why the geochemical plume signature is weaker in the samples of the Roberts Lake Syncline than in the rest of the Cape Smith belt.

Figure 5.11. Selected trace element diagrams for the Roberts Syncline with fields from similar events for comparison. (a) V vs. Ti/1000 (after Shervais, 1982), (b) Th/Yb vs. Nb/Yb (after Pearce, 2008), (c) La/Yb vs. Zr, and Nb/Y vs. Zr/Y. In all diagrams only partial overlap with the combined samples of the Biscotasing LIP and Group 1 samples can be seen. Group 1 falls completely into the range encircled by E-MORB and MIX data from the Beuparlant Formation. The only difference can be seen in the representative La/Yb ratios, the MIX component of the Beuparlant Formation only partially overlaps with Group 1 samples, while showing multiple analyses with higher La/Yb ratios trending towards the OIB composition of the Bauparlant Formation. Data for Biscotasing dykes are from Ernst and Buchan (2010), and for Payne River dykes are from Maurice (2009), and for the Beuparlant Formation are from this thesis.

5.5 Conclusions

Geochemical analysis of the basaltic successions of the Roberts Lake Syncline shows evidence for two distinct events. The lower and more voluminous volcanic package (Group 1) is characterised by elevated trace element patterns similar to E-MORB signatures. Although fractional crystallization was involved in the genesis of these rocks, trace element ratios indicate that mixing of magmas from different mantle sources may also have played a role. Metallogenic indicators Cu, Pd and Pt, suggest that the magmas were sourced from a fertile mantle and possibly underwent a S-saturation event, despite showing no indication of significant degrees of crustal contamination. If any economic segregations of magmatic sulphide do exist in the roots of this event, they are likely only to be found in bodies such as hypothetical layered intrusions where very small amounts of sulphide melt could have accumulated from large bodies of quiescent magma.

Geochemical comparison of the Group 1 lavas of the Roberts Lake Syncline to other regional magmatic units along the margin of the Superior craton shows similarities with both the ca. 2170 Ma Biscotasing LIP (which includes units within the Labrador Trough to the south) and the 1998 Ma Minto-Povungniutk LIP (which includes units within the Cape Smith belt to the north). There is more complete overlap of the trace element data with the latter, and on this basis the Group 1 lavas of the Roberts Lake Syncline are a preferred match with the Beauparlant Formation of the Minto-Povungnituk LIP (Cape Smith Belt).

The upper part of the Roberts Lake Syncline stratigraphy is composed of more depleted rocks with flat patterns on a multielement plot (Group 2). They have been sourced from a fertile and hot mantle and can be linked to the 1880 Ma Circum-Superior LIP based on matching trace element geochemistry. Detailed geochemical comparison suggests a closer resemblance to the chemistry observed within the Circum-Superior LIP units in the Cape Smith belt (Chukotat Group) than to the Circum-Superior LIP units in the Labrador Trough (Cycle 2).

As noted above, geochemical indicators for both magmatic successions within the Roberts Lake Syncline (Groups 1 and 2) indicate more comprehensive similarities with magmatism the Povungnituk and Chukotat events of the Cape Smith belt than to the Cycle 1 and 2 events of the Labrador Trough further south. We therefore provisionally conclude that the Roberts Lake Syncline represents a continuation of the Cape Smith belt, rather than the northernmost part of the Labrador Trough. However, direct U-Pb dating of both Groups of the Roberts Lake Syncline will be required for a definitive assessment of correlations.

5.6 References

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6 CONCLUSIONS AND FUTURE WORK

6.1 Conclusions

Multiple large igneous province (LIP) events have been identified within the eastern arm of the Superior craton and around its margin, spanning in age between 2.2 Ga and 1.9 Ga. This thesis successfully identified and correlated units within the Cape Smith belt and Roberts Lake Syncline with some of these. This research focussed on volcanic units of the northeastern part of the Superior craton, specifically the Povungnituk and Chukotat formations of the Cape Smith belt and the two volcanic series of the Roberts Lake Syncline in the northern Labrador Trough of the New Quebec orogen. The thesis successfully identified and correlated these important Paleo-Proterozoic intraplate magmatic suites with previously constrained LIP events utilising U-Pb geochronology and geochemistry. Their correlations are shown in Figure 6.1.

Figure 6.1. Outline of the eastern arm of the Superior craton. Dotted circles indicate the location of possible plume centers that formed large igneous province (LIP) magmatism in and around the craton. Different LIPs are shown in different colors. Modified from Legault et al. (1994); Buchan et al., 1998; St-Onge et al. (2004); Clark and Wares (2006); Goodfellow (2007); Maurice et al. (2009); Ernst and Bleeker (2010); Nilsson et al., (2010); Hamilton et al., (2017); Sahin and Hamilton (20

Table 6.1. Summary of obtained U-Pb ages of magmatic baddeleyite and zircon.

Sample No.	Rock type	Geological unit	Age (Ma)	Technique
BLS-73-31	Dolerite sill	Povungnituk Group	1998 ± 6	ID-TIMS
BLS-73-183	Dolerite sill	Povungnituk Group	1967 ± 7	ID-TIMS
BL-73-331	Dolerite sill	Chukotat Group	1874 ± 3	ID-TIMS
BL-73-180	Dolerite sill	Chukotat Group	1861 ± 28	IN-SIMS
BL-73-M260	Dolerite dyke	Povungnituk Group (not associated)	819 ± 36	IN-SIMS
BL-73-M326	Dolerite sill	Povungnituk Group	2093 ± 86	LA-ICPMS
SAB-87-273A	Dolerite sill	Povungnituk Group	2079 ± 62	LA-ICPMS

Seven U-Pb dates were determined for magmatic baddeleyite and zircon crystals in samples collected throughout the Cape Smith belt.

Although the identification of U-bearing minerals was possible on multiple samples, greenschist to amphibolite facies metamorphism during the Trans Hudson orogeny resulted in zircon overgrowth on most samples from the Cape Smith belt. Three samples were identified that crystallized baddeleyites large enough for separation and with no visible secondary overgrowths. Additionally, four samples that crystallized large baddeleyite crystals, and which showed visible zircon rims were chosen for in situ dating approaches. The most precise ages were obtained via ID-TIMS and yielded two ages for the Beuparlant Formation of the Povungnituk Group and one age for an interpreted feeder dyke of the Chukotat Group. The samples for the Povungnituk Group yielded ages of 1998 ± 6 Ma and 1967 ± 7 Ma. The older age (ca. 1998 Ma) is interpreted to represent the main pulse for the Beuparlant Formation volcanism and correlates perfectly with ages obtained from the Minto and Lac Shpogan dykes in the interior of the Superior craton, as well as the Watts Group of the Cape Smith belt.

A U-Pb TIMS analysis on a gabbroic dyke within the Nuvilik Formation yielded an age of 1874 ± 3 Ma. Together with previously published ages that correlate with the ore-forming processes related to Chukotat magmatism, this age (ca. 1874 Ma) shows that Chukotat magmatism was active for at least 5 myr and might have lasted as long as 12 myr. Additional magmatic ages obtained via SIMS and LA-ICP-MS techniques show larger error margins. Two ages obtained for sills intruding into the Dumas Formation show ages that, depending on the

calculation method, yield ages that either correlate to the previously dated 2038 ± 3 Ma Korak sill, or show errors that overlap with the entire observed age range for the Povungnituk Group. A SIMS analysis on a sample obtained from the top of the Chukotat Formation yielded an age of 1861 ± 28 Ma, proving that the entire package of the Chukotat Group belongs to the same temporally constrained large magmatic event. The Chukotat Group belongs to the Circum-Superior LIP, which circumscribes most of the Superior craton.

The age obtained for the Beuparlant Formation has significant implications for regional geological correlations.

The ca. 1998 Ma age correlates with ages obtained for the Watts Group in the northern Cape Smith belt and the Minto and Lac Shpogan dykes in the interior of the Superior craton. Additionally, the Minto dykes have been linked on a paleomagnetic basis to the Eskimo Formation of the Belcher Islands. Together, these units form the Minto-Povungnituk LIP, covering an area over 400,000 km². The correlation between these coeval events was further tested on the basis of geochemistry. Based on geochemical observations, two distinct domains are identified within the Minto-Povungnituk LIP. The first domain is located to the north of the Superior craton and comprises the Beuparlant Formation and the Watts Group within the Cape Smith belt. For the Beuparlant Formation 84 samples were analyzed for major and trace element geochemistry, and 8 samples for Nd isotopes. Geochemical data for the Beuparlant Formation varies between two distinct signatures. Major and trace element compositions form an array between E-MORB and OIB-like compositions. This spread does not

correlate with stratigraphy and can also be observed in Nd isotope compositions. It is therefore likely that the different geochemical signatures are the result of mixing between a plume melt and melts of the surrounding mantle lithosphere, which fed the Beuparlant Formation.

The Watts Group shows geochemical signatures similar to the Beuparlant Formation, varying from E-MORB to OIB-like compositions that are interpreted to be the result of the same two mantle endmembers.

The second domain is located to the south in the center of the Superior craton and comprises the Minto dykes and the Eskimo Formation. Most of the southern rocks are highly contaminated and contamination is inversely correlated with magma evolution, suggesting contamination during dyke propagation rather than within the originating magma chamber. The most evolved samples propagated in thin dykes too cold for contamination and represent an isotopic composition close to their parental magma. These Nd-isotopic signatures significantly differ from those of the units in the north. This indicates that the domain was fed by a vastly different source than the northern domain.

Two different models are proposed to explain the geochemical differences observed in the two domains described above. In the first model, magmatism of the southern domain could be explained by plume-induced melting of subcontinental lithospheric mantle beneath the Superior craton and magmatism of the northern domain could be generated by mixing of melts from the plume head and depleted ambient mantle.

The second model involves the direct transport of different primary mantle material either in two adjacent plumes, or within a single plume, from deep mantle sources. In the latter interpretation, the different plume compositions would remain on separate sides of the plume during ascent and would imply that the two spatially separated compositions in the Minto-Povungnituk LIP directly represent the composition of the deep mantle on different sides of the plume at the time of emplacement. The interpretation of two nearby but separate, coeval plumes would also imply two distinct compositions of the deep mantle. Using the Minto-Povungnituk LIP as an example, bilateral asymmetry in LIP events could then be used as a tool to map and locate deep mantle anomalies, like the modern large low-shear-velocity provinces (LLSVPs), back through Earth's history.

The extent of the Minto-Povungnituk LIP could be even larger, including proposed correlations of the Cape Smith belt and the Roberts Syncline. Multiple authors have suggested that the Roberts Lake Syncline represents the eastward and southward continuation of the Cape Smith belt. However, other authors have interpreted the Roberts Lake Syncline volcanism as the northern continuation of the Labrador Trough magmatism. Geochemical analysis of 94 samples from the Roberts Lake Syncline (northern Labrador Trough) have been studied in order to test their correlation with their surrounding geological units. This detailed geochemical sampling throughout the syncline revealed that the basaltic successions of the Roberts Lake Syncline show evidence for two distinct events. The lower and more voluminous volcanic package is characterised by elevated

trace element patterns similar to E-MORB signatures. Trace element signatures of these rocks show that they are not purely related by fractional crystallization and suggest that mixing of two distinct mantle endmembers might have been involved in their formation. Geochemical compositions and metallogenic indicators (Cu, Pd, and Pt) suggest that the magmas were sourced from a fertile mantle and possibly underwent a S-saturation event, despite showing no indication of significant degrees of crustal contamination. Therefore, economic segregations of magmatic sulphide are rather unlikely. The geochemical characteristics of these lower units show similarities to both the 2170 Ma Biscotasing LIP of the eastern Superior craton and its units within the Labrador Trough, as well as the units of the 1998 Ma Minto-Povungnituk LIP of northeastern Superior craton within the Cape Smith Belt. The data overlaps more completely with the observed geochemical variations of the Minto-Povungnituk LIP and a connection to these units is favoured.

The upper volcanic package of the Roberts Lake Syncline is composed of picrites and komatiites that show depleted, MORB-like geochemical signatures. Their concentrations of trace elements and metallogenic indicators show that they were sourced from a fertile and hot mantle and can be linked generally to the 1880 Ma Circum-Superior LIP which includes both the Chukotat Group of the Cape Smith Belt and the Cycle 2 units of the Labrador Trough. Detailed geochemical comparisons suggests a closer resemblance of the upper Roberts Lake Syncline lavas with the Chukotat Group of the Cape Smith belt.

Since both magmatic successions within the Roberts Lake Syncline show a stronger geochemical affinity towards the magmatism observed within the Cape Smith belt to the north, then a connection to these units is favoured. It can therefore be concluded that the Roberts Lake Syncline lavas are a continuation of the Cape Smith belt magmatism, rather than being the northward continuation of the Labrador Trough magmatism.

This thesis therefore enables us to identify a widespread 1998 Ma Minto-Povungnituk LIP event that reached from the center of the eastern arm of the Superior craton along its western margin and all around the northern tip of the craton. The units most distant from the inferred plume center underneath Hudson Bay are located within the Roberts Lake Syncline, whose geological units have been revised and should more probably follow the nomenclature of the Cape Smith belt and be viewed detached from the previously associated Labrador Trough.

6.2 Future Work

Regional correlations are made using the data presented in this thesis, but several key questions still remain.

6.2.1 Geochronology

The exact correlation of the Roberts Lake Syncline with the stratigraphy of the Cape Smith belt or the Labrador Trough can only be confirmed when precise ages are obtained on the two magmatic units. So far the only geochronological constraints can be made with the intrusive units of the Quarqasiaq complex and the Kyak Bay intrusion (of the Roberts Lake Syncline). In both the Beuparlant Formation and the Watts Group (of the Cape Smith belt), only a sample of the enriched endmember has been dated. An age obtained for the depleted part of the northern domain (Cape Smith belt) could help strengthen or modify the interpretations made on possible mixing as a process for magma formation.

The further extent of the Minto-Povungnituk LIP should be tested in regards to the Inukjuak dykes. These dykes have been correlated by previous authors with the Eskimo Formation, Nastapoka basalts and Persillon volcanic rocks based on their geochemistry. Geochronology would test this correlation. Given the difference in dyke trend between the Inukjuak and the Minto dykes, confirmation of membership of the Inukjuak dykes in the Minto-Povungnituk LIP would have significant implications for the magma emplacement history of this LIP.

6.2.2 Geochemistry

A strong geochemical correlation between the Roberts Lake Syncline and the units of the Cape Smith belt and the Labrador Trough requires the identification of mantle source. This will require additional isotopic work on the lavas. Given the metamorphosed character of the rocks, Nd- and Hf-isotopic ratios would be the most appropriate to use. To enable the best possible comparison, this isotopic work needs to be extended into the Labrador Trough for both Nd- and Hf-isotopic systems, where both datasets are currently lacking for the Hellancourt Formation. In the Beuparlant Formation (Cape Smith belt), Hf-isotopic data are currently missing. As a more sensitive isotopic system, it would be helpful to address the mixing proposed in this thesis. Estimated errors of half an epsilon unit means that the trend observed in the Beuparlant Formation only extends with certainty around 1 epsilon unit. The current spread seen in the Nd isotopic system only separates by one epsilon unit, when taken into account the possible error of half an epsilon unit for any given datapoint. Changes in ϵ_{Hf} would be more prominent and would support the theories proposed in this thesis. Platinum group elements indicate a S-saturation event within Group 1 of the Roberts Lake Syncline. A similar dataset for the Beuparlant Formation could help support a correlation between the two units. If the same S-saturation and possible sulfide settling can be observed in the Beuparlant Formation, this could have possible implications for possible ore deposit locations within layered intrusions that have fed the Minto-Povungnituk LIP. The existence of possible layered intrusions for

the Minto-Povungnituk LIP has not been confirmed and represents a future work topic on its own.

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APPENDIX

Appendix 1. U-Pb LA-ICP-MS results for Phalabora

No.	U (ppm)	Th (ppm)	Th/U	Ratio						Age						% Disc.
				²⁰⁷ Pb/ ²³⁵ U	2 σ	²⁰⁶ Pb/ ²³⁸ U	2 σ	²⁰⁷ Pb/ ²⁰⁶ Pb	2 σ	²⁰⁷ Pb/ ²³⁵ U Age (Ma)	2 σ	²⁰⁶ Pb/ ²³⁸ U Age (Ma)	2 σ	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	2 σ	
Phalabora 1																
PB_2	2400	31	0.01	16.080	0.560	0.9040	0.0360	0.1264	0.0014	2858	35	4120	120	2048	20	-101
PB_1	3090	40	0.01	14.390	0.580	0.8090	0.0360	0.1265	0.0016	2751	39	3780	130	2050	22	-85
PB_3	5870	87	0.01	8.450	0.520	0.4750	0.0380	0.1282	0.0076	2276	56	2500	170	2073	104	-25
PB_4	1990	17	0.01	8.740	0.470	0.4850	0.0260	0.1297	0.0072	2308	50	2550	110	2094	98	-24
PB_5	971	9	0.01	16.660	0.600	0.9330	0.0360	0.1268	0.0014	2894	35	4200	120	2054	19	-105
PB_6	1425	12	0.01	13.960	0.580	0.7870	0.0350	0.1253	0.0017	2736	40	3730	130	2033	24	-85
PB_7	513	5	0.01	8.810	0.350	0.5020	0.0360	0.1274	0.0084	2326	34	2620	160	2062	116	-27
PB_8	265	2	0.01	17.410	0.670	0.9760	0.0420	0.1277	0.0017	2916	38	4320	140	2067	23	-109
PB_9	348	3	0.01	14.420	0.580	0.8070	0.0360	0.1290	0.0019	2758	39	3780	130	2084	26	-84
Phalabora 2																
PB_1	1540	28	0.01	15.920	2.200	0.9040	0.0430	0.1244	0.0140	2839	130	4090	150	2020	199	-103
PB_2	1690	29.8	0.01	16.710	2.300	0.9410	0.0450	0.1245	0.0140	2881	130	4210	150	2022	199	-109
PB_3	820	5.02	0.00	16.570	2.300	0.9460	0.0460	0.1237	0.0140	2864	130	4230	150	2010	201	-111
PB_4	908	6.09	0.00	15.300	2.100	0.8790	0.0390	0.1243	0.0140	2809	130	4010	140	2019	200	-101
PB_5	483	7.62	0.01	15.570	2.100	0.8780	0.0400	0.1247	0.0140	2812	130	4010	140	2025	199	-98
PB_6	478	5.03	0.01	14.180	1.900	0.8030	0.0320	0.1253	0.0140	2742	130	3768	120	2033	198	-86

Appendix 2. Standards for measurements of magmatic baddeleyite and zircon

No.	U (ppm)	Th (ppm)	Th/U	Ratio						Age						% Conc.
				$^{207}\text{Pb}/^{235}\text{U}$	2 σ	$^{206}\text{Pb}/^{238}\text{U}$	2 σ	$^{207}\text{Pb}/^{206}\text{Pb}$	2 σ	$^{207}\text{Pb}/^{235}\text{U}$ Age (Ma)	2 σ	$^{206}\text{Pb}/^{238}\text{U}$ Age (Ma)	2 σ	$^{207}\text{Pb}/^{206}\text{Pb}$ Age (Ma)	2 σ	
BL-73-M326																
ZOG1_1	372	474	0.91	29.740	0.790	0.7172	0.0220	0.2960	0.0034	3478	26	3480	82	3449	18	-1
ZOG1_2	431	315	0.52	30.140	0.790	0.7183	0.0220	0.3009	0.0034	3492	26	3495	83	3475	17	-1
ZOG1_3	382	376	0.70	29.670	0.780	0.7110	0.0220	0.2977	0.0034	3474	26	3466	80	3458	18	0
Z_Plesovice_1	1063	128	0.09	0.403	0.013	0.0533	0.0017	0.0546	0.0013	344	10	335	10	396	53	105
Z_Plesovice_2	1405	175	0.09	0.400	0.013	0.0536	0.0017	0.0534	0.0011	341	9	337	10	346	47	68
Z_Plesovice_3	1291	158	0.09	0.394	0.013	0.0531	0.0017	0.0530	0.0013	336	9	333	10	329	56	91
Z_Plesovice_4	1071	128	0.09	0.418	0.014	0.0552	0.0018	0.0543	0.0013	354	10	346	11	384	54	117
Z_Plesovice_5	1260	160	0.09	0.479	0.015	0.0555	0.0017	0.0628	0.0014	397	10	348	11	701	47	52
Z_Plesovice_6	967	117	0.09	0.410	0.013	0.0556	0.0017	0.0535	0.0014	348	10	349	11	350	59	64
Z_02123_1	499	416	0.60	0.336	0.013	0.0464	0.0015	0.0521	0.0018	294	10	292	10	290	79	104
Z_02123_2	545	466	0.61	0.329	0.014	0.0466	0.0016	0.0512	0.0019	289	11	293	10	250	85	99
Z_02123_3	494	408	0.59	0.339	0.014	0.0465	0.0015	0.0530	0.0019	294	11	293	9	329	81	93
Z_91500_1	74	28	0.27	1.857	0.090	0.1769	0.0068	0.0757	0.0036	1066	32	1054	37	1087	95	-7
Z_91500_2	78	29	0.27	1.821	0.086	0.1806	0.0066	0.0742	0.0032	1047	31	1066	36	1047	87	116
Z_91500_3	87	33	0.27	1.877	0.089	0.1830	0.0065	0.0732	0.0033	1074	31	1082	35	1019	91	104
Z_91500_4	83	31	0.27	1.827	0.080	0.1803	0.0063	0.0740	0.0032	1050	29	1067	35	1041	87	80
Z_91500_5	82	31	0.27	1.849	0.079	0.1791	0.0065	0.0747	0.0029	1062	29	1060	36	1060	78	28
Z_91500_6	81	29	0.26	1.858	0.078	0.1742	0.0060	0.0765	0.0026	1055	27	1037	33	1108	68	37
Z_91500_7	80	31	0.27	1.893	0.085	0.1809	0.0064	0.0757	0.0031	1070	30	1078	35	1087	82	-15
Z_91500_8	80	30	0.27	1.848	0.076	0.1763	0.0061	0.0761	0.0027	1062	28	1045	34	1098	71	11
Z_91500_9	79	29.4	0.27	1.790	0.076	0.1784	0.0064	0.0738	0.0030	1043	28	1059	35	1036	82	70
Z_91500_10	78	29	0.27	1.847	0.084	0.1802	0.0066	0.0744	0.0031	1049	30	1068	36	1052	84	54
Z_91500_11	82	31	0.27	1.901	0.085	0.1824	0.0065	0.0744	0.0029	1081	30	1077	35	1052	79	27

No.	U (ppm)	Th (ppm)	Th/U	Ratio						Age						% Conc.
				²⁰⁷ Pb/ ²³⁵ U	2 σ	²⁰⁶ Pb/ ²³⁸ U	2 σ	²⁰⁷ Pb/ ²⁰⁶ Pb	2 σ	²⁰⁷ Pb/ ²³⁵ U Age (Ma)	2 σ	²⁰⁶ Pb/ ²³⁸ U Age (Ma)	2 σ	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	2 σ	
SAB-87-273A																
ZOG1_1	394	285	0.51	29.730	0.380	0.7184	0.0220	0.2968	0.0081	3478	12	3490	82	3454	42	-1
ZOG1_2	383	273	0.51	30.110	0.370	0.7119	0.0210	0.3037	0.0082	3488	12	3464	79	3489	42	1
ZOG1_3	299	257	0.61	29.880	0.390	0.7125	0.0210	0.2982	0.0082	3478	13	3472	80	3461	43	0
Z_Plesovice_1	1470	188	0.09	0.406	0.009	0.0563	0.0017	0.0516	0.0016	345	6	353	10	268	71	111
Z_Plesovice_2	1086	133	0.09	0.412	0.009	0.0549	0.0017	0.0538	0.0018	349	7	344	10	363	75	75
Z_Plesovice_3	1373	174	0.09	0.414	0.008	0.0544	0.0016	0.0540	0.0017	351	6	342	10	371	71	58
Z_Plesovice_4	1085	134	0.09	0.403	0.010	0.0547	0.0017	0.0539	0.0018	345	7	344	10	367	75	79
Z_Plesovice_5	1223	152	0.09	0.415	0.009	0.0557	0.0017	0.0536	0.0018	353	6	349	10	354	76	97
Z_Plesovice_6	1187	144	0.08	0.404	0.009	0.0557	0.0017	0.0525	0.0018	344	6	349	10	307	78	127
Z_02123_1	514	420	0.58	0.335	0.010	0.0462	0.0015	0.0516	0.0022	293	8	291	9	268	98	127
Z_02123_2	505	411	0.58	0.344	0.012	0.0462	0.0015	0.0534	0.0024	299	9	291	9	346	102	107
Z_02123_3	492	395	0.57	0.329	0.012	0.0461	0.0015	0.0514	0.0023	288	9	290	9	259	103	85
Z_91500_1	80	30	0.27	1.841	0.065	0.1770	0.0059	0.0762	0.0034	1057	23	1050	32	1100	89	22
Z_91500_2	80	30	0.27	1.895	0.071	0.1799	0.0061	0.0766	0.0035	1067	25	1070	34	1111	91	39
Z_91500_3	80	30	0.27	1.853	0.066	0.1812	0.0061	0.0753	0.0034	1066	24	1071	33	1077	91	34
Z_91500_4	81	30	0.27	1.816	0.064	0.1788	0.0061	0.0735	0.0034	1048	23	1065	33	1028	94	0
Z_91500_5	82	31	0.27	1.835	0.062	0.1783	0.0061	0.0733	0.0032	1055	22	1057	33	1022	88	24
Z_91500_6	78	29	0.27	1.856	0.067	0.1761	0.0061	0.0758	0.0036	1054	23	1045	33	1090	95	2
Z_91500_7	77	29	0.27	1.883	0.075	0.1817	0.0063	0.0740	0.0035	1057	27	1074	34	1041	95	39
Z_91500_8	80	31	0.27	1.852	0.069	0.1822	0.0063	0.0733	0.0034	1055	24	1078	34	1022	94	31
Z_91500_9	82	30	0.26	1.825	0.066	0.1764	0.0061	0.0768	0.0036	1062	24	1047	33	1116	94	41
Z_91500_10	80	30	0.26	1.874	0.070	0.1818	0.0061	0.0753	0.0035	1070	24	1075	33	1077	93	52
Z_91500_11	80	30	0.26	1.833	0.071	0.1777	0.0062	0.0741	0.0035	1054	25	1058	33	1044	95	90

No.	U (ppm)	Th (ppm)	Th/U	Ratio						Age						% Conc.
				$^{207}\text{Pb}/^{235}\text{U}$	2 σ	$^{206}\text{Pb}/^{238}\text{U}$	2 σ	$^{207}\text{Pb}/^{206}\text{Pb}$	2 σ	$^{207}\text{Pb}/^{235}\text{U}$ Age (Ma)	2 σ	$^{206}\text{Pb}/^{238}\text{U}$ Age (Ma)	2 σ	$^{207}\text{Pb}/^{206}\text{Pb}$ Age (Ma)	2 σ	
Phalabora 1																
ZOG1_1	415	692	1.16	30.140	0.560	0.7180	0.0180	0.3002	0.0036	3489	18	3482	68	3471	19	-1
ZOG1_2	297	373	0.87	30.610	0.540	0.7269	0.0180	0.3023	0.0037	3506	17	3513	67	3482	19	-1
ZOG1_3	329	433	0.91	30.640	0.560	0.7298	0.0180	0.3002	0.0038	3507	18	3535	68	3471	20	-2
Z_Plesovice_1	1713	135	0.06	0.418	0.010	0.0548	0.0014	0.0543	0.0012	354	7	344	9	384	50	94
Z_Plesovice_2	1734	255	0.10	0.498	0.012	0.0563	0.0014	0.0632	0.0014	411	8	353	9	715	47	48
Z_Plesovice_3	952	131	0.10	0.424	0.018	0.0560	0.0023	0.0536	0.0025	359	13	351	14	354	105	28
Z_Plesovice_4	1124	143	0.09	0.410	0.010	0.0550	0.0014	0.0533	0.0013	350	8	346	9	342	55	105
Z_Plesovice_5	1083	135	0.09	0.413	0.011	0.0558	0.0014	0.0537	0.0013	350	8	350	9	358	55	91
Z_Plesovice_6	1508	192	0.09	0.406	0.010	0.0560	0.0014	0.0518	0.0011	345	7	351	9	277	49	42
Z_Plesovice_7	1165	140	0.08	0.408	0.011	0.0555	0.0014	0.0522	0.0012	347	8	348	9	294	52	50
Z_02123_1	550	455	0.58	0.346	0.012	0.0474	0.0013	0.0525	0.0018	302	9	299	8	307	78	116
Z_02123_2	524	448	0.60	0.354	0.012	0.0468	0.0013	0.0541	0.0018	308	9	295	8	375	75	85
Z_02123_3	518	434	0.59	0.341	0.012	0.0479	0.0013	0.0509	0.0018	297	9	302	8	236	82	160
Z_91500_1	80	30	0.27	1.830	0.100	0.1781	0.0066	0.0729	0.0040	1048	36	1056	37	1011	111	36
Z_91500_2	79	29	0.27	1.869	0.078	0.1811	0.0059	0.0768	0.0032	1064	28	1069	32	1116	83	41
Z_91500_3	80	30	0.26	1.853	0.078	0.1800	0.0057	0.0745	0.0030	1060	27	1066	31	1055	81	29
Z_91500_4	80	30	0.27	1.835	0.073	0.1771	0.0055	0.0747	0.0031	1051	26	1052	31	1060	84	69
Z_91500_5	80	30	0.27	1.852	0.070	0.1795	0.0056	0.0749	0.0029	1059	25	1065	31	1066	78	57
Z_91500_6	80	30	0.26	1.847	0.071	0.1804	0.0054	0.0744	0.0028	1062	25	1067	29	1052	76	112
Z_91500_7	80	29	0.26	1.852	0.074	0.1772	0.0054	0.0746	0.0030	1055	25	1052	30	1058	81	49
Z_91500_8	80	31	0.27	1.854	0.074	0.1805	0.0056	0.0757	0.0030	1056	26	1069	31	1087	79	27
Z_91500_9	80	30	0.26	1.825	0.077	0.1783	0.0056	0.0750	0.0031	1044	28	1059	31	1069	83	52
Z_91500_10	80	31	0.27	1.883	0.072	0.1788	0.0053	0.0754	0.0030	1065	25	1060	29	1079	80	45
Z_91500_11	80	30	0.26	1.835	0.075	0.1799	0.0052	0.0743	0.0031	1056	27	1065	28	1050	84	48

No.	U (ppm)	Th (ppm)	Th/U	Ratio						Age							% Conc.
				²⁰⁷ Pb/ ²³⁵ U	2 σ	²⁰⁶ Pb/ ²³⁸ U	2 σ	²⁰⁷ Pb/ ²⁰⁶ Pb	2 σ	²⁰⁷ Pb/ ²³⁵ U Age (Ma)	2 σ	²⁰⁶ Pb/ ²³⁸ U Age (Ma)	2 σ	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	2 σ		
Phalabora 2																	
ZOG1_1	199	237	0.85	30.840	4.000	0.7300	0.0220	0.3022	0.0340	3512	130	3525	81	3482	174	-1.8	
ZOG1_2	338	379	0.80	29.380	3.800	0.7013	0.0210	0.2972	0.0340	3461	130	3423	79	3456	177	1	
ZOG1_3	353	397	0.80	29.280	3.800	0.7102	0.0210	0.2915	0.0330	3459	130	3456	79	3426	176	-0.9	
Z_Plesovice_1	1118	137.5	0.09	0.416	0.058	0.0533	0.0016	0.0560	0.0069	340.5	23	334.8	9.8	452	274	57	
Z_Plesovice_2	921	104.9	0.08	0.414	0.056	0.0540	0.0017	0.0549	0.0066	342.4	22	339	10	408	269	58	
Z_Plesovice_3	1379	174.5	0.09	0.402	0.053	0.0537	0.0016	0.0533	0.0062	337.7	27	337.3	9.8	342	263	20	
Z_Plesovice_4	1077	127.8	0.09	0.402	0.053	0.0545	0.0016	0.0519	0.0060	341.9	38	342	10	281	265	134	
Z_Plesovice_5	1163	142.8	0.09	0.415	0.055	0.0543	0.0016	0.0542	0.0063	348.4	33	340.9	9.8	379	261	100	
Z_Plesovice_6	1215	145.3	0.09	0.390	0.051	0.0538	0.0016	0.0509	0.0058	333.9	37	337.8	9.7	236	263	108	
Z_02123_1	457	423	0.66	0.357	0.050	0.0467	0.0015	0.0543	0.0067	297.6	26	294.1	9.3	384	277	134	
Z_02123_2	476	435	0.65	0.349	0.050	0.0465	0.0015	0.0538	0.0068	291.4	23	292.6	9.1	363	285	100	
Z_02123_3	467	431	0.66	0.366	0.059	0.0462	0.0015	0.0569	0.0086	294	17	291.3	9.3	488	334	180	
Z_91500_1	83	32	0.28	2.030	0.300	0.1798	0.0059	0.0849	0.0130	1065	50	1065	32	1313	297	60	
Z_91500_2	80	31	0.27	2.150	0.410	0.1796	0.0061	0.0860	0.0150	1041	31	1062	33	1338	337	150	
Z_91500_3	79	29	0.26	1.802	0.240	0.1770	0.0058	0.0735	0.0086	1046	86	1051	32	1028	237	-25	
Z_91500_4	82	31	0.27	1.842	0.250	0.1812	0.0059	0.0731	0.0085	1056	88	1071	32	1017	236	6	
Z_91500_5	79	30	0.27	2.380	0.540	0.1796	0.0060	0.0890	0.0160	1057	29	1062	32	1404	344	13	
Z_91500_6	75	28	0.27	1.970	0.280	0.1793	0.0060	0.0784	0.0096	1058	71	1063	33	1157	243	36	
Z_91500_7	82	31	0.27	1.861	0.250	0.1825	0.0068	0.0751	0.0090	1049	84	1071	33	1071	241	100	
Z_91500_8	76	27	0.26	1.832	0.250	0.1748	0.0057	0.0761	0.0091	1041	84	1036	31	1098	239	30	
Z_91500_9	82	31	0.27	1.832	0.250	0.1811	0.0064	0.0738	0.0089	1030	86	1067	34	1036	244	48	
Z_91500_10	80	30	0.27	1.844	0.250	0.1772	0.0059	0.0747	0.0090	1043	85	1053	32	1060	242	60	
Z_91500_11	81	30	0.27	2.050	0.420	0.1831	0.0066	0.0743	0.0091	1040	34	1080	36	1050	247	-26	

Appendix 3. Standards for LA-ICP-MS detrital zircon analyses

No.	U (ppm)	²³⁸ U KCps	Th/U	Ratio		1 σ	²⁰⁷ Pb/ ²⁰⁶ Pb	1 σ
				²⁰⁷ Pb/ ²³⁵ U	²⁰⁶ Pb/ ²³⁸ U			
DD91-1 (2682.4 Ma)	300		0.9	13.038	0.5160		0.1832	
DD85-17 (3002 Ma)	50		0.5	18.231	0.5931		0.2229	
BNB-13-066								
<i>DD91-1-2.1</i>		909	0.51		0.5725	0.0036	0.1855	0.0005
<i>DD91-1-2.2</i>		897	0.53		0.5512	0.0037	0.1864	0.0006
<i>DD91-1-2.3</i>		927	0.53		0.5353	0.0031	0.1847	0.0006
<i>DD91-1-2.4</i>		898	0.55		0.5307	0.0032	0.1850	0.0006
<i>DD91-1-2.4B</i>		898	0.55		0.5307	0.0032	0.1850	0.0006
NK-15-2316								
<i>DD85-17-2.1</i>		122	0.36		0.5733	0.0032	0.2262	0.0008
<i>DD85-17-2.2</i>		149	0.39		0.5532	0.0028	0.2248	0.0007
<i>DD85-17-2.3</i>		74	0.31		0.5809	0.0039	0.2256	0.0009
<i>DD85-17-2.4</i>		119	0.31		0.5671	0.0032	0.2253	0.0008
<i>DD85-17-2.5</i>		128	0.31		0.5577	0.0029	0.2237	0.0008
<i>DD85-17-2.6</i>		156	0.30		0.5564	0.0029	0.2235	0.0008
<i>DD85-17-2.7</i>		111	0.27		0.5555	0.0028	0.2240	0.0007
<i>DD85-17-2.8</i>		174	0.28		0.5539	0.0030	0.2246	0.0007
<i>DD85-17-2.9</i>		129	0.27		0.5521	0.0030	0.2241	0.0008
NK-15-2321								
<i>DD85-17-3.1</i>		54	0.28		0.4526	0.0029	0.2225	0.0016
<i>DD85-17-3.2</i>		48	0.29		0.4372	0.0035	0.2227	0.0022
<i>DD85-17-3.4</i>		46	0.38		0.4602	0.0038	0.2247	0.0018
<i>DD85-17-3.5</i>		54	0.28		0.4396	0.0053	0.2294	0.0022
<i>DD85-17-3.6</i>		54	0.28		0.4390	0.0043	0.2225	0.0021
<i>DD85-17-3.7</i>		53	0.27		0.4356	0.0040	0.2246	0.0023
<i>DD85-17-3.8</i>		50	0.29		0.4531	0.0033	0.2269	0.0017
<i>DD85-17-3.9</i>		32	0.29		0.4491	0.0041	0.2262	0.0023
<i>DD85-17-3.10</i>		35	0.29		0.4569	0.0041	0.2211	0.0021
<i>DD85-17-3.12</i>		41	0.34		0.4430	0.0033	0.2227	0.0018
<i>DD85-17-3.10B</i>		41	0.34		0.4430	0.0033	0.2227	0.0018

Appendix 4. R^2 values for major elements vs. Zr of the mafic Povungnituk Group

	Depleted MPG	Enriched MPG
	R^2 vs. Zr	R^2 vs. Zr
	n = 24	n = 54
SiO ₂	0.06	0.16
Al ₂ O ₃	0.24	0.06
Fe ₂ O ₃	0.45	0.24
CaO	0.10	0.02
MgO	0.09	0.28
Na ₂ O	0.01	0.00
K ₂ O	0.00	0.06
Na ₂ O+K ₂ O	0.00	0.01
TiO ₂	0.64	0.63
P ₂ O ₅	0.83	0.84

Appendix 5. R^2 values for trace elements vs. Zr of the mafic Povungnituk Group

	Depleted MPG	Enriched MPG
	R^2 vs. Zr	R^2 vs. Zr
	n = 24	n = 54
Ba	0.07	0.04
Ce	0.69	0.87
Cr	0.01	0.14
Cs	0.43	0.27
Dy	0.77	0.80
Er	0.69	0.59
Eu	0.73	0.82
Ga	0.20	0.42
Gd	0.90	0.93
Hf	0.94	0.98
Ho	0.77	0.64
La	0.57	0.84
Lu	0.62	0.38
Nb	0.64	0.81
Nd	0.87	0.92
Pr	0.80	0.90
Rb	0.09	0.27
Sm	0.90	0.95
Sn	0.14	0.01
Sr	0.10	0.02
Ta	0.47	0.85
Tb	0.91	0.89
Th	0.02	0.76
Tm	0.64	0.47
U	0.02	0.41
V	0.05	0.00
W	0.07	0.00
Y	0.80	0.66
Yb	0.60	0.48
Zr	1.00	1.00
Ag	NA	0.03
As	0.86	0.09
Cd	1.00	0.26
Co	0.07	0.09
Cu	0.01	0.11
Li	0.03	NA
Mo	NA	0.40
Ni	0.20	0.02
Pb	0.39	0.13
Sc	0.04	NA
Tl	NA	0.03
Zn	0.61	0.00

Appendix 6. Geochemical analyses of the mafic Povungnituk Group

Sample	BLS 199 73	BLS 198 73	BLS 197 73	BLS 196 73	BL 3M 326	BLS 195 73	BLS 194 73
Location	N5-20	N5-17	N5-16	N5-15	N5-15	N5-14	N5-13
Traverse	East	East	East	East	East	East	East
Group	Depleted	Enriched	Depleted	Enriched	Depleted	Enriched	Enriched
Easting	518948	518703	518680	518703	518703	518721	518730
Northing	6804300	6805203	6805430	6805765	6805765	6805978	6806146
SiO2	45.7	47.3	47.3	47.4	48.4	45.2	49.5
Al2O3	16.8	13.1	12.85	11.5	14.8	14.4	13
Fe2O3	10.95	15.6	15.6	18.9	12.95	17.2	15.9
CaO	10.05	7.8	10.6	11.15	9.18	7.28	7.79
MgO	6.43	6.38	6.05	5.73	5.81	5.29	5.93
Na2O	2.14	3.38	2.48	0.91	3.22	3.52	3.54
K2O	0.07	0.26	0.11	0.11	0.67	0.77	0.31
TiO2	1.09	2.05	1.39	2.02	1.62	4.24	2.06
MnO	0.14	0.23	0.21	0.28	0.17	0.21	0.31
P2O5	0.12	0.18	0.1	0.11	0.14	0.41	0.2
LOI	4.68	2.55	1.94	2.37	1.91	3.15	2.87
Total	98.23	98.86	98.66	100.51	98.95	101.75	101.44
Mg#	33.61	26.07	25.06	20.72	27.89	20.96	24.33
Ba	23.2	57.2	30.6	37.1	340	302	83.8
Ce	21.7	27.1	14.1	13.3	19.9	70.5	28.4
Dy	3.24	5.65	4.34	4.49	4.2	5.31	6.15
Er	1.88	3.26	2.83	2.63	2.44	2.47	3.54
Eu	1.07	1.6	0.97	1.14	1.35	2.68	1.57
Ga	22.2	23.3	19.4	20.2	24.1		20.6
Gd	3.38	5.75	3.87	3.66	4.56	6.81	5.58
Hf	2.2	3.7	2.3	2.1	3.1	5.5	4.2
Ho	0.64	1.2	0.97	0.96	0.92	0.98	1.22
La	9.3	10.4	5.3	5.2	7.6	31.2	11.3
Lu	0.27	0.47	0.42	0.39	0.33	0.27	0.47
Nb	5.3	10.4	5.4	4.8	6.9	38.7	10.9
Ta	0.2	0.5	0.2	0.2	0.4	1.8	0.6
Nd	14.1	19.6	11	9.6	14.7	38.7	19.4
Pb	3	6	4	4	2	3	4
Pr	2.93	3.87	2.13	1.92	2.94	9.16	3.94
Rb	1	5.5	1.2	2.2	29.2	17.9	5.7
Sc	28	39	44	46	35	15	37
Sm	3.18	5.66	3.41	2.91	4.11	7.9	5.17
Sr	383	112.5	172.5	286	266	331	101
Tb	0.52	0.96	0.71	0.62	0.76	0.96	0.94
Th	0.62	0.98	0.47	0.46	0.64	2.54	1.19
Tm	0.28	0.46	0.4	0.39	0.34	0.33	0.52
U	0.14	0.3	0.16	0.13	0.19	0.56	0.33
V	229	420	370	839	337	291	425
Y	17.1	29.8	24.3	23	22.5	22.8	30.1
Yb	1.83	2.81	2.59	2.77	2.07	2.15	3.17
Zr	77	139	79	73	106	205	149
Cr	160	90	80	20	70	10	80
Co	49	51	50	58	42	45	48
Cu	38	86	170	271	102	12	41
Ni	129	65	64	56	60	17	56
Zn	86	153	94	117	77	135	160

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 186 73	BL 73 331	BLS 187 73	BLS 188 73	BLS 189 73	BLS 190 73	BLS 191 73
Location	N5-10	N5-10	N5-09	N5-08	N5-06	N5-05	N5-04
Traverse	East	East	East	East	East	East	East
Group	Enriched	Depleted	Enriched	Depleted	Enriched	Enriched	Enriched
Easting	520062	520062	519885	519757	519603	519444	519331
Northing	6808310	6808310	6807933	6807738	6807448	6807162	6806940
SiO2	42.6	44.3	44.3	48.5	44.4	46.1	44.6
Al2O3	12.05	11.4	12.35	12.7	12	13.25	12.4
Fe2O3	12.85	21	17.15	15.55	16.4	16.5	16.2
CaO	9.66	10.55	10.05	5.93	6.69	6.29	7.66
MgO	6.17	4.27	4.94	6.03	6.93	5.23	5.45
Na2O	2.05	1.28	2.36	3.81	1.08	2.13	3.68
K2O	0.3	0.4	0.69	0.27	0.1	0.75	0.28
TiO2	1.64	2.68	3.44	1.96	3.68	3.42	3.62
MnO	0.17	0.34	0.27	0.22	0.2	0.24	0.22
P2O5	0.14	0.12	0.59	0.17	0.47	0.58	0.7
LOI	10.45	2.72	2.79	4.14	7.2	5.36	3.39
Total	98.14	99.17	99.03	99.31	99.17	99.88	98.26
Mg#	29.28	14.92	19.89	25.06	26.70	21.46	22.48
Ba	61.3	287	266	77.1	21.1	75.7	196
Ce	16.4	14.4	74.7	23.6	70.2	66.5	71.6
Dy	3.86	5.53	7.19	5.75	7.03	7.56	7.94
Er	2.24	3.96	3.81	3.52	3.88	4.29	4.35
Eu	0.71	1.93	3.33	1.54	3.13	2.68	3.18
Ga	18.5	25.1	26.3	18.7	22.8	22.2	22
Gd	3.67	5.22	9.11	5.39	7.98	8.1	8.81
Hf	2.4	2.9	5.9	3.2	6.1	6	6.3
Ho	0.79	1.29	1.43	1.24	1.38	1.48	1.56
La	6.5	5.5	32.5	9.5	31.5	29.2	30.7
Lu	0.26	0.51	0.49	0.48	0.48	0.6	0.57
Nb	5.5	5.7	30.2	9.3	35.3	31.5	33.2
Ta	0.3	0.3	1.4	0.5	1.7	1.4	1.6
Nd	11.7	12	44.3	16.4	40.4	38.9	42.6
Pb			4	5		3	2
Pr	2.47	2.29	9.82	3.33	9.12	8.82	9.53
Rb	9.1	12.3	27.2	10.5	2.6	33.4	9.6
Sc	32	40	35	41	35	38	37
Sm	3.55	3.74	9.41	4.62	9.02	8.72	9.43
Sr	189	684	385	103.5	74.4	124.5	215
Tb	0.62	0.89	1.35	0.9	1.25	1.26	1.3
Th	0.52	0.43	3.16	0.9	2.5	2.29	2.17
Tm	0.3	0.57	0.53	0.55	0.55	0.55	0.6
U	0.19	0.12	0.72	0.23	0.53	0.63	0.61
V	309	298	428	452	452	468	446
Y	18.9	32.3	34.6	30.4	33.8	35.8	37.8
Yb	1.83	3.15	3.32	3.34	3.54	3.73	4
Zr	85	105	241	117	246	251	267
Cr	130		130	70	60	60	100
Co	44	55	40	44	37	49	44
Cu	49	9	56	129		54	86
Ni	74	3	26	50	39	34	45
Zn	64	159	145	128	147	150	133

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 192 73	BLS 193 73	BL3 570	BL3 568	BL3 536	BLS 265 73	BLS 264 73	BLS 263 73
Location	N5-02	N5-01	N4-20	N4-19	N4-12	N4-03	N4-02	N3-05
Traverse	East	East	East	East	East	East	East	East
Group	Enriched	Depleted	Enriched	Enriched	Enriched	Enriched	Enriched	Enriched
Easting	519240	519154	519112	518961	519669	519722	519628	519008
Northing	6806754	6806604	6810159	6810358	6812648	6815614	6815745	6817609
SiO2	50.3	39.9	40.9	36.7	47.1	47.1	43	46.9
Al2O3	12.45	11.95	12.15	12.05	14.1	13.6	14.25	13.5
Fe2O3	14.5	16.3	14.8	14.9	15.55	13	16.75	14.4
CaO	7.74	12.2	9.28	10.85	9.6	10.35	8.46	10
MgO	6.53	2.09	5.37	4.78	5.44	4.77	6.82	6.67
Na2O	3.96	2.89	2.18	1.52	1.05	2.15	2.41	3.37
K2O	0.05	0.13	0.02	0.98	0.14	0.49	0.61	0.54
TiO2	1.51	3.04	3.23	4.16	1.55	1.9	3.63	1.79
MnO	0.21	0.15	0.15	0.22	0.21	0.21	0.27	0.24
P2O5	0.11	0.75	0.27	0.4	0.18	0.16	0.57	0.15
LOI	2.69	9.7	9.09	11.6	3.12	5.89	2.93	1.59
Total	100.07	99.13	97.47	98.2	98.11	99.66	99.78	99.2
Mg#	27.97	9.95	23.83	21.67	23.17	24.03	25.98	28.54
Ba	12.2	58	4.6	132.5	53.4	105	164	139.5
Ce	13.7	94.1	45.9	66.1	40.4	26.9	67.4	23.8
Dy	4.34	8.57	6.02	7.11	5.14	5.5	7.52	5.17
Er	2.86	4.35	2.68	3.07	2.67	3.53	4.16	2.94
Eu	1	3.13	2.51	2.16	1.83	1.47	2.69	1.34
Ga	16.9	27.8	24.4	30.3	28.5	22.8	26	21.5
Gd	3.83	11.75	7.41	8.87	5.74	5.71	8.74	5.23
Hf	2.1	8.6	5.5	6.7	4	3.2	6	2.9
Ho	0.99	1.56	1.08	1.29	1.02	1.18	1.51	1.05
La	5.4	35.9	19.8	26.8	16	11	28.4	9.7
Lu	0.42	0.51	0.29	0.36	0.39	0.47	0.65	0.45
Nb	5.2	46.5	20	27.4	17.1	10.8	33.9	9.7
Ta	0.3	2	1.2	1.6	0.8	0.5	1.5	0.4
Nd	10.5	58.8	29.4	43	26.1	18.9	42.5	16.4
Pb		4		5		3	4	4
Pr	2.05	13.1	6.54	9.01	5.55	3.8	9.42	3.43
Rb	0.3	3.5	0.2	24.2	2.4	13.3	18.2	10.6
Sc	44	21	26	23	38	42	34	42
Sm	3.19	12.9	7.86	9.65	5.8	4.94	9.58	4.58
Sr	100.5	142	201	306	496	181	353	200
Tb	0.63	1.57	1.07	1.29	0.88	0.92	1.27	0.87
Th	0.68	3.69	1.49	2.46	1.06	1.18	1.81	1.06
Tm	0.43	0.56	0.37	0.46	0.38	0.49	0.58	0.47
U	0.11	1.29	0.39	0.6	0.26	0.6	0.41	0.3
V	425	81	342	379	335	384	361	363
Y	23.6	38.9	26.1	31.1	25.1	29.8	38.2	27.3
Yb	2.62	3.45	2	2.62	2.44	3.09	3.82	2.81
Zr	73	338	204	257	154	122	249	113
Cr	80		100	20	90	90	120	100
Co	45	36	47	31	53	43	40	46
Cu	28	103	125	10	43	57	17	36
Ni	59	8	73	36	43	48	49	54
Zn	92	91	126	69	135	130	135	104

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 262 73	BLS 261 73	BLS 260 73	BLS 259 73	BLS 258 73	BLS 257 73	BLS 256 73
Location	N3-06	N3-07	N3-08	N3-09	N3-10	N3-11	N3-12
Traverse	East	East	East	East	East	East	East
Group	Enriched	Enriched	Enriched	Enriched	Enriched	Enriched	Enriched
Easting	519296	519338	519372	519411	519383	519289	519178
Northing	6817537	6817727	6817890	6818067	6818320	6818490	6818712
SiO2	47.3	46.1	46.9	44.7	45	42.6	46.9
Al2O3	13.3	13.3	13	12.5	13.05	12.8	13.8
Fe2O3	14.3	13.6	14.65	14.65	14	12.15	12.35
CaO	11.1	10.9	9.37	9.01	10.25	10.1	11.5
MgO	6.2	5.93	6.13	5.9	5.74	5.31	7.31
Na2O	2.72	2.52	3.05	3.2	2.08	2.74	2.03
K2O	0.25	0.32	0.27	0.37	0.24	0.5	0.02
TiO2	1.81	1.68	2.07	2.41	2.25	2.25	1.46
MnO	0.21	0.19	0.21	0.2	0.19	0.16	0.19
P2O5	0.13	0.14	0.19	0.24	0.23	0.26	0.13
LOI	1.6	5.26	2.98	5.44	5.27	NSS	3.04
Total	98.97	100	98.88	98.68	98.37	88.95	98.81
Mg#	27.21	27.32	26.51	25.77	26.12	27.37	33.79
Ba	57.2	169.5	111.5	178	127.5	271	27.6
Ce	22.6	25	33.2	52.5	36.7	51.2	20.3
Dy	4.58	4.28	5.07	5.42	5.35	5.22	3.79
Er	2.63	2.48	2.49	3.08	2.94	2.6	2.07
Eu	1.35	1.17	1.47	1.8	1.53	1.81	1.17
Ga	20.7	19.6	22.3	23	24.1	21.3	21.2
Gd	4.48	4.68	5.42	6.09	6.25	6.06	3.94
Hf	2.7	2.7	3.5	4.3	3.7	3.9	2.4
Ho	0.97	0.86	1.02	1.15	1.13	1.07	0.76
La	8.7	10.1	14.1	16.6	15.4	22.8	8.3
Lu	0.36	0.33	0.41	0.42	0.44	0.37	0.28
Nb	11.8	10.2	15.1	18.2	16.8	29.2	8.5
Ta	0.5	0.5	0.7	0.9	0.8	1.3	0.4
Nd	16.3	16.8	21.6	24.9	23.2	29.5	14.5
Pb		3	2	3	2	3	5
Pr	3.22	3.41	4.57	5.22	4.99	6.54	2.81
Rb	2.3	9.3	3.9	8.8	5.4	9.3	
Sc	44	39	37	37	36	33	35
Sm	4.25	4.3	5.41	6.1	5.94	6.67	3.71
Sr	191	174	272	231	323	270	423
Tb	0.76	0.71	0.83	0.99	0.97	0.94	0.64
Th	0.88	1.09	1.33	1.61	1.56	2.49	0.78
Tm	0.37	0.36	0.43	0.46	0.43	0.38	0.29
U	0.25	0.25	0.32	0.42	0.44	0.62	0.19
V	375	327	359	383	379	326	303
Y	23.7	21.9	25.5	28.7	27.5	25.6	19.2
Yb	2.31	2.13	2.36	2.68	2.7	2.49	1.97
Zr	100	102	128	163	147	158	92
Cr	100	100	140	90	110	130	190
Co	46	47	42	45	44	43	46
Cu	68	83	48	60	75	69	79
Ni	61	56	64	48	58	73	82
Zn	79	113	115	121	111	99	89

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 255 73	BLS 254 73	BL3 453	BLS 252 73	BLS 251 73	BLS 250 73	BLS 249 73
Location	N3-13	N3-14	N3-16	N3-19	N3-20	N3-21	N3-23
Traverse	East	East	East	East	East	East	East
Group	Depleted	Enriched	Enriched	Enriched	Enriched	Enriched	Enriched
Easting	519071	518914	518828	518501	518488	518481	518474
Northing	6818885	6819159	6819319	6819908	6820119	6820241	6820442
SiO2	47.7	45	48.1	50.7	46	48.4	47.1
Al2O3	14.55	13.4	13	15	14	13.35	13.95
Fe2O3	12.05	17.35	16.05	9.09	12.95	12.75	12.7
CaO	11	9.73	9.42	10.85	12.6	11.4	11.1
MgO	6.48	5.58	4.24	5.84	5.82	5.76	6.34
Na2O	2.33	2.33	3.64	3.82	2.89	3.23	3.01
K2O	0.03	0.42	0.91	0.55	0.33	0.38	0.57
TiO2	1.14	2.83	2.96	3.14	1.72	1.63	2.3
MnO	0.17	0.24	0.3	0.13	0.15	0.14	0.21
P2O5	0.11	0.29	0.78	0.52	0.16	0.14	0.18
LOI	3.41	2.89	0.98	1.85	2.3	2.99	1.69
Total	99.02	100.09	100.43	101.53	98.97	100.2	99.22
Mg#	31.68	21.71	18.55	35.65	27.93	28.03	30.09
Ba	18.8	96.2	240	112.5	29.1	33.2	170.5
Ce	16.9	44.5	94.6	78.1	37.3	21.6	30.8
Dy	3.73	7.69	10.15	7.77	4.79	5.56	5.02
Er	2.24	4.64	5.41	3.99	2.54	3.27	2.82
Eu	1.03	2.16	3.29	2.97	1.32	1.35	1.62
Ga	18.9	28	28.9	27.6	24.5	18.9	25.6
Gd	3.65	8.15	11.9	8.52	4.62	4.84	5.14
Hf	2	4.8	8.3	5.5	3.7	3.5	3.6
Ho	0.81	1.65	2.02	1.55	0.9	1.16	1.12
La	7.1	18.2	40.3	32.2	16.9	8.2	12.9
Lu	0.34	0.65	0.72	0.52	0.39	0.46	0.41
Nb	5.3	15.3	49.2	34.5	13	11.8	17.4
Ta	0.2	0.7	2.5	1.9	0.8	0.7	0.9
Nd	10.9	30.5	57.7	42.5	19.5	16.1	19.5
Pb	4			3	5	4	6
Pr	2.34	6.13	12.75	10.1	4.82	3.35	4.29
Rb	0.2	9.8	42	16	5.4	6.7	12.3
Sc	36	43	32	35	34	41	38
Sm	3.37	7.71	12.9	9.63	4.95	5.16	5.55
Sr	230	157.5	211	144.5	234	157.5	231
Tb	0.63	1.31	1.78	1.28	0.77	0.87	0.87
Th	0.71	2.07	3.96	1.97	1.92	1.38	1.27
Tm	0.33	0.63	0.76	0.58	0.38	0.49	0.41
U	0.17	0.5	1.08	2.39	0.51	0.38	0.27
V	300	513	182	437	330	395	459
Y	19.7	40.6	50.4	36.8	23.9	28.4	25.1
Yb	2.09	4.03	4.94	3.62	2.58	3.06	2.54
Zr	77	186	345	229	144	121	135
Cr	120	30	20	90	150	100	120
Co	44	48	29	25	36	39	38
Cu	75	44	15	61	34	94	8
Ni	45	20	16	22	55	44	73
Zn	84	90	108	51	51	35	86

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 245 73	BLS 244 73	BLS 243 73	BLS 248 73	BLS 247 73	BLS 246 73	BLS 185 73	BLS 184 73
Location	N3-25	N3-26	N3-27	N3-29	N3-31	N3-32	C1-43	C1-42
Traverse	East	East	East	East	East	East	Central	Central
Group	Enriched	Enriched	Enriched	Enriched	Enriched	Enriched	Depleted	Depleted
Easting	518460	518446	518436	518581	518762	518762	461122	461278
Northing	6820653	6820886	6821111	6821516	6821919	6822078	6797933	6798245
SiO2	45.7	48.1	41.6	45.4	42	43.6	46.2	46.6
Al2O3	12.45	12.75	13.4	13.4	15.45	15.25	10.25	11.8
Fe2O3	14.1	14.65	12.65	16.75	14.3	13.7	12.4	12.05
CaO	8.47	10.95	7.7	10.8	5.46	7.44	9.54	8.76
MgO	7.09	5.89	7.34	6.02	10.35	7.56	13.95	12.35
Na2O	3.04	2.83	2.63	1.67	2.58	2.58	0.26	0.97
K2O	0.18	0.2	0.03	1.12	1.02	1.23	1.68	1.78
TiO2	2.2	1.93	2.02	2.87	2.12	2.48	0.79	0.74
MnO	0.16	0.21	0.16	0.28	0.17	0.16	0.2	0.19
P2O5	0.19	0.16	0.33	0.29	0.36	0.45	0.06	0.06
LOI	5.48	2.83	10.35	2.21	4.31	3.51	3.61	3.79
Total	99.08	100.58	98.25	100.86	98.18	98.03	99.2	99.27
Mg#	30.24	25.74	33.35	23.66	38.42	32.24	49.24	46.91
Ba	42.8	92	14.6	195.5	256	342	584	459
Ce	24.6	26.7	42.5	39.3	44.7	55.2	6.8	6.4
Dy	5.3	5.27	4.54	7.56	5.25	6.33	2.84	2.72
Er	2.99	2.95	2.46	4.26	2.67	3.21	1.53	1.8
Eu	1.39	1.54	1.6	1.89	1.67	2.18	0.71	0.61
Ga	23.5	25.3	20.6	26.7	23.3	24.4	13	13
Gd	5.41	5.07	4.83	7.34	5.31	6.8	2.24	2.36
Hf	3.6	3.2	3.8	5.3	4	4.6	1.1	1.2
Ho	1.11	1.11	0.89	1.62	1.01	1.2	0.59	0.64
La	9.8	11	18.7	16.1	19.4	23.4	2.6	3.5
Lu	0.41	0.43	0.35	0.64	0.37	0.49	0.22	0.25
Nb	12.8	10.8	18.6	17.7	20.9	25.6	2.1	2.2
Ta	0.7	0.6	1.1	1.1	1.1	1.3	0.1	0.1
Nd	18.1	16.9	23.6	24.9	24.1	32.4	5.3	5.5
Pb	2	5	4	2	2	4	2	
Pr	3.55	3.76	5.49	5.42	5.8	7.24	1.02	1.06
Rb	4.6	3.5	0.5	37.7	32.6	29.6	37.8	40.4
Sc	38	39	25	42	19	26	34	38
Sm	5.03	5.11	5.59	7.04	5.77	7.54	1.88	1.65
Sr	101.5	503	152	105	73.2	93.7	125.5	84.8
Tb	0.92	0.87	0.77	1.19	0.82	1.01	0.41	0.43
Th	1.31	1.03	1.22	1.67	1.37	1.35	0.19	0.23
Tm	0.44	0.41	0.34	0.64	0.39	0.48	0.21	0.28
U	0.33	0.26	0.25	0.48	0.31	0.29		0.05
V	396	438	238	582	273	304	265	278
Y	27.1	26.4	21.5	39.5	25.3	31.1	14.1	15.2
Yb	2.83	2.8	2.16	4.17	2.53	3.11	1.5	1.53
Zr	136	114	152	186	168	193	41	40
Cr	50	100	120	90	140	140	1270	850
Co	44	46	48	40	48	43	64	56
Cu	3	70	42	22	5	8	20	66
Ni	49	58	104	43	84	79	440	266
Zn	115	107	106	181	99	57	83	83

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS-183-73	BLS 182 73	BLS 181 73	BLS 180 73	BLS 179 73	BLS 178 73	BLS 177 73
Location	C1-41	C1-40	C1-38	C1-36	C1-34	C1-33	C1-31
Traverse	Central	Central	Central	Central	Central	Central	Central
Group	Enriched	Enriched	Enriched	Enriched	Enriched	Enriched	Enriched
Easting	462536	462608	462814	462892	462775	462707	462585
Northing	6801174	6801726	6802125	6802515	6803016	6803362	6804429
SiO2	46.4	40.5	40.2	43.6	46.3	27.1	48.4
Al2O3	13.95	14.4	15.25	14.75	14.45	9.1	14.35
Fe2O3	13.3	13.6	16.35	13.25	12.5	7.47	11.5
CaO	9.01	8.51	10.7	11.5	7.18	24	10.45
MgO	8.06	3.69	5.03	6.82	7.08	4.86	7.07
Na2O	2.99	4.76	2.35	2.14	4.46	0.77	3.86
K2O	0.17	0.53	0.73	0.21	0.35	1.37	0.02
TiO2	1.84	3.07	3.67	2.37	2.09	1.57	0.88
MnO	0.25	0.34	0.25	0.18	0.17	0.16	0.19
P2O5	0.12	0.42	0.43	0.28	0.29	0.33	0.07
LOI	2.94	8.31	3	3.33	4.07	21.8	2.65
Total	99.12	98.18	98.09	98.53	99.05	98.64	99.46
Mg#	34.32	18.96	20.96	30.74	32.81	35.94	34.64
Ba	262	213	362	132	227	347	95.6
Ce	23.6	50.6	54.7	45.1	37.7	57.9	12.4
Dy	3.41	4.93	4.47	5.15	4.33	3.24	3.22
Er	2.09	2.36	2.14	2.87	2.49	1.68	1.91
Eu	1.25	2.5	2.28	1.95	1.91	1.74	1.01
Ga	18.6	20.9	25.9	22.5	19.5	11	17
Gd	3.91	6.29	5.63	5.64	5.14	4.49	2.8
Hf	2.1	4	3.7	3.7	2.9	2.4	1.7
Ho	0.7	0.93	0.86	1.01	0.93	0.59	0.73
La	10.3	19.8	23.9	19.1	16.1	28.2	5.2
Lu	0.26	0.31	0.26	0.38	0.32	0.21	0.28
Nb	8.7	31.3	25	19.8	14.9	29.9	4.9
Ta	0.5	1	1	0.9	0.8	1.7	0.3
Nd	15	32.3	30.7	26.2	22.1	27.8	7.8
Pb	2	2		2	3	4	
Pr	3.19	7.09	7.02	5.92	5.16	6.81	1.7
Rb	2.3	7.5	7.8	4.2	5.4	16.9	0.5
Sc	39	19	25	32	32	16	36
Sm	3.76	7.27	6.71	6.07	5.49	5.45	2.2
Sr	368	244	730	492	371	222	94.3
Tb	0.63	0.9	0.84	0.85	0.79	0.62	0.51
Th	0.7	2.45	2.22	1.58	0.97	2.66	0.63
Tm	0.29	0.33	0.29	0.39	0.36	0.21	0.29
U	0.15	0.74	0.5	0.37	0.21	1.58	0.18
V	333	245	432	247	255	140	268
Y	18	23.8	20.4	24.4	22	14.9	16.8
Yb	1.63	2.14	1.79	2.56	2.19	1.3	1.83
Zr	72	159	144	134	106	98	58
Cr	140		30	240	290	270	30
Co	52	32	55	46	43	27	41
Cu	117	34	70	104	80	39	123
Ni	90	16	41	111	80	93	75
Zn	99	115	106	97	87	54	91

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 176 73	BLS 175 73	BLS 174 73	BLS 51 73	BLS 52 73	BLS 53 73	BLS 54 73	BLS 55 73
Location	C1-30	C1-29	C1-25	K1-34	K1-33	K1-32	K1-31	K1-30
Traverse	Central	Central	Central	West	West	West	West	West
Group	Enriched	Depleted	Depleted	Enriched	Enriched	Enriched	Depleted	Depleted
Easting	462536	462512	462345	403267	403048	402874	402334	402205
Northing	6804639	6804741	6805533	6779049	6779195	6779219	6779317	6779411
SiO2	22.2	49	49.1	48.7	45	46.6	48.4	48.5
Al2O3	4.31	14.1	14.1	12.65	12.3	12.5	13.25	13.1
Fe2O3	11.55	11.65	11.7	15.2	16.7	15.7	13.15	14.85
CaO	25.8	7.79	6.96	8.18	10.25	10.05	11.95	10.45
MgO	5.3	7.67	7.6	5.57	5.46	6.21	6.81	6.5
Na2O	1.16	4.2	4.28	2.64	2.85	2.34	2.14	2.56
K2O	1.2	0.31	0.64	0.64	0.15	0.23	0.16	0.14
TiO2	3.64	0.96	0.93	2.17	2.25	1.77	1.13	1.37
MnO	0.17	0.18	0.17	0.22	0.26	0.23	0.21	0.21
P2O5	0.24	0.09	0.07	0.28	0.23	0.16	0.09	0.12
LOI	23.5	3.1	3.41	2.72	2.57	2.38	2.53	2.06
Total	99.47	99.47	99.16	99.05	98.08	98.24	99.85	99.9
Mg#	28.35	36.21	35.90	24.01	21.99	25.43	30.87	27.40
Ba	2340	3520	1420	355	41.3	58.7	39.6	59.4
Ce	133.5	14.2	11.6	39.9	30.9	22.5	14.1	14.4
Dy	4.42	3.45	3.19	6.78	7.02	5.46	4.13	4.18
Er	1.63	2.21	1.86	3.94	3.85	3.36	2.34	2.78
Eu	2.89	0.76	0.74	1.62	1.8	1.6	1.08	1.24
Ga	12.7	16.3	16.4	19.4	22.3	21.4	16.5	21.5
Gd	7.27	3.1	2.74	6.95	7	5.43	3.85	4.27
Hf	8.4	1.9	1.5	4.4	4.4	3.3	1.9	2.3
Ho	0.76	0.75	0.67	1.45	1.45	1.19	0.89	0.97
La	60.4	6	5.1	17	12.7	9.4	5.7	5.8
Lu	0.15	0.32	0.31	0.53	0.51	0.45	0.3	0.42
Nb	78.3	5.5	4.1	15.6	12.1	8.8	3.5	4.9
Ta	5	0.3	0.2	1.1	0.8	0.5	0.3	0.3
Nd	64.2	9.2	7.5	25.5	21.2	15.9	10.1	11
Pb	5	4	2					
Pr	16.3	1.91	1.64	5.73	4.65	3.36	2.18	2.22
Rb	43.4	2.2	8.4	13.1	0.9	3.6	3.2	2.9
Sc	21	39	38	40	38	39	43	40
Sm	10.65	2.62	2.4	6.78	6.08	4.73	3.09	3.56
Sr	538	240	116	198.5	338	321	159	194
Tb	0.92	0.53	0.53	1.14	1.09	0.86	0.62	0.75
Th	6.19	0.8	0.88	1.52	1.16	0.86	0.47	0.51
Tm	0.19	0.33	0.31	0.6	0.6	0.49	0.36	0.4
U	1.5	0.2	0.24	0.43	0.32	0.24	0.12	0.15
V	235	299	294	392	486	458	345	420
Y	17.1	19	16.5	34.3	33.5	29.7	20.3	25.3
Yb	1.08	2.14	1.8	3.83	3.5	2.96	2.22	2.51
Zr	325	65	56	175	167	118	71	85
Cr	550	190	150	80	40	140	60	70
Co	58	45	44	46	46	49	54	51
Cu	131	98	107	62	11	117	157	148
Ni	477	93	84	49	46	71	74	62
Zn	76	72	79	129	145	136	84	93

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 56 73	BLS 57 73	BLS 58 73	BLS 59 73	BLS 60 73	BLS 50 73	BLS 49 73	BLS 48 73
Location	K1-29	K1-28A	K1-28	K1-27	K1-26	K1-25	K1-24	K1-23
Traverse	West	West	West	West	West	West	West	West
Group	Depleted	Depleted	Depleted	Depleted	Depleted	Depleted	Enriched	Depleted
Easting	402042	401889	401739	401593	401415	401094	400938	400638
Northing	6779601	6779706	6779810	6779869	6779935	6780057	6780113	6780270
SiO2	46.7	47.4	47.8	45.8	46.8	47.3	47.5	48.4
Al2O3	13.05	13.5	12.25	12.95	12.15	12.65	12.65	13.4
Fe2O3	11.7	13.45	15	12.8	13.85	14.9	15.55	15.2
CaO	12.4	12.3	11.15	14.55	11.7	11.1	8.96	8.78
MgO	5.03	6.3	6.7	5.67	5.61	6.15	6.88	6.66
Na2O	1.74	1.23	1.75	2.01	1.84	1.49	2.2	3.43
K2O	0.06	0.02	0.25	0.1	0.07	0.14	0.32	0.08
TiO2	1.14	1.03	1.52	1.18	1.29	1.42	1.68	1.8
MnO	0.18	0.22	0.24	0.21	0.21	0.22	0.23	0.22
P2O5	0.09	0.07	0.11	0.09	0.1	0.12	0.14	0.21
LOI	6.47	4.12	2.46	3.56	5.27	2.95	3.9	3.75
Total	98.59	99.68	99.29	98.97	98.94	98.49	100.06	101.97
Mg#	27.04	28.77	27.80	27.64	25.88	26.25	27.61	27.42
Ba	26.6	5.3	80.5	22.8	28.9	41	79.4	18.2
Ce	10.5	10.4	14.9	12.2	12.2	13.9	18.2	27.2
Dy	3.94	3.79	4.85	4.01	4.84	4.72	4.88	4.29
Er	2.34	2.25	2.94	2.71	3.02	2.96	2.95	2.52
Eu	0.98	0.93	1.26	1.05	1.11	1.16	1.46	1.38
Ga	16.9	16.5	17.8	19.1	17.8	19.5	19.2	19.3
Gd	3.68	3.4	4.79	3.56	4.03	4.43	4.58	4.7
Hf	1.8	1.8	2.5	2.1	2.2	2.5	2.9	3.3
Ho	0.89	0.87	1.05	0.87	1.04	1.08	1.1	0.9
La	4.2	4.2	5.9	5	4.7	5.3	7.5	11.1
Lu	0.36	0.31	0.43	0.36	0.43	0.43	0.44	0.3
Nb	3.2	3.1	4.7	4.2	4	5.3	6.6	10.9
Ta	0.3	0.2	0.3	0.2	0.3	0.3	0.4	0.7
Nd	8.1	7.9	11.1	9	9.5	10	13.4	17.9
Pb				2				
Pr	1.66	1.59	2.31	1.91	1.93	2.1	2.88	4.01
Rb	1.4		4.9	1.4	1.5	3.5	9.4	0.4
Sc	43	41	42	40	41	43	42	36
Sm	2.6	2.67	3.64	2.8	3.27	3.3	3.9	4.62
Sr	162	222	209	237	183	314	277	168.5
Tb	0.58	0.57	0.77	0.64	0.68	0.74	0.81	0.75
Th	0.35	0.36	0.46	0.32	0.38	0.41	0.56	0.79
Tm	0.37	0.36	0.44	0.39	0.47	0.46	0.44	0.33
U	0.49	0.13	0.15	0.07	0.11	0.12	0.17	0.22
V	347	329	431	373	399	422	427	368
Y	20.1	20.2	25	22.1	24.4	25.9	26.9	23.1
Yb	2.19	2.43	2.92	2.23	2.94	2.93	2.75	2.14
Zr	62	63	91	66	78	80	99	122
Cr	90	80	160	140	110	140	80	160
Co	48	53	54	45	47	49	53	49
Cu	160	164	119	131	181	136	164	146
Ni	69	74	104	78	67	76	63	79
Zn	88	100	109	87	95	106	116	122

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 47 73	BLS 46 73	BLS 45 73	BLS 44 73	BLS 43 73	BLS 42 73	BLS 41 73	BLS 40 73
Location	K1-22	K1-20	K1-19	K1-18	K1-17	K1-16	K1-14	K1-13
Traverse	West	West	West	West	West	West	West	West
Group	Enriched	Enriched	Enriched	Enriched	Enriched	Enriched	Depleted	Enriched
Easting	400429	399586	399429	399230	397531	397356	397143	397013
Northing	6780395	6780796	6780858	6780931	6780111	6780273	6780486	6780699
SiO2	49.5	46.4	46.5	47.7	48.8	48.5	45.8	47
Al2O3	12.75	12.15	13.5	13.3	13.5	13.3	13.8	13.5
Fe2O3	15.15	16.05	13.35	15.55	15.45	16.2	16.65	16.3
CaO	9.44	10	15.45	9.63	8.48	8.72	9.3	9.88
MgO	6.18	6.14	5.83	6.88	5.33	5.82	6.72	5.57
Na2O	3.27	2.35	1.23	2.64	3.93	3.37	2.47	3.3
K2O	0.24	0.33	0.11	0.13	0.22	0.34	0.62	0.22
TiO2	1.56	2.07	1.32	1.46	1.98	1.81	2.69	2.51
MnO	0.22	0.23	0.24	0.24	0.21	0.23	0.24	0.24
P2O5	0.15	0.17	0.11	0.11	0.18	0.19	0.3	0.22
LOI	2.98	3.08	3.22	3.21	3.04	2.8	3.17	3.03
Total	101.51	99.02	100.88	100.88	101.15	101.33	101.82	101.81
Mg#	26.02	24.80	27.35	27.61	22.93	23.65	25.81	22.76
Ba	75.8	77	18.7	32.2	68.4	115.5	207	46.3
Ce	26.4	25.1	16.8	16	24.3	24.4	40.6	32
Dy	4.6	5.58	3.75	4.01	5.08	5.14	5.79	5.4
Er	2.98	3.25	2.1	2.25	2.84	2.96	3.18	2.73
Eu	1.18	1.7	1.03	1.13	1.29	1.39	1.92	1.97
Ga	16.7	22.8	19.8	16.9	18.4	17.8	20.7	24.5
Gd	4.53	6.34	3.6	3.73	4.89	4.85	6.3	6.31
Hf	3.2	4.1	2.2	2.2	3.1	2.9	3.9	4.4
Ho	1.1	1.23	0.8	0.88	1.16	1.2	1.16	1.11
La	10.9	10.1	7	7.4	9.9	9.6	17.6	12.5
Lu	0.47	0.44	0.29	0.35	0.45	0.45	0.43	0.32
Nb	11.8	10.7	5.8	5.8	9.5	9.7	21.3	11.2
Ta	0.7	0.6	0.4	0.4	0.6	0.6	1.3	0.7
Nd	16.8	17.5	11.3	11.4	15.9	16.4	25.6	23
Pb			2					
Pr	3.81	3.77	2.43	2.5	3.49	3.46	5.7	4.75
Rb	4.5	9.2	1.1	2.1	2.8	6.1	11.8	2.4
Sc	42	38	37	44	43	44	36	33
Sm	4.33	5.02	3.16	3.22	4.54	4.36	6.2	6.28
Sr	323	261	87.4	178	100.5	225	147	113
Tb	0.76	1.04	0.62	0.61	0.83	0.8	1.03	0.95
Th	0.94	0.97	0.51	0.48	0.75	0.79	1.44	0.86
Tm	0.46	0.45	0.3	0.35	0.44	0.46	0.45	0.39
U	0.3	0.28	0.15	0.18	0.22	0.19	0.32	0.26
V	347	499	311	388	438	426	417	451
Y	25.1	29.4	19.1	21.1	26.7	27.6	29.3	26
Yb	2.88	2.73	1.98	2.09	2.65	2.9	2.78	2.32
Zr	119	138	78	73	110	111	149	153
Cr	150	100	50	70	90	90	110	110
Co	49	55	47	55	50	51	52	53
Cu	169	119	87	130	159	151	46	126
Ni	68	60	55	64	63	63	70	81
Zn	128	129	107	120	125	129	129	132

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 39 73	BLS 38 73	BLS 37 73	BLS 35 73	BL 73 105	BLS 34 73	BLS 33 73	BLS 32 73
Location	K1-12	K1-11	K1-10	K1-08	K1-08	K1-07	K1-06	K1-05
Traverse	West	West	West	West	West	West	West	West
Group	Enriched	Depleted	Depleted	Depleted	Depleted	Depleted	Depleted	Depleted
Easting	396871	396781	396697	396503	396503	396393	396147	395979
Northing	6780809	6780951	6781080	6781364	6781364	6781526	6781655	6781713
SiO2	47.4	46.8	48.1	46.4	45.8	48	47.4	48.4
Al2O3	13.55	13.3	13.6	13.2	14.2	13.3	13.5	14.35
Fe2O3	13.95	14.45	14.75	12.55	11.55	14.65	15.05	13.55
CaO	13	13.15	11.15	11.3	13	10.5	9.74	9.6
MgO	5.74	6.07	6.99	9.29	8.51	7.02	6.64	6.82
Na2O	0.89	1.24	1.76	1.45	0.8	2.44	2.83	3.3
K2O	0.02	0.07	0.13	0.61	0.78	0.12	0.11	0.11
TiO2	1.35	1.32	1.45	1.11	1.09	1.52	1.76	1.31
MnO	0.2	0.2	0.2	0.17	0.17	0.2	0.21	0.18
P2O5	0.1	0.11	0.11	0.08	0.09	0.11	0.13	0.1
LOI	4.89	4.31	3	3.45	3.51	2.84	2.91	2.87
Total	101.12	101.05	101.29	99.68	99.55	100.76	100.3	100.63
Mg#	26.19	26.59	29.01	38.96	38.85	29.24	27.56	30.26
Ba	9.3	14.7	31.4	107	134.5	22.2	23.4	32.5
Ce	15.5	12.8	15.3	11.4	12.9	16.1	20.1	12.5
Dy	4.87	3.99	4.44	3.06	3.26	4.31	5.01	3.83
Er	2.91	2.26	2.73	2.01	2.33	2.25	3.1	2.39
Eu	1.31	1.07	1.16	0.89	0.73	1.38	1.43	1.11
Ga	25.3	20.7	19	18.2	20.7	18.8	19.3	17.1
Gd	4.19	3.8	4.01	3.25	3.43	4.24	5.18	3.78
Hf	2.2	2.2	2.4	1.8	2.1	2.6	2.9	2.1
Ho	1.02	0.89	0.97	0.71	0.74	0.89	1.2	0.88
La	6	5.1	6	4.8	5.6	6.3	8.1	4.7
Lu	0.41	0.36	0.38	0.29	0.29	0.32	0.41	0.32
Nb	5.1	4.7	5.4	4.2	4.3	5.5	7.4	5
Ta	0.3	0.3	0.4	0.2	0.3	0.3	0.5	0.3
Nd	11.2	9.8	11.5	8.4	9.6	12.3	15	9.7
Pb	2		3					
Pr	2.29	1.98	2.35	1.74	1.93	2.5	3.1	2.04
Rb	0.3	0.6	1.9	16.5	20.2	1.4	1.2	1.3
Sc	38	40	43	38	38	40	44	45
Sm	3.49	3.02	3.54	2.67	2.94	3.54	4.43	3.22
Sr	81.8	66.5	261	124.5	79.4	325	37.6	84.5
Tb	0.73	0.64	0.67	0.52	0.57	0.67	0.83	0.63
Th	0.44	0.46	0.47	0.37	0.38	0.41	0.55	0.37
Tm	0.42	0.34	0.4	0.26	0.31	0.35	0.43	0.32
U	0.1	0.14	0.15	0.11	0.12	0.12	0.17	0.13
V	434	374	398	351	347	399	429	335
Y	25.5	21.7	23.9	17.4	19	21	27.2	20.1
Yb	2.65	2.18	2.48	1.88	1.97	1.94	2.85	2.1
Zr	78	75	84	61	69	85	99	72
Cr	130	100	110	300	230	100	160	190
Co	42	50	55	51	44	57	53	63
Cu	83	148	216	95	67	156	119	137
Ni	65	80	77	121	110	72	76	117
Zn	98	109	113	93	76	103	119	99

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Sample	BLS 31 73	BL 73 40	BLS 30 73
Location	K1-04	K1-04	K1-03
Traverse	West	West	West
Group	Enriched	Enriched	Enriched
Easting	395889	395889	395539
Northing	6781752	6781752	6781887
SiO2	44.8	44.7	47.8
Al2O3	12.85	11.7	13.9
Fe2O3	16.85	15.55	14.15
CaO	8.23	12.3	8.01
MgO	5.34	5.75	6.25
Na2O	2.83	1.6	3.34
K2O	0.57	0.56	0.17
TiO2	3.03	3.27	2.04
MnO	0.24	0.23	0.17
P2O5	0.37	0.28	0.25
LOI	3.01	3.31	4.66
Total	98.18	99.32	100.81
Mg#	21.46	24.17	27.58
Ba	199.5	224	95.9
Ce	50.5	44.7	38.8
Dy	5.93	5.46	4.52
Er	3.26	3.31	2.52
Eu	2.55	2.23	1.89
Ga	25.8	21.9	20.2
Gd	7.23	6.43	5.48
Hf	5.4	4.7	4.3
Ho	1.3	1.15	1.02
La	20.9	18.8	16.8
Lu	0.45	0.42	0.35
Nb	22.8	20.9	19.7
Ta	1.3	1.2	1
Nd	31.9	28	23.4
Pb		6	
Pr	6.98	6.19	5.32
Rb	12.9	12	1.8
Sc	37	45	36
Sm	7.11	6.47	5.78
Sr	216	368	241
Tb	1.12	1.05	0.82
Th	1.79	1.5	1.78
Tm	0.46	0.41	0.36
U	0.42	0.37	0.41
V	548	733	340
Y	31.1	28.2	23.4
Yb	3	2.58	2.17
Zr	220	198	157
Cr	80	110	180
Co	52	50	48
Cu	75	145	43
Ni	32	54	67
Zn	139	125	115

Coordinates in NAD27, zone 18V

Major elements in wt. %

Trace elements in ppm

Appendix 7. Major element concentrations for samples of the Roberts Lake Syncline with re-analysed trace elements

Sample No.	JMPAR99-38	JMPAR99-38	JMPAR99-40	JMPAR99-46	JMPAR99-53	JMPAR99-54
description	basalt - pillow	basalt - pillow	basalt - pillow	basalt	basalt	basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	432192	432192	432044	431727	431069	430874
UTMY	6672064	6672064	6672044	6671995	6672183	6672197
SiO2	47.96	47.96	45.76	49.40	49.44	48.74
TiO2	1.42	1.42	1.41	2.12	1.94	1.16
Al2O3	14.17	14.17	14.20	13.97	13.93	14.79
Fe2O3	13.03	13.03	14.45	14.45	14.41	12.61
MnO	0.20	0.20	0.22	0.19	0.22	0.21
MgO	6.43	6.43	7.00	4.82	6.46	7.50
CaO	11.62	11.62	12.32	9.82	7.97	10.42
Na2O	2.75	2.75	1.77	3.07	4.05	2.92
K2O	0.16	0.16	0.05	0.16	0.14	0.19
P2O5	0.11	0.11	0.10	0.16	0.13	0.08
LOI	2.41	2.41	2.90	1.92	2.19	2.23
Total	100.27	100.27	100.17	100.10	100.88	100.85
Mg#	30	30	29	22	28	34

Coordinates in NAD83, zone 19V

Concentrations in wt. %

Mg# = molar MgO/Mgo+FeO^T*100

Sample No.	JMPAR99-61	JMPAR99-63	JMPAR99-67	JMPAR99-73	JMPAR99-77	JMPAR99-86
description	basalt	basalt	basalt	basalt - sheet	basalt - sheet	basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	429853	429648	429052	426393	426672	427596
UTMY	6671869	6671855	6672150	6671801	6671985	6672538
SiO2	49.16	45.72	47.63	49.59	46.33	47.90
TiO2	1.42	1.34	1.95	1.86	1.48	1.71
Al2O3	14.34	15.23	14.38	13.64	14.59	13.98
Fe2O3	13.93	14.10	15.37	14.41	14.53	15.02
MnO	0.25	0.22	0.23	0.24	0.22	0.21
MgO	6.19	7.48	6.35	6.01	7.26	6.11
CaO	8.56	12.80	9.30	9.26	12.33	10.27
Na2O	3.96	1.55	3.25	3.75	1.77	3.01
K2O	0.29	0.04	0.17	0.09	0.11	0.22
P2O5	0.10	0.09	0.14	0.14	0.10	0.13
LOI	2.74	2.39	2.16	1.76	2.04	1.96
Total	100.94	100.97	100.92	100.74	100.74	100.52
Mg#	28	31	26	26	30	26

Coordinates in NAD83, zone 19V

Concentrations in wt. %

Mg# = molar MgO/Mgo+FeO^T*100

Sample No.	JMPAR99-88	JMPAR99-32	JMPAR99-35	JMPAR99-37	JMPAR99-42	JMPAR99-58
description	basalt	basalt - sheet	basalt - sheet	basalt - pillow	basalt - pillow	grey basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	428131	432653	432428	432284	431880	430450
UTMY	6672200	6671940	6671992	6672089	6672014	6672061
SiO2	47.07	48.64	47.90	40.56	48.26	50.34
TiO2	1.22	2.68	2.31	2.72	2.51	2.41
Al2O3	14.77	12.56	13.56	14.14	13.03	13.64
Fe2O3	13.84	16.02	16.36	13.84	16.58	13.84
MnO	0.20	0.22	0.23	0.18	0.23	0.18
MgO	8.36	6.01	6.17	4.20	5.23	6.41
CaO	10.04	9.91	8.24	11.29	9.29	6.33
Na2O	2.43	1.93	2.94	4.28	1.57	4.70
K2O	0.29	0.04	0.20	0.14	0.21	0.42
P2O5	0.09	0.20	0.20	0.23	0.20	0.20
LOI	2.61	2.21	2.59	7.04	3.14	2.39
Total	100.91	100.44	100.71	98.61	100.24	100.85
Mg#	34	24	25	21	21	29

Coordinates in NAD83, zone 19V

Concentrations in wt. %

Mg# = molar MgO/Mgo+FeO^T*100

Sample No.	JMPAR99-66	JMPAR99-84	JMPAR99-85
description	basalt	basalt	basalt
Group	Group 1	Group 1	Group 1
UTMX	429241	427347	427432
UTMY	6672011	6672318	6672401
SiO2	53.03	47.47	49.39
TiO2	3.07	2.48	2.70
Al2O3	10.70	13.07	13.12
Fe2O3	14.23	17.12	14.30
MnO	0.24	0.25	0.23
MgO	4.15	4.93	5.62
CaO	9.63	10.04	9.77
Na2O	2.68	2.12	2.48
K2O	0.32	0.56	0.21
P2O5	0.31	0.20	0.23
LOI	1.93	2.37	2.37
Total	100.29	100.61	100.42
Mg#	20	20	25

Coordinates in NAD83, zone 19V

Concentrations in wt. %

Mg# = molar MgO/Mgo+FeO^T*100

Sample No.	NQ0-13-02	NQ0-13-03	NQ0-13-06	NQ0-13-13	NQ0-13-14
description	aphyric basalt	aphyric basalt	aphyric basalt	aphyric basalt	aphyric basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	4444163	444318	444692	428647.44	428647.44
UTMY	6653329	6653397	6654412	6670241.66	6670241.66
SiO2	47.80	49.40	45.80	48.00	47.40
TiO2	1.39	2.20	1.52	2.05	1.82
Al2O3	13.60	12.25	14.05	11.90	12.50
Fe2O3	14.05	17.40	14.60	16.00	16.05
MnO	0.21	0.25	0.23	0.24	0.23
MgO	6.56	5.68	7.36	6.29	6.37
CaO	11.20	9.58	10.45	8.83	9.89
Na2O	2.22	0.97	2.34	3.21	2.80
K2O	0.07	0.08	0.05	0.15	0.24
P2O5	0.10	0.14	0.12	0.16	0.15
LOI	2.02	2.88	2.62	1.64	1.84
Total	99.27	100.85	99.18	98.50	99.35
Mg#	29	22	30	25	25

Coordinates in NAD83, zone 19V

Concentrations in wt. %

Mg# = molar MgO/Mgo+FeO^T*100

Sample No.	JMPAR99-16	JMPAR99-15	JMPAR99-17	JMPAR99-18	JMPAR99-19	JMPAR99-20
description	basalt	basalt	basalt	basalt	basalt	basalt
Group	Group 2	Group 2	Group 2	Group 2	Group 2	Group 2
UTMX	434418	434418	434418	434418	434418	434418
UTMY	6672550	6672550	6672550	6672550	6672550	6672550
SiO2	46.58	47.90	46.84	47.61	46.69	46.42
TiO2	0.67	0.75	0.63	0.73	0.74	0.69
Al2O3	10.37	11.37	8.68	11.02	11.10	10.38
Fe2O3	12.05	12.00	11.94	12.33	11.58	12.12
MnO	0.18	0.17	0.18	0.18	0.17	0.17
MgO	14.56	13.32	17.74	12.56	14.25	15.60
CaO	12.47	8.82	9.75	9.80	10.68	10.86
Na2O	0.66	2.38	1.07	2.36	1.39	0.76
K2O	0.03	0.03	0.02	0.02	0.02	0.11
P2O5	0.05	0.05	0.04	0.05	0.05	0.05
LOI	3.01	3.32	3.81	3.51	3.29	3.72
Total	100.63	100.11	100.71	100.17	99.97	100.89
Mg#	51	49	56	47	51	53

Coordinates in NAD83, zone 19V

Concentrations in wt. %

Mg# = molar MgO/Mgo+FeO^T*100

Sample No.	NQ0-13-07	NQ0-13-08	NQ0-13-09	NQ0-13-10	NQ0-13-11
description	Peridotite	Peridotite	Peridotite	Peridotite	Peridotite
Group	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq
UTMX	428647.44	428647.44	428647.44	428647.44	428647.44
UTMY	6670241.66	6670241.66	6670241.66	6670241.66	6670241.66
SiO2	38.20	37.40	38.20	38.00	39.50
TiO2	0.17	0.18	0.24	0.24	0.28
Al2O3	3.07	3.15	4.20	4.06	4.91
Fe2O3	11.80	12.30	12.60	12.85	12.30
MnO	0.18	0.18	0.18	0.18	0.19
MgO	34.10	33.40	30.50	30.90	28.80
CaO	1.56	1.71	4.34	3.88	4.03
Na2O	<0.01	<0.01	<0.01	<0.01	<0.01
K2O	0.01	0.01	0.01	0.01	<0.01
P2O5	0.02	0.01	0.02	0.02	0.02
LOI	10.05	10.00	8.76	8.86	8.91
Total	99.68	98.86	99.50	99.46	99.39
Mg#	71	70	68	67	67

Coordinates in NAD83, zone 19V

Concentrations in wt. %

Mg# = molar MgO/Mgo+FeO^T*100

Appendix 8. Trace element concentrations for samples of the Roberts Lake Syncline with re-analysed trace elements

Sample No.	JMPAR99-38	JMPAR99-38	JMPAR99-40	JMPAR99-46	JMPAR99-53	JMPAR99-54
description	basalt - pillow	basalt - pillow	basalt - pillow	basalt	basalt	basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	432192	432192	432044	431727	431069	430874
UTMY	6672064	6672064	6672044	6671995	6672183	6672197
Ba	116	116	45.6	71.8	84.1	47
Ce	14.3	14.3	13.1	22.5	17.5	10.3
Cr	170	170	160	90	80	370
Cs	0.03	0.03	<0.01	0.03	<0.01	0.04
Dy	4.16	4.16	3.77	5.62	5.52	3.55
Er	2.34	2.34	2.13	3.58	3.14	2.2
Eu	1.27	1.27	1.03	1.56	1.37	0.76
Ga	22.8	22.8	19.1	23.6	19.1	18.1
Gd	3.95	3.95	3.82	5.39	5.02	3.18
Hf	2.2	2.2	2.1	3.3	3	1.7
Ho	0.85	0.85	0.81	1.19	1.09	0.72
La	5.6	5.6	4.9	9	6.6	3.8
Lu	0.32	0.32	0.3	0.43	0.42	0.3
Nb	5.3	5.3	4.9	8	6.8	4
Nd	10.7	10.7	10	16	13.5	8.2
Pr	2.14	2.14	1.84	3.36	2.6	1.62
Rb	1.4	1.4	0.3	1.4	1.6	6
Sm	3.2	3.2	3.11	4.89	4.13	2.71
Sn	1	1	1	1	1	1
Sr	176	176	329	195	35	105
Ta	0.3	0.3	0.3	0.5	0.4	0.2
Tb	0.68	0.68	0.61	0.99	0.86	0.55
Th	0.49	0.49	0.41	0.67	0.53	0.27
Tm	0.33	0.33	0.32	0.48	0.41	0.31
U	0.15	0.15	0.11	0.19	0.52	0.07
V	371	371	342	474	436	328
W	1	1	1	1	1	1
Y	21.5	21.5	19.4	29.9	27.6	18.2
Yb	2.45	2.45	1.94	2.93	2.73	1.88
Zr	82	82	75	124	107	59
Ag	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
As	<5	<5	<5	5	<5	<5
Cd	<0.5	<0.5	<0.5	<0.5	0.5	<0.5
Co	48	48	48	45	54	48
Cu	84	84	88	106	110	89
Li	10	10	10	10	10	10
Mo	1	1	1	1	1	<1
Ni	81	81	83	59	76	100
Pb	95	95	11	15	19	6
Sc	43	43	42	41	42	44
Tl	<10	<10	<10	<10	<10	<10
Zn	96	96	104	115	131	91
Au	3	3	3	152	10	14
Pt	1.8	1.8	2	0.9	1.9	2.1
Pd	0.2	0.2	1.4	2.6	5.9	9.3

Coordinates in NAD83, zone 19V
Trace element concentration in ppm
Au, Pt, Pd concentrations in ppb

Sample No.	JMPAR99-61	JMPAR99-63	JMPAR99-67	JMPAR99-73	JMPAR99-77	JMPAR99-86
description	basalt	basalt	basalt	basalt - sheet	basalt - sheet	basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	429853	429648	429052	426393	426672	427596
UTMY	6671869	6671855	6672150	6671801	6671985	6672538
Ba	95.1	62	114	30.6	42.2	52.8
Ce	11.5	12.9	23.9	20.3	13.3	15.2
Cr	160	190	190	80	170	100
Cs	0.01	<0.01	0.02	<0.01	0.02	0.02
Dy	4.1	3.83	5.8	5.25	3.81	4.37
Er	2.35	2.38	3.46	3	2.41	2.43
Eu	1.04	1.05	1.57	1.42	1.11	1.11
Ga	18.7	21.2	26.3	18.8	21.8	18.4
Gd	3.59	3.88	5.7	5	3.71	3.9
Hf	2.1	2	3.1	2.7	2	2.4
Ho	0.86	0.86	1.17	1.07	0.79	0.87
La	4.1	4.9	9.7	7.9	5	5.9
Lu	0.33	0.29	0.44	0.36	0.31	0.34
Nb	4.6	4.6	8.4	6.7	4.7	6.3
Nd	9.8	10	16.1	15.2	10.4	11.3
Pr	1.77	1.87	3.33	3.02	1.99	2.26
Rb	4	0.5	1.7	0.7	1.5	2.3
Sm	2.89	2.86	4.53	4.34	3.07	3.55
Sn	1	1	1	1	2	1
Sr	135	210	126	44	200	70
Ta	0.3	0.3	0.5	0.4	0.2	0.4
Tb	0.61	0.65	0.93	0.88	0.62	0.7
Th	0.42	0.36	0.72	0.71	0.41	0.53
Tm	0.33	0.32	0.42	0.43	0.33	0.34
U	0.12	0.17	0.22	0.23	0.12	0.21
V	337	344	457	383	373	335
W	1	<1	2	1	1	2
Y	21.0	20.6	29.4	26.9	20.6	22.4
Yb	2.24	2.02	2.97	2.47	2	2.26
Zr	77	75	117	106	75	91
Ag	<0.5	<0.5	<0.5	<0.5	2.5	<0.5
As	<5	<5	9	7	<5	<5
Cd	<0.5	0.5	<0.5	<0.5	<0.5	0.5
Co	52	50	49	45	49	50
Cu	128	118	96	143	662	112
Li	20	10	10	10	10	10
Mo	<1	1	1	<1	<1	6
Ni	87	103	80	50	94	79
Pb	8	9	21		40	15
Sc	40	41	40	37	43	40
Tl	<10	<10	<10	<10	<10	<10
Zn	103	98	106	141	112	125
Au	7	29	4	9	87	2
Pt	1.3	1.2	1.4	1.8	2.6	1.6
Pd	1.5	1.3	<0.2	11	22	0.9

Coordinates in NAD83, zone 19V
Trace element concentration in ppm
Au, Pt, Pd concentrations in ppb

Sample No.	JMPAR99-88	JMPAR99-32	JMPAR99-35	JMPAR99-37	JMPAR99-42	JMPAR99-58
description	basalt	basalt - sheet	basalt - sheet	basalt - pillow	basalt - pillow	grey basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	428131	432653	432428	432284	431880	430450
UTMY	6672200	6671940	6671992	6672089	6672014	6672061
Ba	58.7	142.5	142.5	75.1	107	152
Ce	11.2	26.6	29.8	32.2	26.8	30.9
Cr	170	60	80	140	50	90
Cs	0.01	0.05	0.03	0.03	0.51	0.04
Dy	3.24	5.97	6.19	6.9	6.74	5.44
Er	1.94	3.71	3.52	3.85	4.01	3.01
Eu	0.79	2.05	1.6	1.61	1.89	1.86
Ga	15.2	26	21.8	22.3	27.1	21.9
Gd	3.01	6.17	6.03	6.66	6.62	5.83
Hf	1.7	3.9	4	4.3	3.8	4.5
Ho	0.66	1.23	1.23	1.35	1.42	1.11
La	4.2	10.6	12	12.5	10.3	11.7
Lu	0.26	0.49	0.43	0.53	0.52	0.35
Nb	3.8	12.9	12.7	13.4	10.3	15.2
Nd	8.1	19.3	20.1	22.1	19.4	21.2
Pr	1.63	3.9	4.2	4.52	3.93	4.44
Rb	4.3	0.3	2.2	1.5	4.5	7
Sm	2.79	5.12	5.73	5.94	5.5	5.65
Sn	1	2	1	2	2	2
Sr	77	293	218	64	421	37
Ta	0.2	0.8	0.7	0.7	0.6	0.9
Tb	0.52	1.07	1.01	1.12	1.08	0.98
Th	0.45	0.96	1.02	1.09	0.85	1.56
Tm	0.28	0.5	0.47	0.56	0.6	0.4
U	0.1	0.31	0.33	0.32	0.26	0.41
V	282	517	429	524	511	415
W	<1	1	1	1	1	<1
Y	17.0	32.0	31.7	32.5	34.6	26.2
Yb	1.73	3.13	3.11	3.37	3.23	2.52
Zr	61	151	146	168	147	176
Ag	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
As	<5	5	<5	<5	<5	16
Cd	<0.5	0.8	0.7	<0.5	0.5	<0.5
Co	51	40	47	46	43	59
Cu	57	61	79	103	104	16
Li	10	20	10	10	10	20
Mo	<1	1	<1	1	1	1
Ni	121	45	47	63	42	51
Pb	4	86	70	17	58	10
Sc	40	40	41	44	39	36
Tl	<10	<10	<10	<10	<10	<10
Zn	101	127	141	132	132	153
Au	<1	4	8	4	7	9
Pt	1.4	1.5	0.8	0.9	0.7	0.7
Pd	3.8	<0.2	<0.2	<0.2	1	0.7

Coordinates in NAD83, zone 19V

Trace element concentration in ppm

Au, Pt, Pd concentrations in ppb

Sample No.	JMPAR99-66	JMPAR99-84	JMPAR99-85
description	basalt	basalt	basalt
Group	Group 1	Group 1	Group 1
UTMX	429241	427347	427432
UTMY	6672011	6672318	6672401
Ba	202	170	45.4
Ce	33.9	29.7	30.7
Cr	100	60	100
Cs	0.42	0.91	0.02
Dy	8.35	6.46	6.73
Er	4.29	3.89	3.94
Eu	2.08	1.62	1.85
Ga	22.6	23.9	22.5
Gd	8.46	6.21	6.56
Hf	5.8	4	4.4
Ho	1.64	1.32	1.31
La	12.5	12	12.2
Lu	0.63	0.51	0.53
Nb	16.9	11.2	12.8
Nd	25.3	20	21.9
Pr	5.15	4.13	4.57
Rb	7.5	12.2	2.7
Sm	7.67	5.65	6.09
Sn	2	2	2
Sr	135	201	97
Ta	1.0	0.7	0.8
Tb	1.38	1.08	1.1
Th	1.39	1.18	1.16
Tm	0.65	0.52	0.53
U	0.4	0.35	0.6
V	428	480	481
W	<1	1	1
Y	40.8	33.6	33.9
Yb	3.89	3.31	3.3
Zr	222	150	168
Ag	<0.5	<0.5	<0.5
As	<5	<5	14
Cd	<0.5	0.6	<0.5
Co	46	48	47
Cu	71	126	35
Li		10	10
Mo	2	1	2
Ni	56	41	58
Pb	8	19	8
Sc	35	42	44
Tl	<10	<10	<10
Zn	120	140	143
Au	20	42	1
Pt	1.4	1.3	1.3
Pd	1.5	<0.2	<0.2

Coordinates in NAD83, zone 19V

Trace element concentration in ppm

Au, Pt, Pd concentrations in ppb

Sample No.	NQ0-13-02	NQ0-13-03	NQ0-13-06	NQ0-13-13	NQ0-13-14
description	aphyric basalt	aphyric basalt	aphyric basalt	aphyric basalt	aphyric basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	4444163	444318	444692	428647.44	428647.44
UTMY	6653329	6653397	6654412	6670241.66	6670241.66
Ba	24.8	24.8	23	92.1	102
Ce	14.8	19.5	15.8	23.1	21.1
Cr	120	20	100	60	110
Cs	<0.01	0.19	<0.01	0.04	0.08
Dy	4.2	5.46	4.84	5.73	5.32
Er	2.28	3.05	2.62	3.16	3.05
Eu	1.11	1.57	1.19	1.76	1.39
Ga	18.8	20.1	21.2	19.6	19.5
Gd	4.12	5.02	4.25	5.07	4.73
Hf	2.3	3	2.7	3.3	3
Ho	0.93	1.14	0.97	1.21	1.14
La	5.6	7.5	5.9	8.7	8.3
Lu	0.33	0.45	0.33	0.48	0.45
Nb	5.2	7.1	6	10	9.1
Nd	10.7	13.7	11.7	15.8	14.4
Pr	2.26	2.83	2.4	3.33	3.08
Rb	0.8	1.5	0.6	1.7	4.7
Sm	3.26	4.07	3.82	4.5	4.37
Sn	1	1	1	1	1
Sr	231	153	216	67	241
Ta	0.3	0.4	0.3	0.6	0.5
Tb	0.66	0.88	0.69	0.88	0.86
Th	0.42	0.57	0.42	0.76	0.7
Tm	0.35	0.45	0.39	0.49	0.44
U	0.14	0.2	0.14	0.25	0.19
V	398	613	387	473	418
W	<1	<1	<1	<1	<1
Y	21.9	27.4	23.9	29.4	27.8
Yb	2.29	3.05	2.48	3.23	3
Zr	78	105	93	116	107
Ag	<0.5	<0.5	<0.5	<0.5	<0.5
As	<5	<5	<5	<5	8
Cd	0.6	1.1	0.5	0.7	<0.5
Co	53	61	58	53	56
Cu	125	71	95	298	129
Li	<10	<10	<10	<10	10
Mo	<1	<1	<1	<1	<1
Ni	80	43	94	55	72
Pb	2	4	<2	5	6
Sc	45	48	44	49	49
Tl	<10	10	<10	<10	10
Zn	108	155	107	52	49

Coordinates in NAD83, zone 19V
Trace element concentration in ppm

Sample No.	JMPAR99-16	JMPAR99-15	JMPAR99-17	JMPAR99-18	JMPAR99-19	JMPAR99-20
description	basalt	basalt	basalt	basalt	basalt	basalt
Group	Group 2	Group 2	Group 2	Group 2	Group 2	Group 2
UTMX	434418	434418	434418	434418	434418	434418
UTMY	6672550	6672550	6672550	6672550	6672550	6672550
Ba	12.6	12.8	6.6	7.3	189	20.5
Ce	5.7	5.1	4.9	5.5	6.5	5.4
Cr	1340	1500	1810	1400	1450	1570
Cs	0.07	0.1	0.14	0.05	0.06	0.08
Dy	2.55	2.77	2.26	2.58	2.82	2.29
Er	1.44	1.63	1.23	1.54	1.72	1.48
Eu	0.62	0.58	0.47	0.59	0.71	0.56
Ga	16	11.9	8.8	12.3	17.4	13.7
Gd	2.2	2.35	1.98	2.25	2.34	2.02
Hf	1	1.1	0.9	1.2	1.1	1
Ho	0.54	0.55	0.47	0.56	0.63	0.53
La	2.1	1.9	2.1	2.1	2.5	2.1
Lu	0.21	0.25	0.18	0.22	0.23	0.2
Nb	1.7	1.8	1.4	1.7	1.9	1.5
Nd	4.3	4.8	4.2	5.2	5.1	4.7
Pr	0.85	0.85	0.77	0.87	0.99	0.84
Rb	1.1	0.5	0.3	0.2	0.5	2.3
Sm	1.54	1.7	1.44	1.72	1.69	1.76
Sn		1	<1	1	1	1
Sr	124.5	105	28.7	98.9	84.2	46.2
Ta	0.1	0.2	0.2	0.2	0.1	0.2
Tb	0.38	0.46	0.34	0.4	0.44	0.36
Th	0.21	0.21	0.15	0.18	0.19	0.14
Tm	0.22	0.27	0.19	0.26	0.26	0.21
U	0.05	0.07	<0.05	0.06	0.13	0.05
V	249	292	229	279	287	257
W		<1	<1	1		<1
Y	13.3	16	12.9	15.1	15.3	14.1
Yb	1.46	1.6	1.33	1.43	1.66	1.33
Zr	37	42	34	40	42	36
Ag		<0.5	<0.5	<0.5		<0.5
As		<5	<5	<5		<5
Cd		0.9	0.7	0.5		0.7
Co	63	65	66	68	64	71
Cu	78	138	68	94	155	147
Li	10	20	20	10	20	20
Mo	2	<1	<1	<1	1	<1
Ni	537	451	561	467	509	551
Pb	3	<2	5	<2	185	<2
Sc	32	35	29	35	34	31
Tl		<10	<10	<10		<10
Zn	78	75	68	79	78	75
Au	18	<1	<1	<1	21	1
Pt	9.4	9.4	8.3	9.1	11.2	9
Pd	8.7	9.7	8.7	8.6	7.3	9.5

Coordinates in NAD83, zone 19V

Trace element concentration in ppm

Au, Pt, Pd concentrations in ppb

Sample No.	NQ0-13-07	NQ0-13-08	NQ0-13-09	NQ0-13-10	NQ0-13-11
description	Peridotite	Peridotite	Peridotite	Peridotite	Peridotite
Group	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq
UTMX	428647.44	428647.44	428647.44	428647.44	428647.44
UTMY	6670241.66	6670241.66	6670241.66	6670241.66	6670241.66
Ba	2.8	2.3	2.1	2.5	2.2
Ce	1.6	1.5	2.3	1.9	3.1
Cr	3860	3850	3350	3390	3350
Cs	0.07	0.09	0.22	0.26	0.22
Dy	0.67	0.6	0.89	0.72	1.13
Er	0.33	0.37	0.63	0.49	0.63
Eu	0.12	0.16	0.28	0.26	0.23
Ga	3.8	3.8	4.8	4.9	5.7
Gd	0.53	0.45	0.82	0.71	0.86
Hf	0.5	0.3	0.4	0.4	0.4
Ho	0.15	0.14	0.19	0.18	0.25
La	0.6	0.6	1	0.8	1.3
Lu	0.07	0.06	0.08	0.07	0.1
Nb	0.5	0.5	0.6	0.6	0.8
Nd	1.1	1	1.8	1.4	2.1
Pr	0.27	0.22	0.34	0.27	0.46
Rb	1.8	1.3	2.1	2.1	2.4
Sm	0.36	0.4	0.65	0.55	0.67
Sn	<1	<1	<1	<1	<1
Sr	2	2	2	2	3
Ta	<0.1	<0.1	<0.1	<0.1	<0.1
Tb	0.11	0.08	0.15	0.11	0.18
Th	0.05	0.07	0.07	0.08	0.11
Tm	0.06	0.06	0.09	0.08	0.1
U	<0.05	0.09	<0.05	<0.05	<0.05
V	94	93	116	114	132
W	<1	<1	1	1	1
Y	3.6	3.4	5.1	4.7	6.1
Yb	0.43	0.37	0.6	0.48	0.68
Zr	14	8	11	13	13
Ag	<0.5	<0.5	<0.5	<0.5	<0.5
As	<5	<5	<5	<5	<5
Cd	<0.5	<0.5	<0.5	0.7	<0.5
Co	131	136	136	162	114
Cu	254	89	136	754	74
Li	<10	<10	<10	<10	<10
Mo	<1	<1	<1	<1	<1
Ni	1840	1720	1840	2320	1410
Pb	3	<2	<2	3	<2
Sc	9	7	17	15	19
Tl	<10	<10	<10	<10	<10
Zn	78	83	93	86	96

Coordinates in NAD83, zone 19V

Trace element concentration in ppm

Appendix 9. Major and trace element concentrations for samples of the Roberts Lake Syncline provided by J.E. Mungall

Sample No.	JMPAR99-39	JMPAR99-41	JMPAR99-45	JMPAR99-47	JMPAR99-51	JMPAR99-52	JMPAR99-55	JMPAR99-60	JMPAR99-62	JMPAR99-65
description	basalt - pillow	basalt - pillow	basalt - pillow	basalt - pillow	basalt - pillow	basalt - pillow	basalt	basalt	basalt	basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	432123	431968	431839	431656	431240	431163	430814	430036	429763	429501
UTMY	6672063	6672044	6671937	6671995	6672139	6672172	6672161	6671778	6671820	6671986
SiO2	48.14	48.51	52.46	45.78	48.21	47.89	47.64	52.61	49.44	49.56
TiO2	1.39	1.34	2.09	1.71	1.65	1.89	1.59	1.36	1.46	1.93
Al2O3	13.63	13.90	12.97	15.25	13.08	13.28	14.40	12.02	13.73	13.54
Fe2O3	13.46	13.15	10.73	15.21	12.44	14.77	13.55	13.42	13.70	14.71
MnO	0.21	0.21	0.19	0.22	0.23	0.28	0.20	0.21	0.21	0.21
MgO	6.91	6.57	3.53	5.73	4.99	6.28	7.15	7.51	7.82	6.78
CaO	12.69	12.90	9.00	12.49	14.77	9.50	11.07	8.07	9.37	7.97
Na2O	1.47	1.72	5.66	1.85	2.15	3.61	2.73	3.97	3.25	4.02
K2O	0.07	0.11	0.34	0.12	0.12	0.12	0.11	0.41	0.16	0.17
P2O5	0.10	0.10	0.18	0.12	0.12	0.14	0.11	0.09	0.10	0.18
LOI	2.34	2.09	2.85	1.88	2.98	2.61	2.24	1.55	2.18	1.96
Total	100.40	100.60	100.01	100.35	100.74	100.36	100.80	101.23	101.42	101.02
Mg#	31	30	22	25	26	27	31	33	33	28
Cr										
Nb	5.20	4.87	9.06	6.60	5.93	6.34	5.68	4.92	5.24	7.97
Rb	2.35	2.23	8.65	3.15	2.86	2.35	2.94	9.63	3.93	3.13
Sr	298	279	86	204	98	46	142	79	73	46
Th	1.94	2.33	1.13	1.02	2.45	1.53	0.91	1.10	0.71	1.51
U	1.12				0.82	0.51			2.02	2.02
V	346	338	403	399	368	389	334	340	347	389
Y	23.7	23.1	30.4	26.9	27.3	30.5	22.2	22.4	25.2	29.3
Zr	80	78	143	100	105	115	87	66	89	118
Cu										
Ni	95	92	58	101	75	77	107	82	94	76

Coordinates in NAD83, zone 19V

Major element concentrations in wt. %

Trace element concentrations in ppm

Sample No.	JMPAR99-68	JMPAR99-72	JMPAR99-74	JMPAR99-78	JMPAR99-80	JMPAR99-82	JMPAR99-87	JMPAR99-33	JMPAR99-34	JMPAR99-36
description	basalt	basalt - sheet	basalt - sheet	basalt - sheet	basalt - pillow	basalt	basalt	basalt - sheet	basalt - sheet	basalt - pillow
Group	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	428874	426336	426460	426697	426811	426929	428001	432560	432499	432361
UTMY	6672074	6671768	6671836	6672015	6672130	6672189	6672231	6671893	6671931	6672031
SiO2	47.84	49.51	45.93	47.20	49.93	48.74	48.02	49.35	48.17	48.79
TiO2	1.96	1.90	1.87	1.41	1.39	1.44	1.06	2.72	2.63	2.54
Al2O3	13.79	13.40	14.59	14.35	13.46	13.76	14.43	12.87	12.49	13.20
Fe2O3	15.44	14.63	16.97	14.22	12.57	13.81	13.25	16.12	16.37	15.46
MnO	0.24	0.22	0.24	0.22	0.22	0.21	0.20	0.23	0.24	0.23
MgO	6.50	5.75	6.90	6.66	7.10	6.33	7.64	5.65	5.63	5.19
CaO	9.41	10.71	9.46	13.30	10.98	11.27	11.74	8.22	9.98	9.41
Na2O	3.23	3.05	2.64	1.51	3.03	1.70	2.33	2.88	2.31	2.98
K2O	0.17	0.11	0.10	0.04	0.22	0.08	0.15	0.18	0.28	0.48
P2O5	0.15	0.14	0.12	0.10	0.10	0.11	0.08	0.22	0.22	0.21
LOI	2.01	1.55	2.50	1.77	1.73	3.07	2.09	2.40	2.08	1.86
Total	100.74	100.97	101.32	100.76	100.73	100.50	101.00	100.86	100.40	100.35
Mg#	27	25	26	29	33	28	33	23	23	22
Cr										
Nb	8.51	7.85	6.78	5.35	4.75	5.23	3.13	13.71	13.32	12.49
Rb	3.75	1.91	2.93	1.31	5.76	3.18	4.04	6.09	7.53	8.93
Sr	97	76	125	242	173	237	151	187	162	152
Th	1.01	2.01	1.82	1.21	2.02	2.16	1.42	2.03	2.95	1.02
U							0.40	1.12		
V	393	389	402	334	324	356	296	492	483	485
Y	28.2	31.0	29.1	24.2	22.0	22.9	18.8	37.5	38.8	37.4
Zr	126	119	110	80	83	88	61	180	174	169
Cu										
Ni	94	64	94	90	88	97	105	62	57	72

Coordinates in NAD83, zone 19V

Major element concentratons in wt. %

Trace element concentrations in ppm

Sample No.	JMPAR99-59	JMPAR99-83	KB00R-1012	KB00R-1013	KB00R-1014	KB00R-1015	KB00R-1016	KB00R-1017	KB00R-1039	KB00R-1040
description	grey basalt	basalt	basalt	basalt	basalt	basalt	basalt	basalt	basalt	basalt
Group	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1	Group 1
UTMX	430342	427140	445943	just south	just south	just south	just south	just south	449435	448627
UTMY	6671818	6672165	6658536	of 1012	of 1012	of 1012	of 1012	of 1012	6663162	6664794
SiO2	49.34	49.82	50.72	48.42	49.67	48.45	49.82	49.25	45.43	48.51
TiO2	2.32	2.50	1.79	1.57	1.55	1.25	1.27	1.26	0.92	1.38
Al2O3	13.30	12.56	13.34	15.55	14.47	14.4	14.21	14.56	13.74	14.2
Fe2O3	13.77	15.32	14.98	12.87	13.79	13.98	13.73	13.9	14.41	12.99
MnO	0.20	0.22	0.22	0.19	0.21	0.21	0.21	0.2	0.24	0.2
MgO	5.78	5.70	7.3	4.83	5.59	5.76	7.86	7.14	8.75	7.01
CaO	10.08	10.29	9.03	11.74	11.43	13.17	8.87	10.08	14.55	11.1
Na2O	3.19	1.42	2.03	1.91	2.11	0.83	2.47	3.34	0.83	2.88
K2O	0.57	0.07	0.2	0.22	0.23	0.12	0.89	0.19	0.22	0.21
P2O5	0.21	0.21	0.17	0.13	0.12	0.09	0.09	0.09	0.05	0.13
LOI	1.64	2.57	0.9	2.19	1.49	2.07	1.27	0.8	1.44	1.8
Total	100.41	100.68	100.67	99.63	100.66	100.32	100.69	100.81	100.57	100.41
Mg#	27	24	30	24	26	26	33	31	34	32
Cr			414	245	254	400	281	287	356	382
Nb	17.92	13.15	6.5	4.7	4.6	3.7	3	3.6	0.7	6.9
Rb	13.97	3.47	3.5	3.6	3.3	2.5	10	4	8.4	3.8
Sr	213	266	81.1	128.2	109.9	171.3	113.2	64.4	149.2	253.8
Th	2.43	2.96	2.3	2.8	1.8	1.5	2	1.7	1.7	2
U	2.13		1.6	-1	-0.4	0.4	0.5	1	0	1.3
V	398	443	379.9	398.7	399.3	337.5	363.7	347.2	312.4	261.2
Y	28.6	35.1	34.1	27.2	27.2	22.7	22.7	21.8	25.8	19.6
Zr	175	168	117.5	92.7	91.3	66.1	68.4	67.7	26.1	81.3
Cu										
Ni	56	74	73	66	79	92	98	112	96	61

Coordinates in NAD83, zone 19V

Major element concentrations in wt. %

Trace element concentrations in ppm

Sample No.	KB00R-1041	KB00R-1042	KB00R-1043	JMPAR99-14
description	basalt	basalt	basalt	basalt
Group	Group 1	Group 1	Group 1	Group 2
UTMX	448574	447258	448064	434418
UTMY	6664786	6664645	6664024	6672550
SiO2	46	51.92	47.84	46.04
TiO2	1.56	0.95	1.19	0.71
Al2O3	14.17	17	13.24	10.45
Fe2O3	15.31	10.96	14.05	12.36
MnO	0.23	0.18	0.24	0.18
MgO	7.28	4.81	8.31	16.66
CaO	10.88	7.78	11.69	7.98
Na2O	2.32	3.79	2.23	1.53
K2O	0.1	0.07	0.13	0.04
P2O5	0.1	0.18	0.11	0.05
LOI	2.22	2.74	1.64	4.34
Total	100.18	100.36	100.67	100.33
Mg#	29	27	34	54
Cr	179	115	214	
Nb	5	5	2.6	1.67
Rb	2.6	1	2.8	2.50
Sr	278.1	235.8	122.7	59.59
Th	-0.1	1.8	1.7	0.00
U	-1.2	-0.6	1.4	
V	354.4	142.8	305.5	259.82
Y	23.6	18.8	20.3	16.15
Zr	76.2	81.8	53.6	41.98
Cu				
Ni	99	21	75	712.26

Coordinates in NAD83, zone 19V

Major element concentrations in wt. %

Trace element concentrations in ppm

Sample No.	JMPAR98-37	JMPAR98-39	JMPAR98-41	JMPAR98-45	JMPAR98-48	4001	4002	4003	4020
description	peridotite	peridotite	peridotite	peridotite	peridotite	peridotite	peridotite	peridotite	peridotite
Group	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq
UTMX	431040	431040	431040	431040	431040	paper map	paper map	paper map	paper map
UTMY	6666500	6666500	6666500	6666500	6666500	in thesis	in thesis	in thesis	in thesis
SiO2	41.50	39.84	38.86	40.17	46.61	39.36	40.57	44.00	42.61
TiO2	0.42	0.26	0.27	0.25	0.60	0.27	0.31	0.36	0.42
Al2O3	7.24	4.66	4.36	4.98	11.51	5.03	5.84	7.53	7.39
Fe2O3	13.48	13.11	11.97	12.86	12.05	13.83	13.24	12.64	13.23
MnO	0.16	0.18	0.17	0.18	0.17	0.18	0.20	0.17	0.15
MgO	24.62	29.64	31.05	29.10	14.72	28.38	26.65	22.78	23.53
CaO	5.46	3.59	3.95	3.82	9.40	3.81	5.20	6.75	5.85
Na2O	0.13	0.07	0.06	0.07	1.74	0.24	0.06	0.27	0.22
K2O	0.03	0.02	N.D.	0.01	0.21	0.02	0.02	0.05	0.04
P2O5	0.03	0.03	0.02	0.03	0.04	0.03	0.03	0.03	0.04
LOI	6.91	8.68	9.18	8.68	3.34	8.56	7.91	5.76	6.49
Total	99.98	100.57	99.91	100.65	100.63	100.29	100.45	100.73	100.47
Mg#	61	66	69	66	51	64	63	61	61
Cr		3017		3906	1566	3004	3011	2615	2678
Nb	1.4	11.0	0.9	11.1	7.4	2.9	2.7	2.7	3.0
Rb	8.5	4.3	2.0	4.2	10.9	2.2	3.2	3.2	4.3
Sr	20	2	7	5	125	5	8	20	11
Th	1.93		1.21			1.53		0.53	0.32
U	0.21			1.85	0.62				0.21
V	191	112	113	114	229	102	124	138	160
Y	11.1	8.9	6.3	10.6	18.2	6.8	8.1	8.4	10.1
Zr	27	4	16	4	11	17	23	22	27
Cu		1103		124	230	1540	403	532	701
Ni	1250	1496	1355	1217	391	1866	1081	897	2019

Coordinates in NAD83, zone 19V
Major element concentratons in wt. %
Trace element concentrations in ppm

Sample No.	4040	4008	4012	4013	4014	4019
description	gabbro	gabbro	gabbro	gabbro	gabbro	gabbro
Group	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq
UTMX	paper map	paper map	paper map	paper map	paper map	paper map
UTMY	in thesis	in thesis	in thesis	in thesis	in thesis	in thesis
SiO2	51.01	49.28	54.07	51.98	52.66	54.33
TiO2	0.72	0.65	0.72	0.69	0.64	0.62
Al2O3	13.39	12.10	13.40	12.67	12.47	12.33
Fe2O3	10.40	11.41	9.66	10.30	9.74	8.90
MnO	0.16	0.16	0.17	0.16	0.15	0.15
MgO	9.90	12.69	6.83	9.91	9.65	9.90
CaO	8.45	7.74	11.05	8.57	9.31	7.03
Na2O	1.73	2.05	2.42	2.35	2.94	2.84
K2O	2.16	0.98	0.45	1.44	0.75	2.14
P2O5	0.06	0.05	0.06	0.06	0.05	0.05
LOI	2.28	2.98	1.52	2.09	1.93	1.89
Total	100.51	100.35	100.45	100.42	100.49	100.43
Mg#	45	49	38	45	46	49
Cr	854	1603	394	1143	1091	1165
Nb	3.7	3.9	4.0	4.0	3.3	4.6
Rb	62.1	33.9	12.1	39.7	21.3	55.8
Sr	207	185	222	237	342	119
Th	2.14	3.18	3.64	0.71	3.15	2.74
U	0.61	0.92	0.51	0.81	0.71	0.61
V	252	214	230	218	212	205
Y	16.5	15.7	18.1	16.4	17.0	14.3
Zr	55	54	81	69	71	72
Cu	143	224	199	165	231	108
Ni	206	347	124	191	189	212

Coordinates in NAD83, zone 19V
Major element concentratons in wt. %
Trace element concentrations in ppm

Sample No.		JMPAR98-01	JMPAR 98-06	JMPAR98-08	JMPAR98-11	JMPAR98-12	JMPAR98-13	JMPAR98-21	JMPAR98-25	JMPAR98-26	JMPAR98-27
description		MS	MS	MS	MS	VEINS	MS	MS	MS	MS	MS
Group		Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq	Qarqasiaq
UTMX		paper map	paper map	paper map	paper map	paper map	paper map	paper map	paper map	paper map	paper map
UTMY		in report	in report	in report	in report	in report	in report	in report	in report	in report	in report
Co	ppm	2530	1829	2727	3191	59	3439	817	1467	1645	1764
Cu	ppm	1703	2160	2009	944	13985	909	4008	3258	3086	3246
Ni	ppm	>20000	>20000	>20000	>20000	603	>20000	4047	11319	15428	14535
Au	ppb	10	15	7	5	4	4	10	9	1	44
Pt	ppb	430	678	988	539	1450	731	90	386	274	418
Pd	ppb	106	2351	140	295	827	174	603	27	127	1970

Coordinates in NAD83, zone 19V