Testing the paleogeography of late Archean supercraton Superia
using pre-breakup 2.51-1.98 Ga dyke and sill provinces – with a
focus on the relationship between the Karelia-Kola and Superior
cratonic fragments

by

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Abstract

Archean cratons are dispersed crustal fragments derived from larger ancestral landmasses. Their individual and combined study in paleogeographic reconstructions reveal details relevant to understanding larger Archean tectonic systems. Canada's Superior craton is considered the central piece of the late Archean Superia supercraton paleogeographic reconstruction that also includes the Karelia-Kola (Finland/Russia), Wyoming (USA), and Hearne (Canada) cratons.

The relationship between Superior and Karelia-Kola is of interest because the position of the Karelia-Kola craton within the context of the Superia reconstruction is not fully resolved. Dyke swarm correlations based on age and geometry indicate that Karelia-Kola and Superior were 'nearest-neighbours', yet paleomagnetic studies disagree. However, these data are complicated due to multiple metamorphic overprints, and are limited by a lack of high-precision U-Pb ages.

The focus of this study is to test the correlation between the Karelia-Kola and Superior cratons by integrating U-Pb geochronology, whole-rock major, trace element and Sm-Nd isotope geochemistry, and dyke swarm geometry. These methods confirm the occurrence of sharply timed large igneous provinces (LIP) in Karelia-Kola correlated to Superior using geochronology, test whether the magmas were from a common source, and refine the paleogeographic Superia reconstruction by incorporating regional dyke trends.

This study presents nine ages from dykes and sills that confirm LIP correlations between the Karelia-Kola and Superior cratons at *ca.* 2.40 Ga, 2.22 Ga, 2.11 Ga and 1.98 Ga, and three ages at *ca.* 2.32 Ga confined to Karelia-Kola. Geochemical studies were completed on mafic magmatic events at *ca.* 2.22 Ga and 2.11 Ga. The evaluation of 2.22

Ga dykes and sills confirms that the Senneterre and Maguire dykes are compatible feeders to the Nipissing sills (Superior) and that the Karjalitic sills (Karelia-Kola) are not their direct continuation. At 2.11 Ga, mafic dykes and sills from Superior, Karelia-Kola, Wyoming and Hearne cratons were generated from two mantle end-member sources that mixed together and assimilated a crustal component.

In support of a nearest-neighbour connection between the Karelia-Kola and Superior cratons, the integration of U-Pb geochronology, geochemistry and dyke trends provide controls for a refined configuration of the Karelia-Kola craton within the Superia supercraton paleogeographic reconstruction.

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Statement of Contributions

Establishment of project

The motivation and large portion of financial support for this thesis comes from the Collaborative Research and Development (CRD) Grant entitled *A new tool for Canada's exploration industry: Using the large igneous province record to reconstruct supercontinents back to 2.7 Ga* (CRDPJ 419503-11). This CRD grant was awarded to Dr. Richard Ernst (as principal investigator) and a team of scientists, and builds on the Superia supercraton paleogeographic reconstruction hypotheses of Dr. Wouter Bleeker (Geological Survey of Canada) and others (Williams *et al.*, 1991; Heaman, 1997; Bleeker, 2003, 2004; Bleeker and Ernst, 2006; Ernst and Bleeker, 2010).

Fieldwork and sample collection

Fieldwork was completed over three summers from 2014 to 2016. In 2014, a few samples from the Superior craton, of the Nipissing and Marathon magmatic events, were collected with W. Bleeker. In the summer of 2015, Sarah Davey and W. Bleeker travelled to Finland and were guided by Drs. Frank Santaguida (formerly of First Quantum Minerals), Alex Neufeld (Anglo American plc), Jouni Vuollo, Hannu Huhma, Esa Heilimo, Asko Käpyaho and Peter Sorjonen-Ward (Geological Survey of Finland) over five weeks. S. Davey and W. Bleeker returned to Finland in 2016 to focus on dykes in the Pudasjärvi block.

Petrography and scanning electron microscope

Petrographic sections were prepared in the Rock Lab at Carleton University by Tim Mount. All transmitted light microscopy was completed by S. Davey at Carleton University. The

operation of the Scanning Electron Microscope was done with the assistance of Patricia Hunt at the Geological Survey of Canada or by Dr. Maryam Shahabi Far at Carleton University while either S. Davey and/or W. Bleeker were present.

Whole-rock major and trace element geochemistry

Twenty-nine sample powders for whole-rock major and trace element geochemistry of sills and dykes from Karelia were processed by S. Davey in the Rock Lab at Carleton University. This included washing, cutting, crushing, and pulverizing rocks, and polishing pucks for thin section. W. Bleeker supplied an additional 26 samples of Nipissing sills and 16 samples of Marathon dykes and sills that S. Davey also prepared for geochemical analysis. S. Davey received five rock cores of Marathon dykes that had been originally collected for paleomagnetic studies from Dr. Henry Halls (University of Toronto), and nineteen samples of Bear Mountain dykes from Dr. Kevin Chamberlain (University of Wyoming), Dr. David Evans (Yale University) and Dr. Taylor Kilian (formerly of Yale University) and these were similarly processed for geochemical analysis by S. Davey. All samples were analyzed at ALS Laboratories [www.alsglobal.com].

Sm-Nd isotopes

Under the guidance of Dr. Shuangquan Zhang and Dr. Brian Cousens, S. Davey completed 33 Sm-Nd isotope analyses at the Isotope Geochronology and Geochemistry Research Centre, Carleton University. S. Davey carried out the dissolution, liquid chromatography and thermal ionization mass spectrometry and S. Zhang spiked each sample. The laboratory assistant Evelyn Tennant prepared and outgassed many of the Re filaments.

U-Pb geochronology

Most of the rock samples for geochronology were crushed and separated by Lab Technicians Aleks Tiago and Kim Kwok at the University of Toronto, Jack Satterly Geochronology Laboratory. S. Davey completed the crushing and separation for the Sipojuntti and Myllykangas dykes. S. Davey picked the baddeleyite and zircon for all U-Pb analyses presented in this thesis except for the Triangle Mountain and Shakespeare intrusion samples, which were done by Drs. Sandra Kamo and Don Davis. S. Davey participated in the sample preparation, which involved annealing and chemical abrasion of zircons, loading samples into TeflonTM capsules, preparing columns, isotope dilution and analyzing the samples on the VG354 mass spectrometer. The role of S. Davey in preparing samples increased with gradual learning, taught by S. Kamo and K. Kwok over the summer of 2017 and 2018 at the Jack Satterly Geochronology Laboratory, University of Toronto.

Thesis and manuscript contributions

The writing and preparation of manuscripts and figures for each chapter of this thesis was completed by S. Davey with two exceptions. W. Bleeker wrote the final version of the geochronology results section in Chapter 4 (on the Nipissing sills) and the short note on the meaning of Finnish locality names, which was included as an appendix in the submitted version of Chapter 2. The figures and manuscript were significantly improved by proof-reading completed by W. Bleeker, R.E. Ernst, B. Cousens and S. Kamo. Chapter 4 benefitted from additional contributions from R.H. Sutcliffe and D. Davis.

Statement of originality

The work contained in this thesis is a continuation of decades-long research dedicated to understanding Earth's ancestral landmasses – the remains of which are fragmented and now embedded within the interiors of continents. My original contributions include the addition and interpretation of whole-rock major, trace element and Sm-Nd isotope geochemistry, tests for common magma sources, and U-Pb geochronology that colleagues and I use to refine the distribution of regional dyke swarms in Finnish Karelia and to test the configuration of the proposed Superia supercraton paleogeographic reconstruction.

Thesis Structure

This is an integrated thesis consisting of four independent chapters intended for publication. For this reason, some repetition will be found in the introduction and methods sections of each chapter. Three of these chapters have either been prepared for submission or are currently under review with peer-reviewed scientific journals. The fourth is an open file report published by the Geological Survey of Canada. While the theme for each paper varies individually, all are connected by the common goal of testing the Superia supercraton paleogeographic reconstruction using large igneous provinces (LIPs) located in Superior (Canada) and Karelia-Kola (Finland and Russia) cratons.

Chapter 2: Archean block rotation in Western Karelia: resolving dyke swarm patterns in metacraton Karelia-Kola for a refined paleogeographic reconstruction of supercraton Superia

Chapter 3: Structural controls on the trends of *ca.* 2.32 and 1.98 Ga dykes, Western Karelia.

Chapter 4: Ni-Cu-PGE potential of the Nipissing sills as part of the *ca.* 2.2 Ga Ungava large igneous province.

Chapter 5: Evidence for a single large igneous province at *ca.* 2.11 Ga across supercraton Superia.

Each chapter has been formatted to comply with the Carleton University thesis template, including sequential table and figure numbers corresponding to each chapter. A compiled list of references is included at the end.

Chapter 2 is currently under review with the journal *Lithos*. In this paper, S. Davey wrote the manuscript, drafted the figures, and prepared samples for major and trace-element whole-rock geochemistry analyzed by ALS laboratories (www.alsglobal.com). W. Bleeker and J. Vuollo participated in fieldwork with S. Davey for sample collection. S. Davey and W. Bleeker completed the SEM work with P. Hunt at the Geological Survey of Canada. Heavy mineral separates for U-Pb geochronology were prepared by A. Tiago, K. Kwok and S. Davey. Zircon and baddeleyite crystals were picked under a binocular microscope by S. Davey and S. Kamo. Chemistry and analyses of U-Pb isotopes for geochronology were completed by S. Davey, S. Kamo and K. Kwok. The manuscript benefitted from revisions by W. Bleeker, R.E. Ernst, S. Kamo and an anonymous reviewer.

Chapter 3 is presented as a pre-submission draft to be submitted to the journal

Tectonics. In this paper, S. Davey wrote the manuscript, drafted the figures, and prepared powders for whole-rock major and trace-element geochemistry. Samples were collected by S. Davey, W. Bleeker and J. Vuollo in Finland. Transmitted light petrography was completed by S. Davey at Carleton University. Backscattered electron images were taken by M. Shahabi Far at Carleton University with S. Davey. The whole-rock major and trace-element geochemical samples were analyzed at ALS Laboratories. Heavy mineral separates for these analyses were prepared by A. Tiago and K. Kwok. U-Pb geochronology was carried out at the Jack Satterly Geochronology Laboratory, University of Toronto by S. Davey under the guidance of S. Kamo and K. Kwok. The manuscript benefitted from revisions by W. Bleeker, R.E. Ernst, and S. Kamo.

Chapter 4 is published as an open file report by the Geological Survey of Canada. Please note that an additional discussion section **4.6** has been added to the original manuscript to provide clarity on how this chapter is relevant to the broader goals of the thesis. In this paper, S. Davey wrote the manuscript except for the geochronology results section, drafted the figures, and prepared powders for whole-rock major and trace-element geochemistry. Samples were provided by W. Bleeker and Dr. Kenneth Buchan and were analyzed by ALS laboratories for whole-rock major and trace element geochemistry. U-Pb geochronology for the Triangle Mountain and Shakespeare Intrusion were completed by S. Kamo and D. Davis, respectively. The geochronology section was co-written by S. Kamo and W. Bleeker. Sm-Nd isotopes were analyzed by S. Davey under the guidance of S. Zhang and B. Cousens. The manuscript benefitted from revision by co-authors S. Kamo, D. Davis, Dr. Michael Easton, and Dr. Richard H. Sutcliffe. Letters of permission from each co-author are presented in **Appendix A**.

Chapter 4 should be cited as follows:

Davey, S., Bleeker, W., Kamo, S., Davis, D., Easton, M., and Sutcliffe, R.H., 2019. Ni-Cu-PGE potential of the Nipissing sills as part of the *ca.* 2.2 Ga Ungava large igneous province; *in* Targeted Geoscience Initiative: 2018 report of activities, (ed.) N. Rogers; Geological Survey of Canada, Open File 8549, p. 403–419. <https://doi.org/10.4095/313675>

Chapter 5 is presented as a pre-submission draft intended for the Journal of Petrology. In this paper, S. Davey wrote the manuscript, drafted the figures, and prepared powders for a subset of the samples evaluated in this paper for major, trace-element and Sm-Nd isotope whole-rock geochemistry. Fieldwork for sample collection in Finland was completed by S. Davey, W. Bleeker, J. Vuollo and H. Huhma. Additional samples of the Marathon and Bear Mountain dykes were provided by W. Bleeker, Henry Halls, K. Buchan, D. Evans, T. Kilian and K. Chamberlain. Major and trace element geochemical analyses were completed by ALS Laboratories. Sm-Nd isotopes were completed at Carleton University by S. Davey under the guidance of B. Cousens and S. Zhang. Bulk mixing and assimilation and fractional crystallization models were calculated by S. Davey following the methods of DePaolo (1981) and after a helpful workshop at Carleton University delivered by Dr. Julian Pearce (Cardiff University; pers. comm, 2019). The work involving U-Pb geochronology was completed by S. Kamo while S. Davey observed. The manuscript benefitted from discussions with N. Kastek and C. Rogers, and revisions by W. Bleeker, R.E. Ernst, and S. Kamo.

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List of Abbreviations, Notations and Definitions

AFC – Assimilation-fractional crystallization

AE-ICPMS – Atomic emission-inductively coupled plasma mass spectrometry

Basement – Igneous or metamorphic rocks making up the crust onto which sedimentary rocks are deposited

Bdy – Baddeleyite, ZrO₂

CA – Chemical abrasion

Dyke – An igneous intrusion that cross-cuts stratigraphy vertically

HREE – Heavy rare earth element

ICPMS – Inductively coupled plasma mass spectrometry

ID – Isotope dilution

LILE – Large ion lithophile element

LIPs – Large igneous provinces

LREE – Light rare earth element

Magnesium (Mg) number:

$$\begin{aligned} &= \frac{MgO \text{ wt. \% sample}}{MgO \text{ molar mass}} \bigg/ \left[\frac{MgO \text{ wt. \% sample}}{MgO \text{ molar mass}} + \frac{FeO \text{ wt. \% sample} \times 0.9}{FeO \text{ molar mass}} \right] \\ &= \frac{MgO \text{ wt. \% sample}}{40.30} \bigg/ \left(\frac{MgO \text{ wt. \% sample}}{40.30} + \frac{FeO \text{ wt. \% sample} \times 0.9}{71.85} \right) \end{aligned}$$

MSWD – Mean square weighted deviation

PM – Primitive mantle. When ‘PM’ occurs as a subscript after an element or element ratio (e.g., Nb_{PM} or Tb/Yb_{PM}), the ‘PM’ subscript indicates that the reported element values are normalized to primitive mantle (Sun and McDonough, 1989; McDonough and Sun, 1995).

$\Delta Nb = 1.74 + \log(Nb/Y) - 1.92\log(Zr/Y)$; calculation by Fitton et al. (1997)

$Nb/Nb^* = Nb_{PM} / [(Th_{PM} \times La_{PM})^{0.5}]$ calculation as in Stepanova *et al.* (2014)

OSZ – Oulujärvi shear zone

Sill – An igneous intrusion that intrudes in parallel with stratigraphy.

TIMS – Thermal ionization mass spectrometry

TTG – Tonalite-trondhjemite-granodiorite

VGP – Virtual geomagnetic pole

Wilfley table – A table that is used to separate dense materials from light materials by shaking.

Zrc – Zircon, $ZrSiO_4$

List of Finnish Words

Finnish locality names can be long and bewildering to a non-Finnish speaker. In the Finnish back country, most localities take their name from the nearest lake (järvi in Finnish) or other geographical feature such as a hill or swamp or bog. With the Finnish language being a non-Indo-European agglutinative language, the geographical features typically get added at the end to other locality specific words to yield typically longer locality names, a simple example relevant to this study being “Pudasjärvi”, translating to Pudas-lake. It is for these reasons that many of the Finnish locality names used in this thesis end in one of the following, which may help in overall comprehension:

- maa (land or terrain, as in Palomaa)
- suo (swamp or bog, hence Suomi, “the land of swamps”)
- niemi (peninsula)
- järvi (lake, such as in Pudasjärvi)
- lampi (small lake, in a bog, without river in or out)
- joki (river or stream)
- koski (rapids or fast flowing river, but not quite a waterfall, as in Taivalkoski)
- kumpu (knoll or hillock, such as in Outokumpu, meaning “strange hill”)
- mäki (small hill, such as in Piippolanmäki)
- kukkula (hill, mound)
- tunturi (large hill or mountain, with bare summit)
- vaara (large hill or ridge, but with trees; such as in Vengasvaara)
- selkä (a ridge or spine)
- lehto (a forest grove, with deciduous trees, typically in a valley; as in Uolevinlehto)
- kangas (a dry upland boreal-type forest, without swamps; as in Myllykangas).

As can be seen from the shortlist above, the Finns or Suomalaiset (those who live in the land of swamps) have a bewildering variety of names for some of their most common landforms. But then again, English is not exactly sparse for words either, such as for varieties of hills: hump, mound, knoll, hillock, hill, rise, ridge, mount, mountain, butte, mesa, peak, massif...

Chapter 1

Introduction

Earth's geologic history is characterized by recurring assembly, rifting and drifting of continental lithosphere (Worsley *et al.*, 1982, 1984). A supercontinent is the union of most continental lithosphere into a single large landmass (Evans *et al.*, 2016). This landmass eventually breaks up, resulting in the independent dispersion of stable continental fragments (i.e., cratons¹). The fragmented cratons drift and collide with arc terranes and other continental fragments, leading to the formation of a new supercontinent. Better known as the supercontinent cycle, our understanding of its episodic nature is founded on the concepts and processes of the Wilson cycle, continental drift, and plate tectonics (Wegener, 1915; Wilson, 1966, 1968; Worsley *et al.*, 1984; Nance *et al.*, 2014).

There have been at least three and possibly four supercontinents over Earth's history (Bleeker, 2003). The most recent, and best understood of these is Pangaea, with its assembly at *ca.* 320 Ma to its break-up starting at 195 Ma (Matte, 2001; Stampfli *et al.*, 2013). Earlier supercontinents formed and fragmented in the Precambrian era including *ca.* 0.9 Ga Rodinia, *ca.* 1.5 Ga Nuna/Columbia, and the widely debated *ca.* 2.5 Ga Kenorland (Bleeker, 2003; Evans *et al.*, 2016). With each step further into deep time, our knowledge of these former or putative supercontinents is reduced by degraded preservation of their

¹ The definition of a craton from the Fifth Edition of the Glossary of Geology (Neuendorf *et al.*, 2005):

“**Craton** (cra'-ton): A part of the Earth's crust that has attained stability and has been little deformed for a prolonged period of time. The term is now restricted to continental areas that have not been pervasively metamorphosed and deformed for at least about one billion years...” *p.* 149.

geology. Moreover, whether the model of the supercontinent cycle was operational and whether the processes responsible are remotely comparable between what is observed today and those operating in the Archean Eon (4-2.5 Ga) are still widely deliberated (Bleeker, 2003; van Hunen and Moyen, 2012; Keppie, 2016).

There are about 35 sizeable fragments of Archean crust (i.e., cratons *sensu lato*) dispersed across the globe that are incorporated within Earth's major continents (**Figure 1-1**; Bleeker, 2003). Note that these ~35 main fragments in addition to smaller slivers are only those that have been sufficiently preserved and that this is not an exhaustive tally, as there surely were other fragments or slivers but these have either been destroyed via subduction erosion, or metamorphosed and deformed beyond recognition (Bleeker, 2003; von Huene *et al.*, 2004). Some of the earliest work on Archean paleogeographic reconstructions materialized from regional studies of continental North America (Hoffman, 1988; Williams *et al.*, 1991). These authors concluded that Archean domains must be fragments of a formerly extensive landmass because they are surrounded by passive (rifted) margins and because their orogenic belts are truncated by younger Paleoproterozoic structures. These observations led to the prediction that the Archean domains embedded within continental North America including the Slave, Superior, Wyoming, Hearne and Rae (formerly North and South Keewatin), and Meta Incognita and Greenland (formerly North Atlantic), are the 'offspring' of a former 'parent' supercontinent called Kenorland (Williams *et al.*, 1991).

Figure 1-1. Distribution of ~35 Archean cratons across the globe with cratons discussed in this study including Superior, Hearne, Wyoming and Karelia-Kola, in red. Modified after Bleeker (2003). See Appendix A for permission to reproduce this figure.

The proposed existence of Kenorland prompted more detailed studies comparing the geology of each of the Archean cratons in North America and sparked the expansion of these ideas on a global scale. Roscoe and Card (1993), for example, completed a stratigraphic study that concluded that the sediments from the Huronian (in Canada) and Snowy Pass basins (in Wyoming, USA), now separated by 2000 km and located on separate cratons, Superior and Wyoming, were deposited in the same epicratonic basin that broke up in the Paleoproterozoic. With the rapid development of U-Pb geochronology resulting in more precise ages on mafic magmatic rocks, Heaman (1997) found that the *ca.* 2.45 Ga Matachewan dyke swarm had age matches in the Karelia-Kola craton and proposed that the Karelia-Kola craton was an adjacent fragment within a single Archean supercontinent involving the Superior and Wyoming cratons.

Based on detailed comparison of basement geology between the Slave and Superior cratons, Bleeker (2003) found that, despite their proximity, these two cratons shared no common geological events aside from very early *ca.* 2.73-2.70 Ga basalt-komatiite volcanic rocks. These detailed comparisons considered the timing of basement formation interpolated from detrital zircon studies, mafic dyke swarms, and general sequence stratigraphy. Consequently, Bleeker (2003) concluded that the Superior and Slave cratons were either distant fragments of a Kenorland-like supercontinent or derived from distinct supercratons, Superia and Sclavia. The Archean paleogeographic record, therefore, does not necessitate the existence of a single supercontinent. Instead, Archean paleogeographic reconstructions may have involved a few supercratons, that during their break-up, produced cratonic fragments with unique basement histories (Bleeker, 2003). The key difference between a supercontinent and supercraton is that a supercraton, while similar in that it is a

large Archean landmass with a stabilized core, does not require the complete amalgamation of all crust and can coexist with other supercratons (**Figure 1-2**).

Figure 1-2. A diagram that shows the difference between (A) the Archean supercontinent Kenorland as proposed by Williams *et al.* (1991) and (B) few co-existing supercratons Sclavia and Superia (among others), modified from Bleeker (2003). See Appendix A for permission to reproduce this figure.

The study of Earth's former supercratons and their paleogeographic reconstruction through time depends on the integration of a variety of methods and datasets, incorporating geological observations from assembly to break-up. The assembly of a supercontinent is recorded by the products of continental collision preserved as metamorphic terranes (e.g., Hoffman, 1988; Williams *et al.*, 1991). The time period in which a supercontinent is assembled is represented by sedimentary and volcanic rocks deposited as supracrustal

assemblages in intra-cratonic basins (e.g., Roscoe and Card, 1993). Widespread mafic magmatism related to large igneous provinces (LIPs) is associated with break-up, and often involves multiple events including many failed rifting attempts (Bleeker, 2003). However, these geological processes operate over vastly different time scales. Orogenic and sedimentary processes occur over hundreds of millions of years, whereas LIPs are typically emplaced within a few or tens of million years. It is this narrow emplacement time that can be refined by high-precision U-Pb ID-TIMS geochronology, and vast areal extent, covering more than one cratonic fragment, that make LIPs the ideal candidate for testing paleogeographic reconstructions at precise temporal intervals through Earth's history (Bleeker, 2004; Bleeker and Ernst, 2006; Ernst and Bleeker, 2010). The applicability of LIPs to the Archean paleogeographic puzzle is best illustrated by the 'magmatic barcode' of Bleeker (2004). The magmatic barcode, a graphical representation of LIPs on a given craton through time, provides a framework with which to test correlations between 'nearest-neighbour' Archean cratons. Formerly adjacent (i.e., 'nearest-neighbour') Archean cratons should have multiple LIPs in common, matching precisely in age, during their supposed time together, within their ancestral supercraton.

1.1 Large igneous provinces (LIPs)

Large igneous provinces (LIPs) are intraplate magmatic episodes involving tremendous volumes of magma manifested as dyke swarms, sill provinces, and volcanic extrusive rocks such as flood basalts and mafic volcanic rocks (Coffin and Eldholm, 1991, 1992, 1994; Bryan and Ernst, 2008; Ernst, 2014). First defined by Coffin and Eldholm (1991, 1992, 1994), the events are estimated to generate enough magma to cover areas of at least 0.1

Mkm² and volumes of at least 0.1 Mkm³ (Bryan and Ernst, 2008). Many LIPs exceed these estimates and are larger than 1 Mkm² and 1 Mkm³. These magmas may be emplaced within continental or oceanic settings, and are predominantly mafic-ultramafic in composition (Coffin and Eldholm, 1992); however, silicic components (termed SLIPs) are recognized as well (Bryan, 2007). LIPs are emplaced rapidly, and in short pulses lasting approximately a million years (i.e., 66 Ma Deccan Trap event; Schoene *et al.*, 2015) and some LIPs may include multiple pulses (i.e., *ca.* 2.48 and 2.45 Ga Matachewan LIP event; Ernst and Bleeker, 2010; Ciborowski *et al.*, 2015). Because LIPs are generated in intraplate settings, the processes that form them are not due to ‘classic’ plate boundary processes such as spreading at the mid-ocean ridge or back-arc extension related to subduction zones (Bryan and Ernst, 2008). Instead, mechanisms for LIP formation include delamination of the lower crust/upper mantle or by the arrival of a buoyant mantle plume ascending from a deeper thermal boundary layer, perhaps the core-mantle boundary (Foulger, 2011; Ernst, 2014).

The relationship between LIPs and continental break-up events have been recognized since their early descriptions (Coffin and Eldholm, 1992). As a driver of continental break-up, the magmatic components provide temporal, petrogenetic and geometric constraints on the cratonic fragments they have divided (i.e., former ‘nearest-neighbours’). With their vast areal distributions, rapid emplacement, and extensive dyke swarms, LIPs bear the ideal characteristics for studying paleogeographic reconstructions in detail. The ‘magmatic barcode’, emphasized by Bleeker (2004), capitalizes on the short emplacement time of LIPs. When a magmatic barcode (LIP events through time in a given craton) has contemporaneous LIP events with other cratons, then it is plausible that they may have shared a common history (Bleeker, 2004; Bleeker and Ernst, 2006). Geometric

control can be obtained from these same magmatic events through paleomagnetic studies, which can provide a paleolatitude and paleo-orientation at the time of magma crystallization. In particular, the radiating patterns associated with dyke swarms provide additional geometric control for paleogeographic reconstructions. Lastly, geochemistry is a tool that can be used to evaluate the tectonic settings and conditions under which these magmas formed, and therefore provide a means of testing whether the dykes, sills, or flood basalts thought to be part of the same LIP event were generated under similar conditions and/or from a similar source. However, as we will demonstrate, geochemistry of LIPs is commonly non-unique and should therefore only be used as a secondary tool that is considered alongside geochronological and/or paleomagnetic evidence.

1.2 Superia supercraton paleogeographic reconstruction

The Superia paleogeographic reconstruction identifies the Superior, Hearne, Wyoming, and Karelia-Kola cratons as nearest-neighbours in an ancestral supercraton from the late Neoproterozoic to early Paleoproterozoic. The reconstruction is based on U-Pb geochronology and distributions of dyke swarms, including their radiating geometry used to identify piercing points (magmatic centres) surrounding the margin of the Superior craton (**Figure 1-3**; Bleeker, 2003; Bleeker and Ernst, 2006; Ernst and Bleeker, 2010).

The barcode method (Bleeker, 2004) reveals multiple LIPs that match in age across these cratons from *ca.* 2.51 to 1.98 Ga (**Figure 1-4**). Some of the most significant age matches between the Superior and Karelia-Kola cratons are at *ca.* 2.45 Ga, *ca.* 2.22 Ga and 2.11 Ga (Heaman, 1997; Ernst and Bleeker, 2010). The positioning of Karelia-Kola in the paleogeographic reconstruction by Bleeker and Ernst (2006), and updated in Ernst and

Bleeker (2010), is largely controlled by these three events. The dykes in Karelia-Kola that are *ca.* 2.45 and *ca.* 2.11 Ga trend towards the magmatic centres of contemporaneous dykes in the Superior craton including the Matachewan and Marathon events, respectively. Although sills do not provide a geometric control, Bleeker and Ernst (2006) hypothesized that the Karjalitic sills at *ca.* 2.22 Ga in Karelia-Kola are the ‘downstream’ continuation of the Nipissing sills.

The conventional application of paleomagnetic studies and their resulting paleolatitudes to paleogeography have so far proven inconclusive for the Karelia-Kola craton (Mertanen et al., 1999; Salminen et al., 2014). Ambiguity in the meaning of paleomagnetic results is due to pervasive overprinting by multiple regional metamorphic events, adding to the complexity of the magnetic components measured in each sample. The mafic magmatic rocks studied in the Karelia-Kola craton include four magnetic components: A, B, D and D’ (Mertanen et al., 1999; Salminen et al., 2014). Components A and B are secondary and unrelated to primary magnetization. Component A is attributed to the metamorphic overprint by the Svecofennian orogeny and B is related to a younger, unspecified regional event (Mertanen et al., 1999; Salminen et al., 2014). The D and D’ VGP (virtual geomagnetic pole) components are variably found (alone and together) in the *ca.* 2.44 Ga gabbro-norite dykes of Mertanen et al. (1999) and in the *ca.* 2.40 Ga WD and *ca.* 2.33 Ga AD13 dykes of Salminen et al. (2014). Both the D (20°S, 278.8°E) and D’ (VGP 9.6°N, 256.2°E), components are candidates for the primary magnetic remanence of the dykes (Mertanen et al., 1999). In cases where only the D component is measured in dykes, the accompanying baked contact tests are negative (such that D was also measured in the host rocks at distances beyond a full dyke width; Mertanen et al., 1999). However,

Mertanen et al. (1999) suggests that the occurrence of D in the host rock may be related to regional heating cause by the ca. 2.44 Ga intrusions themselves. The D' component has been found in one dyke and in the contacts and baked zones of other dykes from Mertanen et al. (1999), and the AD13 dyke of Stepanova et al. (2014). While no positive baked contact test has been obtained, evidence that the D' component is primary includes a higher unblocking temperature (compared to D) and the fact that it has been recorded in at least one dyke sample from the Varisniemi locality.

Use of the D pole or D' VGP for the purposes of reconstruction yield different paleolatitudes and therefore paleogeographic reconstruction solutions. If the D pole is favoured, then the Karelia-Kola craton is located far from the Superior craton (as is the preferred interpretation of Salminen et al., 2014). Conversely, if the D' pole is preferred (as used by Bleeker and Ernst, 2006), the Karelia-Kola craton is located close to Superior.

Figure 1-3. The Superia supercraton paleogeographic reconstruction as presented by Ernst and Bleeker (2010) joining the Superior, Hearne, Wyoming and Karelia-Kola Archean cratons in the Neoproterozoic to early Paleoproterozoic. LIP abbreviations: MR: Marathon, MT: Matachewan, MS: Mistassini, SE: Senneterre, FF: Fort Frances, FN: Franklin dyke, BD: Blue Draw gabbro, BM: Bear Mountain, WR: Wind River, MM: Metamorphosed dykes and sills, KE: Kennedy, SP: Snowy Pass sediments and gabbros, KA: Kaminak, K: Karjalitic. The north arrows show current orientations of each block. The black circle surrounding the Matachewan star indicates the possible plume size. See Appendix A for permission to reproduce this figure.

1.3 Objective

The primary objective of this thesis is to test the hypothesized nearest-neighbour relationship of the Superior and Karelia-Kola Archean cratons. In simple terms, the goal is to answer whether the ancient pieces of continental crust, Superior and Karelia-Kola, were connected 2.5-2.0 billion years ago (Ga) within a single, larger supercraton called Superia. To answer this question, three goals emerge related to the study of LIPs:

1. To improve the magmatic barcode record of the Karelia-Kola and the Superior cratons with U-Pb ID-TIMS geochronology on mafic dykes, sills, and other intrusive bodies.
2. Update the regional dyke swarm mapping in Karelia-Kola, with a specific focus on better defining the dyke swarm patterns in Western Karelia.
3. Compare whole-rock major and trace-element, and Sm-Nd isotope geochemistry, of contemporaneous magmatic events in the Superior and Karelia-Kola craton.

1.4 Approach

As discussed by Bleeker and Ernst (2006), there is a disparity between datasets of the Archean cratons involved in the Superia supercraton paleogeographic reconstruction. Extensive and detailed mapping, U-Pb geochronology, paleomagnetic and geochemical datasets exist for the Superior craton because it is well exposed, well preserved and has been a mineral exploration target for more than a century. It is this abundance of multiple datasets that makes the Superior craton the standard against which others (Karelia-Kola, Hearne, and Wyoming cratons) are compared. While similar datasets are available for the Karelia-Kola craton, their extent and detail are much more limited because of poor

exposure, and significant re-working and overprint by the 1.9 Ga Svecofennian orogeny. This contrast in datasets is the main motivation for concentrating our sample collection efforts in the Finnish portion of the Karelia-Kola craton. Of the 9,931 dykes and sills mapped in the Finnish bedrock digital database (GTK, 2012), 6,076 of these features are unclassified. With 61% of the dykes and sills in Finnish Karelia unclassified, we strategized our sampling collection by primarily targeting dykes and sills categorized either from having preliminary geochronology or trends that have been correlated to widespread events known in the Superior craton. Targeted LIPs include the *ca.* 2.45 Ga Matachewan dyke swarm matching the Pääjärvi dykes, the *ca.* 2.22 Ga Ungava LIP matching the Karjalitic (or Jatulian) sills, and the *ca.* 2.11 Ga Marathon dyke swarm matching the Tohmajärvi dykes, sills and mafic volcanic rocks (**Figure 1-4**). These events were selected as targets, not only because of their widespread and well documented occurrence in the Superior craton, but also because they are variably found in the Hearne and Wyoming cratons, and together span almost the entire time period when Superior and Karelia-Kola were likely adjacent as integral pieces of the ancestral supercraton Superia.

Figure 1-4. *On the next page.* Magmatic barcodes for the Superior, Hearne, Wyoming, and Karelia-Kola cratons. Magmatic events discussed and dated in this thesis are highlighted by their map unit colour. Ch. (chapter) numbers indicate the chapter in which dates and discussion can be found for the particular magmatic event. Age references: Superior craton: Fahrig (1987); Buchan *et al.* (1993, 1996, 2007); Machado *et al.* (1993); Krogh (1994); Heaman (1997); Halls and Heaman (2000); Buchan and Ernst (2004); Halls and Davis (2004); Bleeker and Ernst (2006); Bleeker *et al.* (2008); Halls *et al.* (2008); Maurice *et al.* (2009); Ernst and Bleeker (2010); Bleeker *et al.* (2016); Kastek *et al.* (2018); M. Hamilton, pers. comm. (2019). Hearne craton: Heaman and LeCheminant (1993); Ansdell (2005); Sandeman *et al.* (2013); Grohn *et al.* (2016); Regan *et al.* (2016). Wyoming craton: Redden *et al.* (1990); Cox *et al.* (2000); Bowers and Chamberlain (2006); Dahl *et al.* (2006); Hark (2008); Kilian *et al.* (2016a). Karelia-Kola: Alapieti (1982); Huhma *et al.* (1990); Balashov *et al.* (1993); Amelin *et al.* (1995); Puchtel *et al.* (1997); Smolkin (1997); Vaasjoki (2001b); Vuollo and Huhma (2005); Hanski and Huhma (2005); Hanski *et al.* (2010, 2014); Kulikov *et al.* (2010); Martin *et al.* (2013a, 2013b); Salminen *et al.* (2014); Stepanova *et al.* (2014, 2015, 2017).

1.5 Chapter rationale

LIP events at *ca.* 2.45 Ga, *ca.* 2.22 Ga and *ca.* 2.11 Ga have been previously studied in varying detail in the Superior and Karelia-Kola cratons (e.g., Huhma, 1986; Conrod, 1989; Bates and Halls, 1991; Pekkarinen and Lukkarinen, 1991; West and Ernst, 1991; Noble and Lightfoot, 1992; Vuollo and Phrainen, 1992; Heaman and LeCheminant, 1993; Lightfoot *et al.*, 1993; Nykänen *et al.*, 1994; Amelin *et al.*, 1995; Buchan *et al.*, 1996, 1998; Heaman, 1997; Vaasjoki, 2001b; Hamilton *et al.*, 2002; Vuollo and Huhma, 2005; Karhu *et al.*, 2007; Halls *et al.*, 2008; Maurice *et al.*, 2009; Maurice and Francis, 2010; Stepanova and Stepanov, 2010; Hanski *et al.*, 2010; Ciborowski *et al.*, 2015; Huhma *et al.*, 2018). For this reason, data availability (or lack thereof) guided the collection of new information. **Chapter 2** and **Chapter 3** are concerned with U-Pb geochronology on dykes in Western Karelia. New ages are combined with previously published ages and observed trends to structurally restore the deformation history of the western margin of the Karelia-Kola craton, and consequently reveal regional dyke swarm patterns necessary for paleogeographic reconstructions. **Chapter 4** and **Chapter 5** present geochemical analyses on the *ca.* 2.22 and *ca.* 2.11 Ga events, which aim to characterize the petrogenesis of their magmatic products.

In **Chapter 2**, U-Pb geochronology is presented on three *ca.* 2.40 and *ca.* 2.11 Ga dykes from the Pudasjärvi block, Western Karelia. It has been well documented that the trends of dykes are not continuous from the interior of the Karelia-Kola craton to the Pudasjärvi block. With our new U-Pb geochronology combined with previous studies, we propose a structural restoration that requires $\sim 30^\circ$ of vertical axis rotation.

In **Chapter 3**, we present seven U-Pb ages on *ca.* 2.40, *ca.* 2.32, and *ca.* 1.98 Ga

dykes from the Iisalmi, Kuhmo and Taivalkoski blocks, Western Karelia. These dykes were selected because of their existing assignment to the *ca.* 2.45 Ga Pääjärvi swarm and because no U-Pb age existed for the NE-trending dykes in the Taivalkoski block. These ages add to the U-Pb geochronology database and provide detail on regional dyke swarms.

In **Chapter 4**, our focus shifts to the Superior craton and the *ca.* 2.22 Ga Ungava LIP. In contrast to many mafic magmatic events in the Karelia-Kola craton, the *ca.* 2.22 Ga Karjalitic sills are well studied. Before a geochemical comparison could be completed between the *ca.* 2.22 Ga Ungava and Karjalitic sill events, confirmation of the hypothesis that the Senneterre and Maguire dykes are geochemically compatible feeders to the Nipissing sills was required. The whole-rock major and trace element and Nd isotope geochemical study presented in Chapter 4 confirms that at least some of the dykes of the Ungava LIP are geochemically similar, and therefore viable feeders, to the Nipissing sills.

In **Chapter 5**, U-Pb geochronology is presented for the Nieminen dyke in Western Karelia, precisely dated at 2114 ± 3 Ma, which confirms an age match to dykes in the Superior, Wyoming and Hearne cratons. To fully test for a geochemical match between Karelia-Kola and relevant magmatic rocks in other fragments of the Superia supercraton (Superior, Wyoming and Hearne cratons), 13 dykes and sills from Superior, 20 dykes from Wyoming and 3 dykes from Karelia-Kola cratons were analyzed for whole-rock major and trace element geochemistry. In addition, 27 Sm-Nd isotope analyses were run on samples from Superior (13), Wyoming (11) and Karelia-Kola (3) cratons for more detailed comparison.

Chapter 6 summarizes of our findings and discusses future work that should be completed for further testing of the Superia supercraton paleogeographic reconstruction.

Chapter 2

Archean block rotation in Western Karelia: resolving dyke swarm patterns in metacraton Karelia-Kola for a refined paleogeographic reconstruction of supercraton Superia

2.1 Abstract

Rifting, breakup, and subsequent collision related to the *ca.* 1.92-1.79 Ga Svecofennian orogeny fragmented and deformed the western margin of the Archean Karelia-Kola craton into four crustal blocks: Pudasjärvi, Iisalmi, Kuhmo, and Taivalkoski. Detailed quantification of Svecofennian deformation is limited due to poorly exposed basement geology and an incomplete dyke swarm record.

New U-Pb ID-TIMS geochronological results on baddeleyite and zircon are presented for several key mafic dyke swarms from the Pudasjärvi block, namely the Uolevinlehto, Myllykangas, and Sipojuntti dykes. The age of the 325°-trending Uolevinlehto dyke is estimated at *ca.* 2400±12 Ma from discordant multigrain baddeleyite, suggesting it to be younger than *ca.* 2450 Ma dykes in the area. The 350°-trending Myllykangas dyke has a minimum age of 2135.2+3.6/-3.7 Ma based on two single grains of chemically abraded zircon. Results from three single baddeleyite grains from the 320°-trending Sipojuntti dyke provide a precise upper intercept age of 2128.9±1.2 Ma.

Our new U-Pb ages are integrated with those from the literature to define six major dyke swarms in the Pudasjärvi block: the WNW-trending *ca.* 2450 Ma Pääjärvi, NNE-trending *ca.* 2400 Ma Uolevinlehto, NW-trending *ca.* 2135-2100 Ma Tohmajärvi, WNW-

trending *ca.* 2077±35 Ma Palomaa, NNW-trending *ca.* 1980 Ma Paukkajanvaara and undated “East-West” dykes. A comparison of coeval dykes between blocks reveals that the Svecofennian collision resulted in a ~30° clockwise vertical-axis rotation of the Pudasjärvi block relative to the interior of Karelia. After restoring this vertical axis rotation back in time, we use the trends and ages of regional dyke swarms in Karelia to revise the paleogeographic reconstruction of supercraton Superia during the early Paleoproterozoic prior to continental breakup.

2.2 Introduction

Essential to the study of Earth’s late Archean geological record and reconstruction of Archean crustal fragments (*s.l.* “cratons”) into their ancestral Archean landmasses (“supercratons”), are constraints obtained from geochronology, paleomagnetism, geochemistry, and the geometry of mafic dyke swarms as part of large igneous provinces (LIPs). Bleeker and Ernst (2006) demonstrated the utility of mafic dyke swarms for continental reconstructions by identifying several Archean cratons that may have been “nearest neighbour” fragments within supercraton Superia, the ancestral landmass of which the Superior craton is the central and largest fragment (Bleeker, 2003). Nearest-neighbour Archean cratons are identified by matching multiple contemporaneous LIPs, and further support is drawn from coincident metamorphic events, and correlation of sedimentary cover sequences and latitudes from paleomagnetic studies (Bleeker and Ernst, 2006; Pesonen *et al.*, 2012; Pehrsson *et al.*, 2013). Correlations in the Paleoproterozoic LIP magmatic barcode identified between different crustal blocks (the LIP record of a given

craton through time; **Figure 2-1**; Bleeker, 2004; Bleeker and Ernst, 2006) have linked the Superior craton (Canada) to the Hearne (Canada), Wyoming (USA), and Karelia-Kola (Finland and Russia) cratons. The contemporaneous events for Superior (S), Hearne (H), Wyoming (W) and Karelia-Kola (K) are: 2.45 Ga of the Matachewan dykes (S), Blue Draw Metagabbro (W), and Pääjärvi dykes (K); the 2.22 Ga Ungava dykes and Nipissing sills (S), Clay Butte dykes (W) and Karjalitic sills (K); and the 2.11 Ga Marathon (S), Griffin (H), Bear Mountain (W) and Tohmajärvi dykes (K) (**Figure 2-1**; Heaman, 1997; Mertanen *et al.*, 1999; Bleeker, 2003, 2004; Bleeker and Ernst, 2006; Dahl *et al.*, 2006; Reddy and Evans, 2009; Ernst and Bleeker, 2010). Except for the more refined reconstruction of Superior-Wyoming based on multiple paleomagnetic data (Kilian *et al.*, 2016a, 2016b), the exact positioning of many cratonic fragments within the context of supercraton Superia remains to be determined. Uncertainty in the original paleogeographic configuration of cratonic fragments is especially true for the Karelia-Kola craton where paleomagnetic studies are complicated by overprints and report disparate paleolatitudes and apparent polar wander paths compared with the Superior craton (e.g., Mertanen *et al.*, 1999; Pesonen *et al.*, 2012; Salminen *et al.*, 2014).

Craton-scale dyke swarm patterns in the Karelia-Kola craton are particularly challenging to interpret due to the partially dismembered and variably remobilized Archean basement blocks that comprise it. In the sense of Abdelsalam *et al.* (2002), Karelia-Kola is truly a “metacraton” because of significant reworking during the Svecofennian orogeny. Because the original spatial relationships between the Pudasjärvi, Iisalmi, Kuhmo and Taivalkoski blocks are not fully known and because dyke swarm patterns are inconsistent from one block to another, craton-scale swarm patterns necessary for accurate

paleogeographic reconstruction models remain poorly defined.

Metamorphism from the *ca.* 1.92-1.79 Ga Svecofennian orogeny (Lahtinen *et al.*, 2011) has obscured the magnetic signatures of some dyke swarms typically visible in aeromagnetic maps and variably overprinted the paleomagnetic signatures of dykes in western Karelia. Metamorphism has also compromised the geochronological record (e.g., Vaasjoki, 2001b; Vuollo and Huhma, 2005). Metamorphism can promote zircon growth and Pb loss, as well as secondary zircon overgrowth on baddeleyite, which will lead to discordance and potentially inaccurate age interpretations (Söderlund *et al.*, 2013; Bleeker, 2014). Most of the U-Pb geochronological studies in Finnish Karelia were conducted when multigrain fractions weighing at least a few milligrams were required to circumvent larger Pb blanks and ensure emission of sufficient Pb and U signals (Vaasjoki, 2001b). Subsequent advances in U-Pb geochronological methods and reduction of Pb background levels permit the measurement of smaller quantities of material, as we demonstrate in our U-Pb geochronology of the Sipojuntti dyke (Perttunen and Vaasjoki, 2001).

Figure 2-1. *On the next page.* Magmatic barcode records for the Superior, Wyoming and Karelia-Kola cratons (modified after Bleeker, 2004). Well-established barcode matches are highlighted in light grey and those discussed in this study are highlighted in pink. Age references, Superior craton: Fahrig (1987); Buchan *et al.* (1993, 1996, 2007); Machado *et al.* (1993); Krogh (1994); Heaman (1997); Halls and Heaman (2000); Buchan and Ernst (2004); Halls and Davis (2004); Bleeker and Ernst (2006); Bleeker *et al.* (2008); Halls *et al.* (2008); Maurice *et al.* (2009); Ernst and Bleeker (2010); Bleeker *et al.* (2016); Kastek *et al.* (2018); M. Hamilton, pers. comm. (2019). Hearne craton: Heaman and LeCheminant (1993); Ansdell (2005); Sandeman *et al.* (2013); Grohn *et al.* (2016); Regan *et al.* (2016). Wyoming craton: Redden *et al.* (1990); Cox *et al.* (2000); Bowers and Chamberlain (2006); Dahl *et al.* (2006); Hark (2008); Kilian *et al.* (2016a). Karelia-Kola: Alapieti (1982); Huhma *et al.* (1990); Balashov *et al.* (1993); Amelin *et al.* (1995); Puchtel *et al.* (1997); Smolkin (1997); Vaasjoki (2001b); Vuollo and Huhma (2005); Hanski and Huhma (2005); Hanski *et al.* (2010, 2014); Kulikov *et al.* (2010); Martin *et al.* (2013a, 2013b); Salminen *et al.* (2014); Stepanova *et al.* (2014, 2015, 2017).

The purpose of this study is to refine the dyke swarm record within the Pudasjärvi block of the Karelia-Kola metacraton by 1) contributing three U-Pb isotope dilution thermal ionization mass spectrometry (ID-TIMS) ages, and 2) integrating these new ages with previously published U-Pb geochronology. The ages and trends of this improved dyke swarm classification for the Pudasjärvi block are then compared with that of other Karelian crustal blocks to define the relative rotation between Pudasjärvi and other blocks. Structural restoration of the Pudasjärvi block indicates that this block underwent a $\sim 30^\circ$ clockwise vertical-axis rotation during the Svecofennian orogeny, a conclusion similar to that reached by Vuollo *et al.* (2000), also based on dyke swarm comparisons. Our analysis is accompanied by whole-rock major and trace element geochemistry to build upon the geochemical characterization efforts initiated in previous studies (i.e., Vuollo and Huhma, 2005; Huhma *et al.*, 2018). The regional trends extrapolated by structurally restoring the Pudasjärvi block to its position prior to Svecofennian reworking, together with the new geochronological data, are then used to refine the reconstruction of Karelia-Kola within the context of supercraton Superia.

2.3 Geological setting

The Karelia-Kola metacraton is the Archean component of the Precambrian Fennoscandian or Baltic shield (**Figure 2-2**; Gaál and Gorbatshev, 1987; Gorbatshev and Bogdanova, 1993; Hölttä *et al.*, 2008). For the purposes of paleogeographic reconstruction, the Karelia and Kola cratons are treated as a single Archean metacraton because of the three magmatic barcode matches at *ca.* 2.5 Ga, 2.45 Ga, and 1.9 Ga (**Figure 2-1**; Smolkin, 1997; Bleeker

Figure 2-2. Map of the Karelia-Kola craton with relevant Archean blocks, geological provinces (names in red), Belomorides (grey text), and Karelia Supergroup with selected schist belts labelled (italics). Geology of the Karelia-Kola craton modified from Hölttä *et al.* (2008); Geological Survey of Finland, GTK (2012); and Lahtinen (2012). The Pudasjärvi block and approximate extent of **Figure 2-3** are outlined by the black rectangle. Inset: divisions of the East European craton. Abbreviations as follows: N - Norway; S - Sweden; Fin - Finland; Rus - Russia, modified from Gorbatshev and Bogdanova (1993).

and Ernst, 2006; Salminen *et al.*, 2014). We do recognize, however, that the Paleoproterozoic history of the Belomorides (located between Karelia and Kola), and microcontinents in Kola, requires some independent movement and re-amalgamation, details of which remain to be resolved and are beyond the scope of this paper with its focus on western Karelia (Mints, 2015a; Lubnina *et al.*, 2016a).

The early history of the Kola and Karelia Archean cratons has been characterized as amalgamated composites of smaller crustal blocks, termed cratonic embryos or microcontinents by (Mints, 2015b), that were generated contemporaneously, but independently, by high-temperature melting of mafic or tonalite-trondjemite-granodiorite (TTG) crust (Mints *et al.*, 2015). Cratonic embryos recognized in Kola include the 2.9-2.7 Ga Murmansk and Inari-Kola microcontinents (Mints, 2015b). Their independent generation is recognized by their differing geology and bounding 2.8 Ga greenstone and sedimentary terranes (Mints *et al.*, 2015). While Murmansk is comprised of relatively undeformed TTG suites, diorite and enderbite, Inari-Kola is a complex microcontinent involving sedimentary and igneous terranes that have been pervasively deformed by multiple Archean and Proterozoic metamorphic events (Hölttä *et al.*, 2008; Mints, 2015b). Mints (2015b) explains that Karelia is also a composite of independently produced microcontinents that grew and merged during the Meso- and Neoarchean. The amalgamation of Kola and Karelia is recorded by *ca.* 2.72 Ga high-pressure granulites in the Belomorian belt, located between the two cratons (Hölttä *et al.*, 2008).

As described by Slabunov *et al.* (2006) and summarized by Hölttä *et al.* (2008), Karelia is subdivided into the Vodlozero, Central Karelia and Western Karelia subprovinces by lithology, structure and ages (Hölttä *et al.*, 2012). The Vodlozero

subprovince is centered on 3.2-3.1 Ga granitoids onto which three generations of greenstone belts were accreted from 3.1-2.6 Ga (Slabunov *et al.*, 2008; Hölttä *et al.*, 2012). The Central Karelian subprovince includes 2.75-2.70 Ga TTG suites, sanukitoids, and greenstone belts (Hölttä *et al.*, 2008). Western Karelia is characterized by the presence of Paleoproterozoic to Neoproterozoic TTG orthogneiss and amphibolite belts, and remnants of ancient greenstone belts (Hölttä *et al.*, 2012). The Western Karelia subprovince comprises four Archean crustal blocks including Pudasjärvi, Iisalmi, Kuhmo and Taivalkoski (**Figure 2-2**; Vuollo and Huhma, 2005) and can be further subdivided into the Lentua, Kuopio, Rautavaara, Iisalmi, Manamansalo, Kalpio, Siurua and Ranua complexes (Hölttä *et al.*, 2012). Boundaries between Archean crustal blocks are defined by the Peräpohja, Kuusamo, Kainuu and Kiiminki Proterozoic schist belts and shear zones, or by their temporally and geochemically distinct lithological assemblages (**Figure 2-2**; Hölttä *et al.*, 2012). Spatially, the Pudasjärvi block is the most isolated Archean block of those located in Western Karelia. Its southwestern and northwestern boundaries are delineated by Paleoproterozoic cover, while the eastern boundary is defined by thrust faults, Paleoproterozoic shear zones and related Kuusamo and Kainuu schist belts (**Figure 2-2**; Hölttä *et al.*, 2008, 2012). With areal dimensions of ~200 km by 165 km, the Pudasjärvi block is divided into the Ranua and Siurua complexes (**Figure 2-3**; Hölttä *et al.*, 2012). Near the centre of the block, the *ca.* 3.5 Ga Siurua complex is the oldest unit in Pudasjärvi and in all of Fennoscandia (Mutanen and Huhma, 2003). The surrounding Ranua complex includes the 2.82-2.80 Ga Oijärvi greenstone belt, 2.82-2.73 Ga TTGs, 2.74-2.73 Ga paragneisses, 2.70-2.62 granites, as well as the Ranua diorite-quartz diorite intrusion dated at *ca.* 2.70 Ga (Mutanen and Huhma, 2003; Hölttä *et al.*, 2012; Huhma *et al.*, 2012).

Figure 2-3. Simplified geological map of the Pudasjärvi block, Western Karelia, showing Paleoproterozoic dykes and sills, *ca.* 2450 Ma layered mafic intrusions (names in italics) and surrounding schist belts (bold italics). Localities A-J from **Table 2-1**. Geology and dyke swarms modified from GTK (2012).

Table 2-1. Sample locations and characteristics of mafic dykes in the Pudasjärvi block.

Map ID	Dyke swarm, age Locality, sample ID	Lat. (°N)	Long. (°E)	Trend (°)	Dip (°)	Width (m)	Age (Ma)	Dating Method	εNd	Magn. susc. avg ±1σ (n)	Visible on Am
<i>Pääjärvi swarm, ca. 2450 Ma</i>											
A	Loljunmaa, 15SD_070	65.9103	25.2021	280	~88	250			-1.2 ¹	0.8±0.1 (10)	No
B	Vengasvaara, A1744	65.4246	26.4098	-	-		2444±4 ¹	U-Pb SHRIMP, Zrc	-5.4 ¹	Not available	Yes
C	16SD_014	65.4911	25.3306	018	90?	20				0.69±0.1 (10)	No
<i>Uolevinlehto swarm, ca. 2400 Ma</i>											
D	Uolevinlehto, 15SD_071	65.6902	27.0607	325	~90	30	ca. 2400; 2386.9±2.4 (min.)	U-Pb TIMS, Bdy	+0.4 ¹	30±3 (5)	No
<i>Tohmajärvi swarm, ca. 2135-2100 Ma</i>											
E	15SD_090	65.4945	25.3163	350	~90	20				1.0±0.3	No
F	Sipojuntti, 15SD_092	65.7310	24.8864	320	-	75	2128.9±1.2	U-Pb TIMS, Bdy	+0.3 ¹	4-40	Yes
G	Myllykangas, 16SD_019	65.5383	25.3329	350	-	30	2135.2+3.6/-3.7 (min.)	U-Pb TIMS, Zrc, Bdy		1.3±0.54 (13)	Yes
<i>Palomaa swarm, ca. 2077 Ma</i>											
H	Palomaa dyke*	65.4618	26.3310	~300	-	30	2077±34 ¹	Sm-Nd*	+1.1 ¹	Not available	Yes

Table 2-1 continues on the next page

Notes: Localities A-J are shown in **Figure 2-3**. Coordinates in WGS1984. Dips reported with a "~" indicate those interpreted from contact-parallel parting planes and/or columnar jointing. Abbreviations: Bdy - baddeleyite; Zrc - chemically abraded zircon; Magn. susc. - Magnetic susceptibility (x10⁻³ SI units). Am: Aeromagnetic map (GTK, 2016). * Sm-Nd age from whole-rock, plagioclase, and pyroxene mineral/internal isochron. Reference: 1. Huhma et al. (2018).

Table 2-1. Continued.

Map ID	Dyke swarm, age Locality, sample ID	Lat. (°N)	Long. (°E)	Trend (°)	Dip (°)	Width (m)	Age (Ma)	Dating Method	ϵNd	Magn. susc. avg $\pm 1\sigma$ (n)	Visible on Am
<i>East-West swarm</i>											
I	15SD_091	65.5168	25.2793	260	~90	25				0.3 $\pm$ 0.1 (10) Chilled margin:	No
J	15SD_101	65.5368	25.2929	275	85N	16				0.3 $\pm$ 0.1 (10), Centre: 0.6 $\pm$ 0.3 (10), max. 20	Yes
C	16SD_015	65.4911	25.4994	270	~90	0.2				5.3 $\pm$ 3.6 (10)	No
G	16SD_020	65.5393	25.3338	280	-	12				22.6 $\pm$ 11.8 (11)	Yes
<i>Paukkajanvaara swarm, ca. 1980 Ma</i>											
-	no samples available										Yes

Notes: Localities A-J are shown in **Figure 2-3**. Coordinates in WGS1984. Dips reported with a "~" indicate those interpreted from contact-parallel parting planes and/or columnar jointing. Abbreviations: Bdy - baddeleyite; Zrc - chemically abraded zircon; Magn. susc. - Magnetic susceptibility ($\times 10^{-3}$ SI units). Am: Aeromagnetic map (GTK, 2016). * Sm-Nd age from whole-rock, plagioclase, and pyroxene mineral/internal isochron. Reference: 1. Huhma et al. (2018).

Hundreds of mafic dykes that are Paleoproterozoic or younger (**Figure 2-3**) cut the basement of the Pudasjärvi block. The most recent digital dyke swarm compilation by the Geological Survey of Finland (GTK) includes 891 polylines of variable orientations representing dykes and dyke segments identified by bedrock mapping programs or interpretation of aeromagnetic data (Vuollo and Huhma, 2005; GTK, 2012, 2016). Despite the high density of dykes within the Pudasjärvi block, only 18% (or 162 polylines) of dykes are classified into either the *ca.* 2450 Ma Pääjärvi, *ca.* 2100 Ma Tohmajärvi or *ca.* 1980 Ma Paukkajanvaara swarms. (For a brief explanation of Finnish locality names the reader is referred to the **List of Finnish Words** on page 29).

2.3.1 Previous dyke swarm classification in the Pudasjärvi block, Western Karelia

2.3.1.1 Pääjärvi swarm (NNE-trending, 2450 Ma)

The *ca.* 2450 Ma Pääjärvi swarm is the oldest set of Paleoproterozoic dykes and sills recognized in Pudasjärvi (Vuollo and Huhma, 2005). The northern extent of the Pudasjärvi block hosts the 2422 ± 52 Ma Loljunmaa gabbro-norite dyke. Trending 280° , the Loljunmaa dyke is at least 250 m wide and is regarded as a feeder to the *ca.* 2.45 Ga Penikat layered mafic intrusion (locality A in **Table 2-1** and **Figure 2-3**; Vuollo and Huhma, 2005; Huhma et al., 2018). Their genetic relationship is supported by similar Sm-Nd whole-rock data of Loljunmaa to the plagioclase and pyroxene mineral isochron of the Penikat intrusion (Huhma *et al.*, 2018). The 2436 ± 8 Ma Kemi intrusion, Suhanko and Portimo complexes (Narkaus and adjacent intrusive bodies) are other layered mafic intrusions at the northern margin of Pudasjärvi that have been correlated with Pääjärvi mafic magmatism (**Figure 2-3**; Huhma *et al.*, 2018). Near the centre of the block is the 2444 ± 4 Ma Vengasvaara

intrusion (locality B in **Table 2-1** and **Figure 2-3**; Huhma *et al.*, 2018). Although the dated sample from the Vengasvaara intrusion is from drill core (Huhma *et al.*, 2018), the associated prominent aeromagnetic signature suggests it has a north-south trend.

2.3.1.2 Uolevinlehto swarm (NW-trending, previously linked with Pääjärvi swarm)

Earlier studies grouped the Uolevinlehto dyke within the Pääjärvi swarm (locality D in **Table 2-1** and **Figure 2-3**; Vuollo and Huhma, 2005; Huhma *et al.*, 2018). The medium-grained 30 m-wide Uolevinlehto dyke (**Figure 2-4a**) trends 325° and, based on observation of parting planes, retains a near vertical dip. Vuollo and Huhma (2005) reported a minimum age of 2378 Ma based on one multigrain baddeleyite fraction analyzed by ID-TIMS, approximately correlating with the 2.45 Ga Pääjärvi swarm, but also consistent with a younger age. Because the multigrain baddeleyite fraction plots on a discordia line between 2.44 and 1.88 Ga, Huhma *et al.* (2018) did not reject a possible connection to the Pääjärvi event. However, herein it is shown to be a different swarm.

2.3.1.3 Tohmajärvi swarm (NW-trending, *ca.* 2130-2100 Ma), and Palomaa swarm (WNW-trending, *ca.* 2077 Ma)

The NW-trending Tohmajärvi dykes in the Pudasjärvi block are dated at the Sipojuntti locality (locality F in **Table 2-1** and **Figure 2-3**). Perttunen and Vaasjoki (2001) report an upper intercept U-Pb age of 2118±14 Ma based on air-abraded multigrain zircon ID-TIMS analyses. The 320°-trending, medium- to coarse-grained Sipojuntti dyke is at least 75 m wide and is faintly visible on aeromagnetic maps (Perttunen and Vaasjoki, 2001; Vuollo

and Huhma, 2005; GTK, 2016). Also included in the Tohmajärvi swarm classification by Vuollo and Huhma (2005) is the 290-300°-trending Palomaa dyke that has a preliminary age of 2077 ± 34 Ma based on a Sm-Nd mineral and whole-rock internal isochron (locality H in **Table 2-1** and **Figure 2-3**; Huhma *et al.*, 2018).

2.3.1.4 Paukkajanvaara swarm (NNW-trending, *ca.* 1980 Ma)

The NNW-trending Paukkajanvaara dykes are identified in the Pudasjärvi block by their prominent high magnetic signatures on aeromagnetic maps (Vuollo and Huhma, 2005). Few are exposed and no dykes belonging to this swarm have been successfully sampled for geochronology or geochemistry (Vuollo and Huhma, 2005; GTK, 2012).

2.4 Methods

2.4.1 U-Pb ID-TIMS geochronology

Medium- to coarse-grained ~10 kg samples from the Uolevinlehto, Sipojuntti and Myllykangas localities (**Figure 2-3**) were crushed and a concentrate of heavy, non-magnetic minerals was prepared for each sample using a Wilfley table, magnetic separator and methylene iodide. Baddeleyite and zircon were handpicked from this heavy mineral separate under a binocular microscope. Additional baddeleyite separation using the Wilfley table, following the method of Söderlund and Johansson (2002), was required for the Uolevinlehto sample. Mineral processing and U-Pb analysis by ID-TIMS methods of the three samples were completed at the Jack Satterly Geochronology Laboratory at the

University of Toronto.

U-Pb analysis was by isotope dilution thermal ionization mass spectrometry methods (ID-TIMS). Prior to analysis, zircon crystals were thermally annealed and chemically etched (chemical abrasion), which has the advantage of penetrative removal of alteration zones where Pb loss has occurred and which generally improves concordance (Mattinson, 2005). This pre-treatment involves placing zircon grains in a muffle furnace at $\sim 900^{\circ}\text{C}$ for ~ 24 -60 hours to repair radiation damage and anneal the crystal lattice, followed by a modified single-step partial dissolution procedure in ~ 0.10 ml of $\sim 50\%$ HF and 0.020 ml 7N HNO_3 in Teflon dissolution vessels at 195°C for 2-12 hours. Zircon and baddeleyite were rinsed with HNO_3 at room temperature prior to dissolution. A ^{205}Pb - ^{235}U spike was added to the Teflon dissolution capsules during sample loading. Zircon and baddeleyite were dissolved using ~ 0.10 ml of concentrated HF acid and ~ 0.02 ml of 7N HNO_3 at 195°C for 3-5 days. All samples were dried to a precipitate and re-dissolved in ~ 0.15 ml of 3N HCl overnight (Krogh, 1973). U and Pb were isolated from the zircon solution using $50\ \mu\text{l}$ anion exchange columns using HCl , deposited onto outgassed rhenium filaments with silica gel (Gerstenberger and Haase, 1997), and analyzed with a VG354 mass spectrometer using a Daly detector in pulse counting mode. Corrections to the ^{206}Pb - ^{238}U ages for initial ^{230}Th disequilibrium in the zircon and baddeleyite have been made assuming a Th/U ratio in the magma of 4.2. All common Pb was assigned to procedural Pb blank for zircon. Dead time of the measuring system for Pb and U was 16 and 14 ns, respectively. The mass discrimination correction for the Daly detector is constant at 0.05% per atomic mass unit. Amplifier gains and Daly characteristics were monitored using the SRM 982 Pb standard. Thermal mass discrimination corrections are 0.10% per atomic mass unit for both Pb and

U. Decay constants are those of Jaffey et al. (1971). All age errors quoted in the text and table, and error ellipses in the concordia diagrams are given at the 95% confidence interval. VG Sector software was used for data acquisition. In-house data reduction software in Visual Basic by D.W. Davis was used. Plotting and age calculations were done using Isoplot 3.00 (Ludwig, 2003). Results of standards (R33, OG1 and EARTHTIME 2.0 Ga synthetic solution of Black et al., 2004; Condon et al., 2015. and Stern et al., 2019) analyzed over the duration of this study are presented in **Appendix C**.

2.4.2 Geochemistry

Whole-rock major and trace element geochemical data were acquired for 10 mafic dykes within the Pudasjärvi block and two dykes from the Kuhmo and Taivalkoski blocks of Western Karelia. Where possible, samples for geochemistry were collected from the freshest dyke rock at ~0.5-1 m in from the chilled margin (samples were not collected directly from the margin of the dyke to avoid the risk of chemical interaction with the host rocks). Samples were crushed in a Mn-steel jaw crusher and pulverized in an agate mill at Carleton University. At least 5 g of each pulverized sample was sent to ALS Laboratories (www.alsglobal.com) for analysis of 13 major oxides, and 42 minor and trace elements by inductively coupled plasma-atomic emission spectrometry (ICP-AES) and inductively coupled plasma-mass spectrometry (ICP-MS). Samples underwent lithium borate fusion before dissolution for trace element analysis by ICP-MS. Note that samples were only included in the geochemical comparison that had major element oxide totals greater than 97% (including loss on ignition).

2.5 Results

2.5.1 Field observations

In the Pudasjärvi block, ten dykes were sampled, including three previously documented localities and five new localities. Two of these new localities include two dykes at each site (**Table 2-1; Figure 2-3**). The *ca.* 2.45 Ga Loljunmaa, *ca.* 2.40 Ga Uolevinlehto and 2.13 Ga Sipojuntti localities have been the subject of previous geochronological and geochemical studies (**Figure 2-3** localities A, D, and F, respectively; Perttunen and Vaasjoki, 2001; Vuollo and Huhma, 2005; Huhma *et al.*, 2018).

Dyke sites 16SD_014 and 16SD_015 are found together in a forested area north of a rural gravel road near the village of Väli-Olhava (locality C in **Table 2-1** and **Figure 2-3**). Dyke 16SD_014 is 15-20 m wide with a 018° trend. Although only one contact was observed, the 15-20 m width is estimated due to the disappearance of outcrop on the other side of the dyke. Cross-cutting this larger dyke (16SD_014) is the 20 cm-wide E-W-trending dyke 16SD_015 (**Figure 2-4b**). The younger dyke (16SD_015) has a porphyritic texture with plagioclase phenocrysts up to 3 mm long in an aphanitic groundmass.

Dyke site 15SD_090 is a roadside outcrop along the east side of Highway 4 (Kemintie road) ~1.2 km north of the junction from Nyby (locality E in **Table 2-1** and **Figure 2-3**). The 350° trend of dyke 15SD_090 is inferred from horizontal columnar jointing in lieu of exposed contacts. From outcrop exposure, the estimated width of this fine- to medium-grained diabase dyke at least 20 m wide.

Figure 2-4. Outcrop characteristics and field relationships of some observed dykes in the Pudasjärvi block. a) Average medium-grained, intergranular, texture of (meta)diabase in the Uolevinlehto dyke, 15SD_071. b) The cross-cutting relationship between the older 018°-trending dyke 16SD_014, Pääjärvi swarm, and younger East-West trending dyke 16SD_015 with columnar jointing. c) Roadside outcrop of dyke 15SD_091 with columnar jointing outlined (white long-dashed lines) and the inferred trend marked by the white dotted line. d) Contact of dyke 15SD_101 with Archean basement.

The Myllykangas (16SD_019) and 16SD_020 dyke sites are found together in a craggy, sparsely forested area ~4.20 km along Pysäkkitie road, and later Kivimaantie road, starting from the Highway 4 (Kemintie road) intersection (locality G in **Table 2-1** and **Figure 2-3**). Myllykangas has a 350° trend and can be followed for 100 m. Despite contacts not being observed, its minimum width is estimated at 30 m based on exposed outcrop. With a green-weathered colour, the dyke is generally medium-grained but does include

rare coarser-grained patches near its presumed centre. Dyke 16SD_020 is 12-15 m wide, near vertical dyke with a 270° trend and is distinguished from 16SD_019 by its brown-weathering and higher magnetic susceptibility (**Table 2-1**). Contacts between the two dykes were not located, but both dykes are visible on aeromagnetic maps.

Dyke site 15SD_091 is 3 km north of 15SD_090 following Highway 4 (locality I **Table 2-1** and **Figure 2-3**). No contacts are exposed but jointing outlining large horizontal columns indicate a trend of $\sim 260^\circ$ (**Figure 2-4c**). The dyke has a uniform fine-grained texture and is at least 30 m wide.

Dyke site 15SD_101 is located in an unnamed quarry ~ 2.75 km along Pysäkkitie road, and later Kivimaantie road, starting from the Highway 4 (Kemintie road) intersection (locality J in **Table 2-1** and **Figure 2-3**). The dyke is at least 16 m wide and has a near-vertical dip. It has a consistent medium- to fine-grained plagioclase-phyric texture including rare pegmatoidal veins. Contacts of the dyke against Archean basement, and an elevated magnetic susceptibility, make it easily visible on aeromagnetic maps, and indicate that the dyke has a 270° trend (**Figure 2-4d**).

2.5.2 Petrographic characteristics

Major rock-forming minerals in mafic dykes in the Pudasjärvi block are plagioclase (40-60%), clinopyroxene (20-35%), primary and secondary hornblende (10-20%), opaque minerals such as magnetite, ilmenite and titanomagnetite (1-5%), and accessory phases of quartz, apatite, titanite $\pm$ pyrite, pyrrhotite, chalcopyrite, pentlandite, sphalerite, orthopyroxene, baddeleyite, and zircon. Common textures are intergranular to subophitic and ophitic. The Uolevinlehto dyke is a well-preserved example that has intergranular to

subophitic texture (**Figure 2-5**). Textures of other, more pervasively altered dykes are shown in **Figure 2-6** to **Figure 2-9**. Plagioclase crystals range from sub- to euhedral with carlsbad, pericline and albite twinning, and rarely have normal zoning. While the average plagioclase composition ranges from bytownite to andesine (An₈₀₋₄₀), albite (An₁₀) is also present as a secondary alteration phase found in association with epidote (Vuollo *et al.*, 1992; Stepanova *et al.*, 2014, 2015). Orthopyroxene is a minor phase specific to the Uolevinlehto dyke and, similar to clinopyroxene, occupies spaces between plagioclase crystals (**Figure 2-5**). Interstices filled with quartz and feldspar occasionally show graphic intergrowth textures, particularly in coarser-grained samples. When present, baddeleyite and zircon are adjacent to amphibole, clinopyroxene or plagioclase grain boundaries, or included within interstitial quartz blebs (**Figure 2-5** to **Figure 2-7**). Alteration intensity varies from minor to pervasive across all swarms. Regardless of intensity, alteration consistently includes sericite and saussuritisation of plagioclase, and chlorite as well as secondary amphibole alteration over pyroxene and primary amphibole (**Figure 2-5** to **Figure 2-9**).

Figure 2-5. Petrographic details of the Uolevinlehto dyke 15SD_071, *ca.* 2400 Ma Uolevinlehto swarm. a) Intergranular to subophitic texture between plagioclase (Plag) and clinopyroxene (Cpx) with minor chlorite alteration, plane polarized light. b) As in a, crossed nicols. c) Backscattered electron image of baddeleyite (in white box) hosted by a quartz bleb within amphibole (Amph) and plagioclase. d) Close-up of baddeleyite (Bdy) grain from c showing a subhedral bladed morphology included in a quartz (Qtz) bleb; very minor secondary zircon (Zrc) overgrowth (light grey) on bottom corner of crystal. e) Single grain baddeleyite fraction Bdy1. f) Baddeleyite fraction Bdy2 including 10 crystals and fragments.

Figure 2-6. Petrographic details of the Myllykangas dyke 16SD_019, 2135.2+3.6/-3.7 Ma, Tohmajärvi swarm. a) Photomicrograph of fine-grained-intergranular plagioclase (Plag) and amphibole (Amph) overprinted by saussurite alteration including actinolite (Act) under plane polarized light. b) As in a, crossed nicols. c) Backscattered electron image of zircon (Zrc); in white box) adjacent to pyrite (Py), plagioclase, apatite (Ap) and secondary actinolite (Act). d) Close-up of poorly preserved zircon grain in c. e-f) Backscattered electron image of baddeleyite (Bdy) with substantial secondary zircon overgrowth hosted between amphibole and plagioclase grains. g) Image of single grain baddeleyite fractions Bdy1, Bdy2, and Bdy3. h) Image of zircon after chemical abrasion Zrc1, Zrc2, and Zrc3.

Figure 2-7. Petrographic details of the pervasively altered Sipojuntti dyke 15SD_092, 2128.9 ± 1.2 Ma, Tohmajärvi swarm. a) Saussuritised plagioclase (Plag) and chlorite on amphibole (Amph) of equigranular-fine-grained gabbro under plane polarized light. b) As in a, crossed nicols. c) Backscattered electron image of two partially metamict zircon (Zrc) crystals surrounded by plagioclase, amphibole and secondary biotite (Bt). d) Close-up of two partially metamict zircon crystals in c. e) Backscattered electron image of baddeleyite (Bdy) with zircon overgrowth hosted in amphibole in the vicinity of plagioclase and titanite (Ttn). f) Single grain baddeleyite fractions Bdy1, Bdy2, and Bdy4.

Figure 2-8. Petrographic details of selected dykes in the Pudasjärvi block with pervasive alteration, chlorite on amphibole (Amph) and clinopyroxene (Cpx), saussuritisation of plagioclase (Plag). Left panels: plane polarized light; right panels: crossed nicols. a-b) Intergranular texture in the Loljunmaa dyke 15SD_070, *ca.* 2450 Ma Pääjärvi swarm. c-d) Preserved subophitic texture with plagioclase embedded in clinopyroxene in dyke 16SD_014, *ca.* 2450 Ma Pääjärvi swarm. e-f) Subophitic texture with plagioclase embedded in clinopyroxene in dyke 15SD_090, *ca.* 2135-2100 Ma Tohmajärvi swarm.

Figure 2-9. Petrographic details of selected dykes within the East-West swarm. Left: plane polarized light; right: crossed nicols. a-b) Dyke 15SD_091 with a fine-grained-subophitic texture with saussuritised plagioclase embedded between mafic crystals completely replaced by secondary chlorite and amphibole. c-d) Dyke 15SD_101 ophitic texture with pervasive chlorite alteration on mafic minerals embedded in saussuritised plagioclase. e-f) Ophitic texture in dyke 16SD_020.

2.5.3 Geochronology

2.5.3.1 Uolevinlehto dyke, 15SD_071

Mineral separation of medium- to coarse-grained gabbro from the interior of this dyke yielded a limited supply of small baddeleyite crystals and crystal fragments (average length $\sim 40\ \mu\text{m}$, max. $60\ \mu\text{m}$; **Figure 2-5**). One larger single grain, two smaller single grains and two multigrain fractions (2 and 10 grains/fragments, respectively) were analyzed (**Table 2-2**). All yield discordant, scattered data points due to variable lead loss or mixing of two components (i.e., magmatic baddeleyite and younger zircon overgrowth). The three least discordant results, Bdy1, Bdy2 and Bdy3 provide an imprecise upper intercept date of $2422 \pm 40/-70\ \text{Ma}$ and a lower intercept of $1590 \pm 290/-230\ \text{Ma}$ (**Figure 2-10a**). The result of a previous baddeleyite fraction reported by Huhma *et al.* (2018) plots on this regression line. The $^{207}\text{Pb}/^{206}\text{Pb}$ date of the least discordant single grain baddeleyite fraction indicates a minimum crystallization age of $2386.9 \pm 2.4\ \text{Ma}$. Given our experience with variably discordant baddeleyites, with analyses trailing off to the left in the concordia diagram as their discordance increases (i.e., indicating not only more but also earlier open system behaviour), the $^{207}\text{Pb}/^{206}\text{Pb}$ date of the least discordant fraction is likely a closer estimate for the age of crystallization, albeit a minimum, than the upper intercept date resulting from the more inclusive regression (Bleeker, 2014). More data would be required but we may interpret the age of this dyke as *ca.* 2400 Ma with an estimated uncertainty of $\pm 10\text{-}12\ \text{Ma}$.

Table 2-2. U-Pb isotopic data for the Uolevinlehto, Myllykangas and Sipojuntti mafic dykes from the Pudasjärvi block.

Fraction (analysis number)	Weight (μg)	U (ppm)	$\frac{\text{Th}}{\text{U}}$	Pbtot (pg)	Pbcom (pg)	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$ Meas.	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$ Meas.	2 σ	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	2 σ	Error Corr.
<i>Uolevinlehto, 15SD_071</i>											
Bdy1 (sk38p189)	0.2	321	0.13	28.9	0.5	3582	9.340	0.031	0.4409	0.0013	0.899
Bdy2, 10 (sk38p240)	0.2	788	0.22	70.8	1.4	3072	9.004	0.041	0.4311	0.0018	0.946
Bdy3, 2 (sk39p154)	0.1	166	0.21	3.6	0.7	348	8.608	0.158	0.4209	0.0055	0.806
Bdy4 (sk39p154)	0.1	355	0.10	7.2	0.8	579	7.913	0.088	0.4036	0.0026	0.738
Bdy5 (sk39p178)	0.1	66	0.05	2.8	0.7	295	8.594	0.490	0.4285	0.0088	0.685
<i>Myllykangas, 16SD_19B</i>											
Zrc1 (sk39p187)	0.2	122	0.90	10.0	0.2	2916	6.954	0.040	0.3798	0.0020	0.933
Zrc2 (sk39p163)	0.1	148	0.75	3.2	0.5	364	6.639	0.149	0.3714	0.0056	0.778
Zrc3, hzy (sk39p189)	0.9	60	0.73	17.9	0.6	1824	5.290	0.024	0.3045	0.0011	0.808
Bdy1, clr (sk38p237)	0.1	217	0.26	7.8	4.6	124	5.894	0.338	0.3492	0.0036	0.856
Bdy2, clr (sk39p162)	0.1	167	0.07	5.6	0.7	514	6.091	0.081	0.3431	0.0024	0.699
Bdy3 (sk38p236)	0.3	172	0.23	17.4	0.8	1394	5.921	0.032	0.3283	0.0013	0.779
<i>Sipojuntti, 15SD_092</i>											
Bdy1, clr (sk38p199)	0.3	1239	0.03	140.4	7.1	1310	7.028	0.038	0.3850	0.0014	0.703
Bdy2, clr (sk38p201a)	0.3	1154	0.04	129.3	0.6	14549	6.934	0.022	0.3798	0.0011	0.949
Bdy3 (sk39p183)	0.4	985	0.02	131.7	0.6	13708	6.918	0.017	0.3794	0.0008	0.937
Bdy4 (sk39p161)	0.2	1111	0.11	83.9	0.5	9855	6.845	0.017	0.3776	0.0008	0.918

Table-2 continues on the following page

Notes: Fractions are single grain analyses unless followed by a value indicating number of grains in the fraction; Zrc-zircon, Bdy-baddeleyite. Clr-clear, hzy-hazy.

Th/U calculated from radiogenic $^{208}\text{Pb}/^{206}\text{Pb}$ ratio and $^{207}\text{Pb}/^{206}\text{Pb}$ age assuming concordance.

Pbtot is amount of Pb in picograms excluding blank. Pbcom is total amount of common Pb in picograms; assigned to the isotopic composition of laboratory blank.

$^{206}\text{Pb}/^{204}\text{Pb}$ corrected for fractionation and common Pb in the spike. Correction for ^{230}Th disequilibrium in $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ 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 $^{207}\text{Pb}/^{206}\text{Pb}$ age.

Uranium decay constants are from Jaffey *et al.* (1971).

Table 2-3. Continued.

Fraction (analysis number)	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	2 σ	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$ Age (Ma)	2 σ	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$ Age (Ma)	2 σ	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$ Age (Ma)	2 σ	Disc. (%)
<i>Uolevinlehto, 15SD_071</i>									
Bdy1 (sk38p189)	0.15365	0.00023	2354.5	5.7	2371.9	3.1	2386.9	2.5	1.6
Bdy2, 10 (sk38p240)	0.15147	0.00023	2310.8	8.1	2338.5	4.2	2362.6	2.6	2.6
Bdy3, 2 (sk39p154)	0.14834	0.00164	2264.4	25.0	2297.4	16.7	2326.9	19.0	3.2
Bdy4 (sk39p154)	0.14221	0.00110	2185.5	12.0	2221.2	10.1	2254.3	13.4	3.6
Bdy5 (sk39p178)	0.14546	0.00659	2299.0	39.7	2295.9	51.3	2293.2	77.0	-0.3
<i>Myllykangas, 16SD_19B</i>									
Zrc1 (sk39p187)	0.13279	0.00028	2075.3	9.4	2105.5	5.2	2135.2	3.7	3.3
Zrc2 (sk39p163)	0.12966	0.00185	2035.8	26.5	2064.6	19.7	2093.4	25.1	3.2
Zrc3, hzy (sk39p189)	0.12601	0.00034	1713.4	5.2	1867.2	3.9	2043.1	4.8	18.4
Bdy1, clr (sk38p237)	0.12243	0.00599	1930.7	17.0	1960.4	50.0	1991.9	87.6	3.6
Bdy2, clr (sk39p162)	0.12876	0.00125	1901.6	11.6	1989.0	11.6	2081.1	17.1	10.0
Bdy3 (sk38p236)	0.13081	0.00045	1830.0	6.1	1964.3	4.7	2108.9	6.0	15.2
<i>Sipojuntti, 15SD_092</i>									
Bdy1, clr (sk38p199)	0.13239	0.00051	2099.6	6.4	2115.0	4.8	2129.9	6.8	1.7
Bdy2, clr (sk38p201a)	0.13240	0.00014	2075.5	5.2	2103.0	2.9	2130.0	1.8	3.0
Bdy3 (sk39p183)	0.13224	0.00012	2073.5	3.7	2100.9	2.2	2128.0	1.6	3.0
Bdy4 (sk39p161)	0.13149	0.00013	2064.9	3.7	2091.6	2.2	2118.0	1.8	2.9

Notes: Fractions are single grain analyses unless followed by a value indicating number of grains in the fraction; Zrc-zircon, Bdy-baddeleyite. Clr-clear, hzy-hazy.

Th/U calculated from radiogenic $^{208}\text{Pb}/^{206}\text{Pb}$ ratio and $^{207}\text{Pb}/^{206}\text{Pb}$ age assuming concordance.

Pbtot is amount of Pb in picograms excluding blank. Pbcom is total amount of common Pb in picograms; assigned to the isotopic composition of laboratory blank.

$^{206}\text{Pb}/^{204}\text{Pb}$ corrected for fractionation and common Pb in the spike. Correction for ^{230}Th disequilibrium in $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ 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 $^{207}\text{Pb}/^{206}\text{Pb}$ age.

Uranium decay constants are from Jaffey *et al.* (1971).

Figure 2-10. U-Pb concordia diagrams for dykes in the Pudasjärvi block. a) Uolevinlehto dyke sample 15SD_071. The complex discordia patterns of multigrain baddeleyite fractions define an upper intercept date (dashed regression line) of 2422+40/-70 Ma. A minimum date of 2386.9±2.4 Ma is constrained by the least discordant and most precise datum (Bdy1) with the solid regression line anchored at 0 Ma. A single baddeleyite fraction (light grey ellipse) from Huhma et al. (2018) is also shown and plots along the regression line defined by other fractions in this study. b) Myllykangas dyke, sample 16SD_019B. Single grain zircon (dark grey ellipses) and baddeleyite (white ellipses) show discordant and scattered Pb loss. An upper intercept at 2149.7+6.0/-6.1 Ma is calculated from the most and least discordant zircon fractions and is considered a maximum date. The $^{207}\text{Pb}/^{206}\text{Pb}$ age of the least discordant zircon constrains a minimum age of 2135.2+3.6/-3.7 Ma. c-d) Sipojuntti dyke, sample 15SD_092. c) Large multigrain air-abraded zircon fractions (dark grey ellipses) of Perttunen and Vaasjoki (2001) provide an upper intercept of 2118±14 Ma (grey dashed line) onto which fraction Bdy4 of this study also plots. d) The three least discordant data (white ellipses) plot along a regression line anchored at 0 Ma with an upper intercept age of 2128.9±1.2 Ma.

2.5.3.2 Myllykangas dyke, 16SD_019B

Zircon and baddeleyite were separated from a partially pegmatitic sample from the interior of this 350°-trending dyke. A population of ~50 subhedral, tabular zircon grains ranging in length from 50 to 200 µm were handpicked. Baddeleyite crystals and fragments, many of which were encrusted by secondary zircon overgrowth, were abundant and ranged between 50 and 100 µm in length. Single grain zircon and baddeleyite analyses yielded variably discordant results (**Table 2-2; Figure 2-6; Figure 2-10b**). An upper intercept date calculated using the more precise but discordant zircon analyses (Zrc1 and Zrc3) is 2149.7+6.0/-6.1 Ma, with a lower intercept of *ca.* 644+36/-38 Ma. The age of 2149.7+6.0/-6.1 Ma likely represents a maximum estimate of the age of crystallization. A minimum age of crystallization at 2135.2+3.6/-3.7 Ma is provided by the $^{207}\text{Pb}/^{206}\text{Pb}$ age of the least discordant fraction Zrc1. Because of the scattered lead loss pattern in both the baddeleyite and zircon results, this minimum age of 2135.2+3.6/-3.7 Ma is the best estimate for the age of crystallization of the Myllykangas dyke (locality G in **Table 2-1** and **Figure 2-3**).

2.5.3.3 Sipojuntti dyke, 15SD_092

Zircon and baddeleyite were recovered from the coarse-grained gabbro collected from the Sipojuntti dyke (locality F in **Table 2-1** and **Figure 2-3**). Zircon crystals ranged in size from 50 to 200 µm and included larger, blocky and fragmented grains, as well as smaller subhedral translucent (not clear but hazy) crystals. A selection of the clearest zircon crystals were thermally annealed and briefly etched (1 hour) but this did not produce usable material (**Figure 2-7**). Baddeleyite crystals with blocky, subhedral morphologies, varied between clear and light to dark brown (**Figure 2-7**). Four single grain baddeleyite fractions

yielded 1.7-3.0% discordant results (**Table 2-2; Figure 2-10c-d**). Anchored at 0 Ma, a regression through the three least-discordant fractions (Bdy1, Bdy2, and Bdy3) yield an upper intercept date at 2128.9 ± 1.2 Ma, which we consider the best estimate for the crystallization age of the Sipojuntti dyke (**Figure 2-10d**). Baddeleyite fraction Bdy4 plots well to the left, yet is collinear on the regression line of Perttunen and Vaasjoki (2001), defined by data for eight discordant multigrain air-abraded zircon grains with an upper age intercept at 2118 ± 14 Ma and a lower intercept at 257 ± 190 Ma (**Figure 2-10c**). The scatter between fraction Bdy4 and others of this study, and the results of Perttunen and Vaasjoki (2001), can be attributed to Pb loss in zircon and baddeleyite, and to post-magmatic growth of zircon on baddeleyite (Heaman and LeCheminant, 1993; Rioux *et al.*, 2010; Söderlund *et al.*, 2013).

2.5.4 Geochemistry

Whole-rock major and trace element geochemistry is reported for 12 dykes from the Pudasjärvi, Kuhmo and Taivalkoski blocks (**Table 2-4**). All dykes are classified as tholeiitic basalt to basaltic andesite on total alkali silica and Nb/Y versus Zr/Ti diagrams (**Figure 2-11** and **Figure 2-14**). On an Al-Fe_{tot}+Ti-Mg ternary diagram (Jensen, 1976), the most dykes plot in the high-Fe tholeiitic basalt field (**Figure 2-11**). Outliers from this main cluster include one dyke from the East-west swarm, which plots in the high Mg-tholeiitic field and three dykes belonging to either the Uolevinlehto or Pääjärvi swarms, which plot in the komatiitic basalt field (**Figure 2-11**). Dykes with high Fe-tholeiitic compositions show weak correlations between Al₂O₃, TiO₂, MgO, FeO_{tot} wt. % and no correlation with

Figure 2-11. Major element classification of the *ca.* 2450 Ma Pääjärvi, *ca.* 2400 Ma Uolevinlehto, *ca.* 2135-2100 Ma Tohmajärvi, and East-West dyke swarms. Top: Total alkali silica classification diagram (Le Bas et al., 1986). Bottom: Al-Fe_{tot}+Ti-Mg ternary classification diagram (Jensen, 1976).

CaO wt. % and P₂O₅ wt. % against SiO₂ wt. % (**Figure 2-12**).

Dykes classified as komatiitic basalt (Loljunmaa, Viianki, and dyke 15SD_014) on the Jensen diagram (**Figure 2-11**) have lower Al₂O₃ and FeO_{tot} wt. % values and higher MgO wt. % values compared to the main population (**Figure 2-12** and **Figure 2-13**). These dykes have elevated Mg numbers between 71 and 77 – values that are significantly higher than other dykes in the Pudasjärvi block, which generally range from 43 to 59 (**Figure 2-14b**). Despite variable Mg numbers, SiO₂ wt. % is comparable for all dykes (**Figure 2-12** and **Figure 2-14**). The Th/Nb_{PM} ratio (PM: normalized to primitive mantle; Sun and McDonough, 1989) compared with Mg number allows for the further distinction of two populations of dykes: those with high Th/Nb_{PM} ratios (>4) coupled with high Mg numbers and those with low Th/Nb_{PM} ratios (<4) coupled with lower Mg numbers (**Figure 2-14b**).

Similarities across dyke swarms continue into the multi-element patterns with flat heavy rare earth element (HREE) slopes, slightly elevated light rare earth elements (LREEs) indicated by shallow negative slopes from La to Gd, negative Nb anomalies of variable magnitude, and variable enrichment in Th with respect to La (**Figure 2-14c** and **f**). In particular, the trace element ratios Nb/La_{PM}, La/Sm_{PM}, Tb/Yb_{PM} are similar in nearly all dykes. The Myllykangas locality, however, is distinct from the general population of dykes (**Figure 2-14e**), locality G in **Table 2-1** and **Figure 2-3**). With lower absolute abundances of trace elements and a shallow positive slope, Myllykangas is neither enriched in Th nor depleted in Nb relative to La (**Figure 2-14e**).

Figure 2-12. Major element oxides Al_2O_3 , CaO , TiO_2 , P_2O_5 , MgO and Fe_{tot} against SiO_2 for the ca. 2450 Ma Pääjärvi, ca. 2400 Ma Uolevinlehto, ca. 2135-2100 Ma Tohmajärvi, and East-West dyke swarms. Two-sigma error is based on reproducibility of the 10-LT-05 basalt standard (see methods and **Table 2-4**). Errors not shown on the Y-axis when 2σ is less than 0.2 (or less than the thickness of the SiO_2 error bar).

Figure 2-13. Major element oxides Al₂O₃, CaO, TiO₂, P₂O₅, Na₂O, K₂O, SiO₂ and Fe_{tot} against Mg number. Two-sigma error is based on reproducibility of the 10-LT-05 basalt standard (see methods and **Table 2-4**). Errors not shown where 2σ is less than 0.2.

Figure 2-14. Geochemical characteristics of the *ca.* 2450 Ma Pääjärvi, *ca.* 2400 Ma Uolevinlehto, *ca.* 2135-2100 Ma Tohmajärvi, and East-West dyke swarms. a) Zr/Ti versus Nb/Y diagram (Pearce, 1996). b) Th/Nb versus Mg number (PM: normalized to primitive mantle; Sun and McDonough, 1989). c-f) Multi-element diagrams normalized to primitive mantle: c) Pääjärvi, *ca.* 2450 Ma. d) Uolevinlehto, *ca.* 2400 Ma. e) Tohmajärvi, *ca.* 2135-2100 Ma and f) East-West dyke swarms.

Table 2-4. Whole-rock major and trace element geochemistry for dykes in the Pudasjärvi, Taivalkoski and Kuhmo block.

Swarm, locality, sample ID	Geochemical	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	CaO	MgO	Na ₂ O	K ₂ O	Cr ₂ O ₃	TiO ₂	MnO	P ₂ O ₅	SrO	BaO	LOI	Total
<i>Detection limit</i>	<i>group</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>
<i>Pääjärvi swarm, ca. 2450 Ma</i>																
Loljunmaa, 15SD_070 ^A	2	51.00	11.10	10.80	9.38	12.15	1.64	0.50	0.17	0.70	0.17	0.10	0.03	0.02	2.20	99.96
16SD_014 ^B	2	51.50	8.49	9.73	7.45	15.50	1.83	0.62	0.24	0.63	0.16	0.08	0.02	0.03	3.20	99.48
<i>Uolevinlehto swarm, ca. 2400 Ma</i>																
Uolevinlehto, 15SD_071 ^A	1	52.80	14.50	13.05	9.80	5.80	2.45	0.57	0.02	1.06	0.19	0.13	0.02	0.02	-0.53	99.88
WD dyke (15SD_073) ^A	1	48.00	13.50	15.35	9.40	6.35	2.09	0.49	0.03	2.40	0.22	0.31	0.02	0.01	0.57	98.74
Viianki dyke (15SD_084A, 2m from contact) ^A	2	51.40	11.85	10.80	8.48	12.95	1.81	0.59	0.18	0.60	0.17	0.08	0.02	0.02	0.38	99.33
Viianki dyke (15SD_084B, center) ^A	2	52.70	10.90	10.75	8.26	14.70	1.62	0.45	0.21	0.49	0.17	0.06	0.02	0.02	0.15	100.50
<i>Tohmajärvi swarm, ca. 2135-2100 Ma</i>																
15SD_090 ^A	1	51.00	13.15	14.00	8.22	5.61	3.00	1.42	0.03	1.81	0.21	0.26	0.02	0.04	1.07	99.84
Sipojuntti, 15SD_092 ^A	1	46.40	12.55	14.05	9.13	8.59	2.69	0.70	0.08	2.29	0.18	0.16	0.04	0.02	2.02	98.90
Myllykangas, 16SD_019A ^B	3	49.40	13.15	15.35	10.70	6.82	1.96	0.27	0.03	1.04	0.25	0.09	0.01	0.01	0.97	100.05
Myllykangas, 16SD_019A* ^B	3	49.30	13.05	15.80	10.95	6.89	1.94	0.29	0.03	1.01	0.25	0.08	0.01	0.01	1.00	100.61
Myllykangas, 16SD_019B ^B	3	55.30	12.25	12.85	8.23	5.01	3.51	0.20	<0.01	1.66	0.19	0.04	0.01	<0.01	0.93	100.18
<i>East-West swarm</i>																
15SD_091 ^A	1	54.30	14.45	9.52	9.14	6.32	2.31	0.92	0.05	0.61	0.16	0.08	0.02	0.03	1.39	99.30
15SD_101B ^A	1	50.10	13.35	16.45	10.40	5.76	2.37	0.46	0.01	1.34	0.23	0.10	0.02	0.01	0.27	100.87
16SD_015 ^B	1	50.10	13.50	14.00	8.93	6.60	3.35	0.53	0.01	1.11	0.28	0.10	0.03	0.03	1.32	99.89
16SD_020A ^B	1	51.60	14.35	14.30	9.21	5.37	2.34	0.79	0.03	1.43	0.23	0.17	0.02	0.03	0.09	99.96
10-LT-05 Standard – Batch A		48.70	17.9	9.69	9.61	6.5	3.52	0.99	0.03	1.4	0.14	0.37	0.11	0.06	0.59	99.63
10-LT-05 Standard – Batch B		49.60	18.8	9.35	9.47	6.52	3.63	1.01	0.03	1.4	0.14	0.38	0.11	0.06	0.82	101.3
10-LT-05 Standard - Average		48.88	18.18	9.54	9.52	6.61	3.60	1.02	0.03	1.44	0.14	0.38	0.11	0.06	0.75	100.27
2σ		1.16	0.50	0.29	0.32	0.21	0.20	0.03	0.01	0.06	0.00	0.02	0.00	0.00	0.34	1.69

Table 2-3 continues on the next page

Notes: Major elements in wt. % and trace elements in ppm. *Duplicate analysis. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B are provided in addition to the average composition and 2σ values of 25 analyses spanning 6 years. Superscript letters A or B following each sample number identify the corresponding batch.

Table 2-3: Continued

Swarm, locality, sample ID	Ba	Ce	Cr	Cs	Dy	Er	Eu	Ga	Gd	Hf	Ho	La	Lu	Nb	Nd	Pr	Rb	Sm	Sn	Sr	Ta
<i>Detection limit</i>	<i>0.5</i>	<i>0.5</i>	<i>10</i>	<i>0.01</i>	<i>0.05</i>	<i>0.03</i>	<i>0.03</i>	<i>0.1</i>	<i>0.05</i>	<i>0.2</i>	<i>0.01</i>	<i>0.5</i>	<i>0.01</i>	<i>0.2</i>	<i>0.1</i>	<i>0.03</i>	<i>0.2</i>	<i>0.03</i>	<i>1</i>	<i>0.1</i>	<i>0.1</i>
<i>Pääjärvi swarm, ca. 2450 Ma</i>																					
Loljunmaa, 15SD_070 ^A	178.0	22.9	1230	0.34	2.64	1.63	0.84	14.3	3.15	2.1	0.61	11.7	0.25	3.1	12.3	2.98	17.9	2.77	1	234	0.2
16SD_014 ^B	228.0	21.3	1750	0.64	2.10	1.23	0.78	12.0	2.56	2.7	0.48	8.8	0.17	3.4	11.1	2.49	17.8	2.87	1	159	0.3
<i>Uolevinlehto swarm, ca. 2400 Ma</i>																					
Uolevinlehto, 15SD_071 ^A	199.5	24.5	120	0.22	4.13	2.53	1.19	19.3	3.98	2.4	0.93	11.9	0.38	4.3	15	3.25	15.7	3.5	1	177	0.3
WD dyke (15SD_073) ^A	112.5	39.1	240	0.79	7.70	4.77	1.90	22.0	7.82	5.4	1.71	17.5	0.68	12.9	25.6	5.65	23.2	6.68	2	131	0.8
Viiianki dyke (15SD_084A) ^A	227.0	20.4	1330	0.57	2.26	1.28	0.70	13.5	2.47	1.9	0.52	9.6	0.21	2.6	10.9	2.62	18.3	2.44	<1	178	0.2
Viiianki dyke (15SD_084B) ^A	169.5	15.3	1580	0.53	1.80	1.15	0.55	12.6	1.96	1.3	0.41	8.0	0.15	2	8.1	1.98	14.8	1.73	1	170	0.1
<i>Tohmajärvi swarm, ca. 2135-2100 Ma</i>																					
15SD_090 ^A	365.0	46.5	230	0.64	6.78	3.94	1.72	20.0	6.43	5.1	1.39	22.4	0.54	16.2	24.9	6.15	47.4	6.15	2	189	1
Sipojuntti, 15SD_092 ^A	185.5	28.7	600	0.25	4.04	2.17	1.50	22.0	4.94	3.6	0.77	12.5	0.28	10.4	18.5	4.15	24.3	4.57	1	283	0.7
Myllykangas, 16SD_019A ^B	54.3	8.5	210	0.09	5.13	3.42	0.93	17.3	3.89	2.0	1.16	3.1	0.51	2.9	7.6	1.31	5.1	2.77	1	95.1	0.3
Myllykangas, 16SD_019A* ^B	51.3	8.3	190	0.08	5.08	3.37	0.88	16.7	3.95	1.9	1.16	3.2	0.55	2.8	6.6	1.33	5.4	2.63	1	92.4	0.2
Myllykangas, 16SD_019B ^B	34.9	7.6	10	0.06	2.96	1.99	0.74	15.4	2.28	1.4	0.65	3.3	0.34	2.5	4.8	1.08	3.9	1.75	1	132	0.1
<i>East-West swarm</i>																					
15SD_091 ^A	271.0	28.0	340	0.37	2.80	1.58	0.77	17.6	2.55	2.3	0.58	14.0	0.21	3.9	13.6	3.26	25.7	3.12	1	199	0.3
15SD_101B ^A	118.0	16.7	100	0.4	4.68	2.76	1.20	19.3	4.10	2.6	0.92	7.3	0.41	4.6	11.5	2.50	13.6	3.36	1	145	0.3
16SD_015 ^B	237.0	12.9	100	0.86	3.67	2.41	0.90	17.2	3.36	2.2	0.87	5.6	0.38	4.4	8	1.74	26.6	2.57	1	250	0.3
16SD_020A ^B	285.0	36.1	200	0.39	5.06	3.14	1.42	20.0	4.98	3.6	1.12	16.1	0.41	8.3	20.1	4.63	21.6	4.58	1	203	0.6
10-LT-05 Standard – Batch A	539.0	45.8	180	0.21	3.70	2.03	1.54	19.5	4.50	3.5	0.67	22.0	0.27	10.8	24.7	6.08	12.0	5.20	1	904	0.6
10-LT-05 Standard – Batch B	556.0	47.5	190	0.21	3.77	2.11	1.66	20.1	4.75	3.5	0.72	21.9	0.23	11.5	26.2	6.01	12.2	5.63	1	960	0.7
10-LT-05 Standard - Average	538	48.00	184	0.23	3.63	1.97	1.59	19.93	4.45	3.58	0.73	22.4	0.25	10.9	25.55	6.21	12.5	5.15	1.2	938	0.6
2σ	36.6	3.90	23	0.06	0.29	0.15	0.16	1.37	0.44	0.28	0.07	2.0	0.02	0.93	1.97	0.66	0.82	0.51	0.8	38.5	0.1

Table 2-3 continues on the next page

Notes: Major elements in wt. % and trace elements in ppm. *Duplicate analysis. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B are provided in addition to the average composition and 2σ values of 25 analyses spanning 6 years. Superscript letters A or B following each sample number identify the corresponding batch.

Table 2-3. Continued

Swarm, locality, sample ID	Tb	Th	Tm	U	V	W	Y	Yb	Zr	Ag	As	Cd	Co	Cu	Li	Mo	Ni	Pb	Sc	Tl	Zn
<i>Detection limit</i>	<i>0.01</i>	<i>0.05</i>	<i>0.01</i>	<i>0.05</i>	<i>5</i>	<i>1</i>	<i>0.5</i>	<i>0.03</i>	<i>2</i>	<i>0.5</i>	<i>5</i>	<i>0.5</i>	<i>1</i>	<i>1</i>	<i>10</i>	<i>1</i>	<i>1</i>	<i>2</i>	<i>1</i>	<i>10</i>	<i>2</i>
<i>Pääjärvi swarm, ca. 2450 Ma</i>																					
Loljunmaa, 15SD_070 ^A	0.45	2.19	0.26	0.39	199	<1	15.1	1.62	78	<0.5	<5	<0.5	58	90	10	<1	375	9	29	<10	82
16SD_014 ^B	0.40	3.45	0.19	0.64	171	1	12.5	1.16	87	<0.5	<5	<0.5	69	73	10	<1	751	<2	22	<10	83
<i>Uolevinlehto swarm, ca. 2400 Ma</i>																					
Uolevinlehto, 15SD_071 ^A	0.61	1.95	0.39	0.47	302	<1	22.7	2.60	91	<0.5	<5	<0.5	45	122	<10	<1	83	2	37	<10	106
WD dyke (15SD_073) ^A	1.24	1.78	0.73	0.42	303	2	43.1	4.65	223	<0.5	<5	0.5	50	428	10	<1	144	2	36	<10	134
Viiianki dyke (15SD_084A) ^A	0.36	1.81	0.18	0.38	197	<1	13.0	1.28	62	<0.5	<5	0.5	64	94	10	<1	411	3	30	<10	86
Viiianki dyke (15SD_084B) ^A	0.30	1.45	0.18	0.30	182	<1	10.1	1.07	48	<0.5	<5	<0.5	62	80	10	<1	440	<2	27	<10	74
<i>Tohmajärvi swarm, ca. 2135-2100 Ma</i>																					
15SD_090 ^A	1.10	4.77	0.60	1.13	389	<1	34.1	3.60	188	<0.5	<5	<0.5	42	51	20	<1	51	42	39	<10	157
Sipojuntti, 15SD_092 ^A	0.72	2.33	0.29	0.54	362	<1	19.5	1.85	117	<0.5	<5	<0.5	51	60	20	<1	276	2	26	<10	98
Myllykangas, 16SD_019A ^B	0.79	0.24	0.52	0.09	364	1	31.5	3.48	67	<0.5	<5	0.8	58	137	20	<1	85	5	52	<10	109
Myllykangas, 16SD_019A* ^B	0.70	0.30	0.51	0.09	335	<1	31.1	3.48	63	<0.5	<5	0.9	53	120	10	<1	82	<2	48	<10	96
Myllykangas, 16SD_019B ^B	0.42	0.81	0.34	0.24	508	<1	17.8	2.09	48	<0.5	<5	<0.5	38	160	10	<1	44	6	40	<10	75
<i>East-West swarm</i>																					
15SD_091 ^A	0.47	4.08	0.24	0.87	197	1	15.4	1.52	87	<0.5	<5	<0.5	47	110	20	<1	65	13	30	<10	82
15SD_101B ^A	0.71	1.03	0.40	0.21	377	1	24.0	2.73	85	<0.5	<5	<0.5	50	223	10	<1	67	<2	39	<10	98
16SD_015 ^B	0.63	0.91	0.38	0.21	337	<1	22.3	2.51	76	<0.5	<5	0.7	50	334	10	<1	80	5	47	<10	87
16SD_020A ^B	0.78	2.86	0.37	0.68	307	1	27.4	2.91	125	<0.5	<5	1	47	203	20	1	61	6	36	<10	133
10-LT-05 Standard – Batch A	0.66	2.01	0.27	0.56	154	1	17.4	1.64	137	<0.5	<5	<0.5	39	41	10	<1	116	<2	17	<10	91
10-LT-05 Standard – Batch B	0.67	1.88	0.25	0.59	165	1	18.9	1.72	152	<0.5	<5	<0.5	43	45	10	1	124	<2	18	<10	100
10-LT-05 Standard - Average	0.66	1.99	0.27	0.59	163.8	1	18.6	1.66	150	NA	7	1	39	42	10	1	115	4	17	10	92
2σ	0.06	0.22	0.04	0.10	21.7	0	1.5	0.21	14	NA	1	0	4	4	0	1	9	3	2	0	7

Notes: Major elements in wt. % and trace elements in ppm. *Duplicate analysis. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B are provided in addition to the average composition and 2σ values of 25 analyses spanning 6 years. Superscript letters A or B following each sample number identify the corresponding batch.

2.6 Discussion

2.6.1 Updated dyke swarm classification in the Pudasjärvi block and correlations to the Taivalkoski and Kuhmo blocks

Building on the overview by Vuollo and Huhma (2005), we present an updated classification scheme for dyke swarms in the Pudasjärvi block based on U-Pb geochronology and dyke trends. This classification involves six swarms but leaves hundreds of dykes still unclassified. The six dyke swarms are: *ca.* 2450 Ma NE-trending Pääjärvi, *ca.* 2400 Ma NW-trending Uolevinlehto, *ca.* 2135-2100 Ma NW-trending Tohmajärvi, *ca.* 2077 Ma WNW-trending Palomaa, *ca.* 1980 Ma NNW-trending Paukkajanvaara and undated “East-West” dykes (**Table 2-1; Figure 2-3**). The only swarm not modified by our reclassification is the *ca.* 1980 Ma Paukkajanvaara swarm.

Of the six dyke swarms in the Pudasjärvi block described above, the three dated in this study by U-Pb ID-TIMS, have age matches with dykes in the Kuhmo and Taivalkoski blocks (**Table 2-5; Figure 2-15a**; Pekkarinen and Lukkarinen, 1991; Vuollo *et al.*, 1992; Mertanen *et al.*, 1999; Vuollo and Huhma, 2005; Stepanova *et al.*, 2014, 2015; Huhma *et al.*, 2018). While no matches to the Palomaa or East-West swarms have been identified in the Kuhmo and Taivalkoski blocks, there are multiple sets of undated and unclassified dykes with appropriate trends after restoration that could be the continuation of these swarms.

Figure 2-15. Position of the Pudasjärvi block with respect to the Taivalkoski and Kuhmo blocks. a) Current position of the Pudasjärvi block and Paleoproterozoic mafic dykes in Pudasjärvi, Taivalkoski and Kuhmo. Localities B, D, G, and F from **Table 1**. Localities 1-11 from **Table 4**. b) Suggested 30° rotation counter-clockwise to restore the Pudasjärvi block to its likely original position relative to Taivalkoski and Kuhmo blocks and the main part of the Karelia craton. Star indicates rotation axis.

2.6.1.1 Pääjärvi swarm (NE-trending, Pudasjärvi block; NW regional trend, *ca.* 2450 Ma)

The *ca.* 2450 Ma Pääjärvi swarm is reclassified to include a branch of NE-trending dykes mapped by Vuollo and Huhma (2005) based on aeromagnetic data. These dykes are represented by the subparallel 16SD_014 dyke and 2444 ± 4 Ma Vengasvaara intrusion (**Figure 2-3**).

Dykes dated at *ca.* 2.45 Ga in the Taivalkoski and Kuhmo blocks include the Suoperä (BD) and Lupzenga (XD) dykes in the Taivalkoski block and the Lashkujärvi dyke in the Kuhmo block (**Table 2-5; Figure 2-15a**; Mertanen *et al.*, 1999; Stepanova *et al.*, 2015; Huhma *et al.*, 2018). Mertanen *et al.* (1999) also noted a thin (43-80 cm) 2.45 Ga NW-trending dyke at the Varisniemi (VN) locality, identified by its similar demagnetization behaviour compared to the Suoperä (BD) dyke. Additionally, a number of layered mafic intrusions and sills dated at 2.45 Ga have been recognized across Karelia-Kola (e.g., Alapieti *et al.*, 1990; Lobach-Zhuchenko *et al.*, 1998; Hanski *et al.*, 2001). Some of these mafic intrusions belong to the 2450 Ma Pääjärvi event but because they do not contribute the same geometric information as dykes, they will not be discussed further.

Suoperä (2447 ± 10 Ma) and Lashkujärvi (2450 ± 12 Ma) are the only dykes with U-Pb TIMS ages. Despite their nearly identical ages, the NW orientations of the subparallel Suoperä (BD), Lupzenga (XD) and Varisniemi (VN) localities are nearly perpendicular to the NE-trending Lashkujärvi dyke of Stepanova *et al.* (2015). Regional structural analyses of the Karelia-Kola craton have linked NW-trending 2.45 Ga dykes with regionally extensive parallel structures to northeast-southwest extension (Skyttä *et al.*, 2019). Local variations in swarm trends are not an uncommon occurrence for regional swarms. The *ca.*

2.22 Ga Ungava swarm in the Superior craton, for example, involves a regional radiating swarm pattern defined by the Maguire, Senneterre and Klotz dykes. The 2.22 Ga Ungava event, however, also includes the volumetrically minor Anuc, Kogaluk Bay and Couture dykes with distinct orientations (Maurice *et al.*, 2009; Davey *et al.*, 2019).

2.6.1.2 Uolevinlehto swarm (NW-trending, *ca.* 2400 Ma)

The NW-trending Uolevinlehto dyke, formerly assigned to the *ca.* 2450 Ma Pääjärvi swarm of Vuollo and Huhma (2005), is reclassified as a distinct swarm at *ca.* 2400 Ma based on dating herein (section **2.5.3.1**). Dykes with subparallel trends in the vicinity of Uolevinlehto are also assigned to the Uolevinlehto swarm. A separate swarm distinct from the Pääjärvi swarm is proposed because of the conspicuously younger U-Pb age of 2400 ± 12 Ma compared with the ages of the *ca.* 2450 Ma Pääjärvi dykes and layered mafic intrusions across Karelia. Additionally, the *ca.* 2.40 Ga Uolevinlehto dyke has a distinct trend from the Vengasvaara intrusion and the 16SD_014 dyke.

Age matches to the *ca.* 2400 Ma Uolevinlehto dyke include the E-W-trending Viianki and Taivalkoski (WD) dykes in the Kuhmo and Taivalkoski blocks, respectively (**Table 2-5; Figure 2-15a**; Vuollo and Huhma, 2005; Huhma *et al.*, 2018). Preliminary U-Pb analyses on multigrain baddeleyite fractions provide an imprecise upper intercept age at 2404 ± 29 Ma for the Viianki dyke (**Chapter 3**). Huhma *et al.* (2018) report an age for the Taivalkoski (WD) dyke at 2407 ± 35 Ma based on a Sm-Nd whole-rock, plagioclase and pyroxene mineral/internal isochron. Additional ages of dykes *ca.* 2.40 Ga have been identified in Russian Karelia by Stepanova *et al.* (2017) and further support the proposed existence of a wide-spread dyke swarm at this time.

2.6.1.3 Tohmajärvi swarm (NW-trending, *ca.* 2135-2100 Ma), and Palomaa swarm (WNW-trending *ca.* 2077 Ma)

The Tohmajärvi swarm described by Vuollo and Huhma (2005) is divided into two separate swarms, Tohmajärvi and Palomaa, on account of their differing trends. Tohmajärvi dykes are constrained based on the ages and NW trends of the 2128.9 ± 1.2 Ma Sipojuntti (15SD_092) and $2135.2 \pm 3.6/3.7$ Ma Myllykangas (16SD_019) dykes. The Palomaa dyke swarm is distinguished by a WNW trend offset $\sim 30^\circ$ from the Tohmajärvi swarm (locality H in **Table 2-1** and **Figure 2-3**). While tentative, the Palomaa swarm is dated by a Sm-Nd whole-rock, plagioclase and pyroxene mineral/internal isochron at 2077 ± 35 Ma (Huhma *et al.*, 2018). Aeromagnetic data highlight a number of subparallel dykes in the vicinity of the Palomaa locality and these are included in the Palomaa swarm.

The Sipojuntti and Myllykangas dykes of the Tohmajärvi swarm in the Pudasjärvi block have age matches in the Kuhmo block, including the NW-trending Pirtguba, Piippolanmäki and Kutsu dykes (**Table 2-5; Figure 2-15a**; Pekkarinen and Lukkarinen, 1991; Stepanova *et al.*, 2014). Located in Russian Karelia, the Pirtguba dyke is the eastern-most dyke thus far identified as part of the *ca.* 2.11 Ga age group by a U-Pb age at 2126 ± 5 Ma (locality 7 in **Table 2-5; Figure 2-15**; Stepanova *et al.*, 2014). The Kutsu and Piippolanmäki dykes are also grouped into the *ca.* 2.11 Ga dyke swarm by U-Pb geochronology with an age of 2113 ± 4 Ma (**Table 2-5**; Pekkarinen and Lukkarinen, 1991).

2.6.1.4 Paukkajanvaara swarm (NNW-trending, *ca.* 1980 Ma)

Although we were unsuccessful in sampling *ca.* 1.98 Ga Paukkajanvaara dykes in the Pudasjärvi block, a dense swarm of *ca.* 1980 Ma dykes with comparable aeromagnetic

signatures cuts across the Kuhmo block (**Table 2-5; Figure 2-15a**). This dense swarm of Paukkajanvaara dykes in the Kuhmo block has been previously linked to lineaments with high aeromagnetic signatures in the Pudasjärvi block (Vuollo *et al.*, 2000; Vuollo and Huhma, 2005). Paukkajanvaara swarm ages in the Kuhmo block include the Kivikevätti (KD) dyke at 1980 ± 4 Ma based on analyses of four multigrain baddeleyite fractions (Huhma *et al.*, 2018) and the Paukkajanvaara dyke at 1963 ± 8 Ma from TIMS analyses on zircon (Vuollo *et al.*, 1992; Huhma *et al.*, 2018). This swarm is broadly linked to units (mafic volcanic rocks, dykes and sills) through Karelia-Kola, defining the Pechenga-Onega LIP (Lubnina *et al.*, 2016a, 2017).

2.6.1.5 “East-West” dykes in the Pudasjärvi block

We introduce the undated “East-West” trending dykes as a new swarm in the Pudasjärvi block (**Table 2-1; Figure 2-3**). The cross-cutting relationship where the dykes 16SD_014 (Pääjärvi swarm) and 16SD_015 intersect indicate that the East-West dykes are younger than the *ca.* 2450 Ma Pääjärvi dykes (**Figure 2-4b**). An EW-trending dyke (16SD_020) was also found at the Myllykangas (NW-trending dyke site 16SD_019 at $2135.2 \pm 3.6/3.7$ Ma) locality but no contacts were located. Dyke site 16SD_020 has high magnetic susceptibility (up to $22 \cdot 10^{-3}$ SI units compared to $<2 \cdot 10^{-3}$ SI units in Myllykangas) and has better preserved primary magmatic minerals compared to Myllykangas. These features suggest that the East-West dykes are also younger than the *ca.* 2135-2100 Ma Tohmajärvi swarm (see **Figure 2-6 to Figure 2-9** to compare petrography of the Tohmajärvi and East-West dykes).

Table 2-5. Locations and characteristics of mafic dykes in the Taivalkoski and Kuhmo blocks.

Map ID	Dyke swarm Locality, sample	Block	Latitude (°N)	Longitude (°E)	Trend (°)	Dip (°)	Width (m)	Age (Ma)	Dating Method	εNd	References
Pääjärvi swarm, ca. 2450 Ma											
1	Suoperä, BD/A1415	Taivalkoski	65.8031	30.3276	NNW	90 [†]	50-60	2447±10	U-Pb TIMS, Bdy	-2.4	Mertanen <i>et al.</i> (1999); Huhma <i>et al.</i> (2018)
2	Lupzenga, XD1	Taivalkoski	66.1447	30.9893	NNW	90 [†]	30	2421±32	Sm-Nd*	-1.4	Mertanen <i>et al.</i> (1999); Huhma <i>et al.</i> (2018)
3	Lupzenga, XD3	Taivalkoski	66.1412	30.9831	NW	90 [†]	30	2476±35	Sm-Nd*	+1.7	Mertanen <i>et al.</i> (1999); Huhma <i>et al.</i> (2018)
4	Lashkujärvi/Ca-230	Kuhmo	65.0931	30.5435	NE	-	-	2450±12	U-Pb TIMS, Bdy	-1.7	Stepanova <i>et al.</i> (2015)
Uolevinlehto swarm, ca. 2400 Ma											
5	Taivalkoski, WD	Taivalkoski	65.5784	28.2640	E (090)	74S	10	2407±35	Sm-Nd*	+1.6	Huhma <i>et al.</i> (2018)
6	Viianki, VD	Kuhmo	64.7727	30.0731	E (090)	~80N	25	min. 2395	U-Pb, Bdy	-1.2	Vuollo and Huhma (2005); Huhma <i>et al.</i> (2018)
Tohmajärvi swarm, ca. 2135-2100 Ma											
7	Pirtguba/Ca-255	Kuhmo	64.9786	30.7358	NNW (325)	-	70	2126±5	U-Pb SIMS, Zrc	+2.8; +1.4	Stepanova <i>et al.</i> (2014)
8	Piippolanmäki/465	Kuhmo	62.4440	30.2870	NW	90 [†]	-	2113±4	U-Pb TIMS, Zrc		Pekkarinen and Lukkarinen (1991)
9	Kutsu/A427		62.2900	30.5750		90 [†]					
Paukkajanvaara swarm, ca. 1980 Ma											
10	Kivikevätti, KD/A1409	Kuhmo	64.7548	29.7921	NW (330)	84NE	60	1980±4	U-Pb TIMS, Bdy	+0.3	Huhma <i>et al.</i> (2018)
11	Paukkajanvaara/All44A	Kuhmo	62.9310	30.0580	NW	-	-	1963±8	U-Pb TIMS, Zrc	+1.4	Vuollo <i>et al.</i> (1992); Huhma <i>et al.</i> (2018)
-	A1175	Kuhmo	-	-	NW	-	-	1972±5	U-Pb TIMS, Zrc	+0.9	Vuollo <i>et al.</i> (1992); Vuollo and Huhma (2005)

Notes: Localities 1-11 shown in **Figure 2-15**. Dips reported with a "~" indicate those interpreted from contact-parallel parting planes and/or columnar jointing. Abbreviations: Bdy - baddeleyite; Zrc - chemically abraded zircon. † Dip is reported as subvertical. *Sm-Nd age from whole-rock, plagioclase, and pyroxene mineral/internal isochron.

2.6.2 Geochemical characteristics of the dyke swarms

Three geochemical groups are recognized within the subalkaline tholeiitic basalt to basaltic andesite dykes and related intrusions of the *ca.* 2450 Ma Pääjärvi, *ca.* 2400 Ma Uolevinlehto, *ca.* 2135-2100 Ma Tohmajärvi and undated East-West dyke swarms in the Pudasjärvi block and contemporaneous dykes from the Kuhmo and Taivalkoski blocks (**Figure 2-14** and **Figure 2-15**). The Fe-tholeiitic basalts of Group 1 include the largest population of dykes and layered mafic intrusions (57% of the sample set) with representatives identified in the Uolevinlehto, Tohmajärvi and East-West swarms. Group 2 includes dykes and intrusions with “boninitic” characteristics, recognized by their elevated Mg numbers with respect to SiO₂ wt. %. Lastly, Group 3 includes dykes with MORB-type characteristics and, while apparently rare in the Pudasjärvi block, Stepanova *et al.* (2014) have reported dykes in the Russian part of the Kuhmo block and Central Karelia with identical geochemistry.

2.6.2.1 Group 1: Fe-tholeiitic basalts

Fe-tholeiitic basalts have been described in many Karelian dykes (Vuollo *et al.*, 1992, 1995; Nykänen *et al.*, 1994; Vuollo and Huhma, 2005). Dykes with Fe-tholeiitic basalt chemistries have FeO/MgO ratios greater than 1.4 (**Figure 2-16a**) and plot within the continental flood basalt field of the V versus Ti diagram (**Figure 2-16b**; Shervais, 1982). In a Nb/Y versus Zr/Y diagram, dykes and layered mafic intrusions belonging to Group 1 plot below the Iceland array (**Figure 2-16c**; Fitton *et al.*, 1997). These samples indicate a negative correlation between ΔNb ($\Delta\text{Nb} = 1.74 + \log[\text{Nb}/\text{Y}] - 1.92\log[\text{Zr}/\text{Y}]$), as calculated in Fitton *et al.*, 1997) and Th/Nb (**Figure 2-16**) in which lower ΔNb values

correlate to increased distance from the Iceland array in **Figure 2-16c**. Their multi-element patterns include flat to shallow HREE slopes, shallow negative LREE slopes, a negative Nb anomaly and minor enrichment in Th with respect to La (**Figure 2-16e**). Flat HREE slopes are characteristic of shallow upper mantle melting above the garnet stability field, or where no garnet is present in the residue due to high percentage melting (Pearce, 2008). The overall abundance of Nb is dependent on the composition of the source and degree of partial melting in which lower degrees of partial melting yield higher concentrations of Nb relative to higher degrees of partial melting of the same source (Pearce, 2008). According to Vuollo et al. (1992) and Nykänen et al. (1994) these magmas were produced by 10-20% partial melting, which is in agreement with the higher Nb/Y and Zr/Y ratios shown in **Figure 2-16c**. The negative Nb anomalies on the multi-element diagram correspond to ΔNb values that plot below the Iceland array on the Nb/Y vs. Zr/Y diagram and correlation between ΔNb and Th/Nb indicate that the magmas incorporated a crustal component (**Figure 2-9c and d**; Fitton *et al.*, 1997; Pearce, 2008). Crustal assimilation is supported by previous modelling of assimilation and fractional crystallization (Nykänen *et al.*, 1994). The limited range of existing ϵNd values from +0.3 to +1.7 (**Table 2-1** and **Table 2-5**) supports a common source and similar lithospheric contamination for these dykes, despite their occurrence in many dyke swarms.

2.6.2.2 Group 2: High Mg-tholeiitic basalts

Elevated SiO_2 wt. % and Mg numbers are recognized within a subset of dykes and layered mafic intrusions older than 2400 Ma, corresponding to the *ca.* 2450 Ma Pääjärvi and *ca.* 2400 Ma Uolevinlehto swarms within the Pudasjärvi block (**Figure 2-11**, **Figure 2-13**,

Figure 2-14, and **Figure 2-16**; Alapieti *et al.*, 1990; Vuollo *et al.*, 1995). Samples part of this group plot in the Komatiite field on a Jensen classification diagram (**Figure 2-13**). Their siliceous-high-Mg composition is superficially comparable to basalts and andesites in the Izu-Bonin arc, Japan. For this reason, the term “boninitic” has often been used in reference to their geochemistry without invoking a similar genetic model (Alapieti *et al.*, 1990; Vogel *et al.*, 1998; Srivastava, 2008; Pearce and Reagan, 2019). The dykes and layered mafic intrusions belonging to the high Mg-tholeiitic basalt Group 2 have lower FeO/MgO ratios (**Figure 2-16a**) compared to those of Group 1 but overlap the arc and continental flood basalt fields on a V versus Ti diagram (**Figure 2-16b**; Shervais, 1982). Dykes belonging to Group 2 are similar to those from Group 1 on a Nb/Y versus Zr/Y diagram (**Figure 2-16b**) but the associated layered mafic intrusions have overall lower Nb/Y and Zr/Y values and plot separately within the depleted MORB and Iceland array fields (**Figure 2-16c**; Luttinen, 2018). The parallel multi-element patterns of the dykes and layered mafic intrusions (**Figure 2-16f**) include flat HREE slopes, steep negative LREE slopes, a pronounced negative Nb anomaly and enrichment in Th with respect to La (with Th/Nb_{PM} values greater than 4). The overall abundances of trace elements vary by an order of magnitude, especially in the sills (**Figure 2-16f**). However, parallel multi-element patterns with varying trace element abundances can be related by cumulates and liquids from the same batch of magma or by fractional crystallization (Rollinson, 1993). The proportionally lower Nb/Y and Zr/Y ratios, that separate many of the layered mafic intrusions from the dykes, can be explained by varying degrees of partial melting (**Figure 2-16c and f**). In all dykes and layered mafic intrusions, the presence of flat HREE slopes and V/Ti ratios around 20 indicate that they were derived from shallow upper mantle

similar to Group 1 in either a continental or arc setting. The pronounced negative Nb anomalies and Th enrichment on the multi-element diagram, and negative correlation between ΔNb and Th/Nb (**Figure 2-16d**), indicate that these magmas either assimilated a lithospheric component (crustal or metasomatized mantle) or are derived from an arc setting. The geochemical signature of Group 2 has been recognized in younger 1.98 Ga dykes from the Kuhmo block by Vuollo *et al.* (1992); the T2 Group. The petrogenetic study by Vuollo *et al.* (1992) of these dykes concluded that their magmas were generated in an arc setting by 30% partial melting of the mantle. However, the same geochemical features reported by Vuollo *et al.* (1992), such as V/Ti ratios ~ 20 , LREE enrichment and High Field Strength Element (HFSE) depletion may also be explained by crustal assimilation or having been derived from metasomatized mantle. The uniform and uniquely negative ϵNd values between -1.2 and -5.2 are distinct from Groups 1 and 3 and are also suggestive of having incorporated a lithospheric component but does not provide conclusive evidence for either crustal assimilation, sourcing from metasomatized mantle or generation within an arc setting (**Table 2-1** and **Table 2-5**).

Figure 2-16. *On the next page.* Dykes and selected intrusions classified by their geochemical characteristics into Group 1 (Fe-tholeiitic basalt), 2 (Mg-tholeiitic basalt) and 3 (MORB-type basalt). Colours correspond to geochemical group and shapes correspond to dyke swarm assignment based on age and trend. a) $\text{FeO}_{\text{Total}}$ versus MgO with Rayleigh crystal fractionation curve as in Vuollo *et al.* (1992). b) V versus Ti/1000 (Shervais, 1982). c) Nb/Y versus Zr/Y diagram after Fitton *et al.* (1997) and depleted MORB field and labels from Luttinen (2018). d) ΔNb versus Th/Nb (Fitton *et al.*, 1997; Luttinen, 2018). e-g) Multi-element plots of Groups 1, 2 and 3 normalized to primitive mantle (Sun and McDonough, 1989). Data for the Pirtguba, Lashkujärvi and layered mafic intrusions from Stepanova *et al.* (2014; 2015) and GTK (2018).

e. Group 1: Fe-tholeiitic basalt

Includes representatives from the Pääjärvi, Uolevinlehto, Tohmajärvi and East-West swarms.

f. Group 2: Mg-tholeiitic basalt

Includes representatives from the Pääjärvi and Uolevinlehto swarms.

g. Group 3: MORB-type tholeiitic basalt

Includes dykes from the Tohmajärvi swarm.

Geochemical groups

Group 1: Fe-tholeiitic basalt

○ ■ ◆ Dykes

▲ Layered mafic intrusions

Group 2: Mg-tholeiitic basalt

◆ ▲ Dykes

△ Layered mafic intrusions

Group 3: MORB-type basalt

● Dykes

Dyke swarm, age (Ma)

□ East-West

○ Tohmajärvi, ca. 2130-2100

◇ Uolevinlehto, ca. 2400

△ Pääjärvi, ca. 2450

2.6.2.3 Group 3: MORB-type basalts

Group 3 MORB-types basalts are distinguished by $+\Delta\text{Nb}$ values and flat multi-element patterns (**Figure 2-16 c, d, and g**) but have Mg numbers (**Figure 2-14b**), FeO/MgO and Ti/V (**Figure 2-16a and b**) ratios similar to those of Group 1. In the Pudasjärvi block, only the 2135 Ma Myllykangas dyke has this geochemistry but identical geochemistry is reported by Nykänen *et al.* (1994) and Stepanova *et al.* (2014) in the contemporaneous *ca.* 2.11 Ga dykes, as well as other dykes ranging in age between 2.14 and 2.06 Ga in Russian Karelia. With flat HREEs and only minor evidence for a lithospheric contribution, these authors attribute the generation of MORB-type dykes to 10-15% partial melting of shallow upper mantle under thin lithosphere within an extensional tectonic environment. The ϵNd values of +1.4 and +2.8 of the Pirtguba dyke presented by Stepanova *et al.* (2014) distinguish Group 3 isotopically from Group 2 but coincide with ϵNd values of Group 1, suggesting that Groups 1 and 3 may have been derived from a similar mantle source. 2σ

2.6.3 A unique geochemical fingerprint?

Attempts thus far to identify a unique “geochemical fingerprint” for each dyke swarm in the Pudasjärvi have been unsuccessful. This is not unexpected for two reasons: (1) more than one geochemical pattern is often identified within a single dyke swarm in which their swarm assignments and age consistency are confirmed by U-Pb geochronology and (2) nearly identical geochemical patterns are found in various swarms in which their distinct ages are confirmed by U-Pb geochronology (Ernst, 2014). Both circumstances are reported in LIPs from other cratons. For example, the in-situ and undeformed *ca.* 66 Ma Deccan continental flood basalt lavas with the duration of eruption constrained to less than a

million years, are an example of lavas with heterogeneous chemostratigraphy (Lightfoot *et al.*, 1990; Schoene *et al.*, 2015). Geochemical heterogeneity is also identified within the dyke swarm record. For example, the *ca.* 2.22 Ga Maguire and Klotz dykes in the northern Superior craton are geochemically distinct but both belong to the Ungava LIP (Maurice *et al.*, 2009; Davey *et al.*, 2019). Maurice *et al.* (2009) identify geochemical matches to the 2.22 Ga Klotz dykes in the younger 2.0 Ga Povungnituk basalts part of the Minto-Povunituk LIP, and these too are geochemically heterogeneous. Kastek *et al.* (2018) report geochemical variation in the 2.0 Ga Ma Minto-Povunituk LIP that can be divided geographically in which the northern domain hosts MORB and OIB-like suites, and the southern region, represented by Minto dykes and Eskimo volcanics, is dominated by an E-MORB-like signature. Maurice *et al.* (2009) and Kastek *et al.* (2018), propose that magmas were generated from heterogeneous mantle and plume sources that variably incorporated enriched lithospheric components. Identical geochemical signatures can be produced when the same mantle is repeatedly tapped and when their magmas are incorporating similar lithospheric components (Maurice *et al.*, 2009; Kastek *et al.*, 2018). These examples and their subsequent interpretation illustrate the occurrence of multiple geochemical signatures within a single LIP and the reoccurrence of identical geochemical signatures in different LIPs through time. Geochemical heterogeneity in flood basalt and dyke swarm components throughout the LIP record therefore emphasizes the necessity of high-precision U-Pb geochronology for accurate swarm assignment.

2.6.4 Rotation of the Pudasjärvi block

Rotation of the Pudasjärvi block was first discussed by Vuollo *et al.* (2000). Their

preliminary analysis of dyke swarm orientations in the Pudasjärvi block relative to those in the Kuhmo and Taivalkoski blocks required a counter-clockwise vertical axis rotation of 50° for the *ca.* 2450 Ma Pääjärvi dykes, 40° for the 2135-2100 Ma Tohmajärvi dykes and 30° for the 1980 Ma Paukkajanvaara dykes to restore the Pudasjärvi block to its apparent initial position. We apply a similar approach, but base the restoration on the *ca.* 2135-2100 Ma Tohmajärvi swarm because it includes consistent NW trends reported by multiple authors from across Western Karelia (Vuollo *et al.*, 1992; Nykänen *et al.*, 1994; Stepanova *et al.*, 2014). We conclude that dextral transpression during the *ca.* 1.92-1.79 Ga Svecofennian orogeny caused the Pudasjärvi block to rotate $\sim 30^\circ$ clockwise (on a vertical axis; **Figure 2-15b**). Rotation along horizontal axes (tilting), either locally or regionally, can be discounted because all dykes, with their various trends, still maintain (sub)vertical dips in the Pudasjärvi block (**Table 2-1**). Clockwise rotation is consistent with dextral transpressional deformation previously recognized and described along the Archean-Proterozoic collisional boundary (Gaál, 1990; Sorjonen-Ward, 2006; Hölttä *et al.*, 2012; Lahtinen *et al.*, 2015) as well as along the N-S Hirvaskoski Shear Zone, which divides the Pudasjärvi block from the Kuhmo and Taivalkoski blocks (Kärki *et al.*, 1993). This “best-fit” restoration of the Pudasjärvi block (**Figure 2-15b**) aligns dyke 16SD_014 to the NW-trending Lupzenga (XD) and Suoperä (BD) dykes and not with the NE-trending Lashkujärvi dykes (**Figure 2-15b**). The regional trend of the *ca.* 2450 Ga Pääjärvi dykes is therefore interpreted to be represented by the NNW-trending dykes and these are related to regionally extensive parallel structures associated with northeast-southwest extension (Skyttä *et al.*, 2019). The structural restoration also roughly aligns dykes with *ca.* 2400 Ma ages since the Uolevinlehto dyke becomes approximately parallel with the Taivalkoski

(WD) and Viianki dykes. Similar to the conclusions reached by Vuollo *et al.* (2000), the undated Paukkajanvaara dykes in the Pudasjärvi block would align in parallel with the dated Paukkajanvaara dykes in the Kuhmo and Taivalkoski blocks.

Rotation of the Pudasjärvi block has wider implications for the greater Karelia-Kola metacraton because a block cannot undergo deformation-induced rotation without influencing adjacent crust. Whether vertical axis rotation of the Pudasjärvi block has significantly affected the positioning of the Iisalmi block and/or Lapland to as far as northwest as the “Norbotten craton” requires additional study. An additional translational component (related to orogenic deformation and crustal shortening) has also not been quantified by our proposed structural restoration. While the swarms clearly indicate that Pudasjärvi was a local rifted block or fragment of Western Karelia prior to the Svecofennian orogeny, its initial distance from the margin of the Kuhmo and Taivalkoski blocks remains poorly constrained. With additional age matches, it may be possible to identify more precise piercing points to constrain the translational components of Pudasjärvi’s reworking.

2.6.5 Superia reconstruction

Our proposed rotation of the Pudasjärvi block has constrained the regional trends of the *ca.* 2450 Ma Pääjärvi, *ca.* 2400 Ma Uolevinlehto and *ca.* 2135-2100 Ma Tohmajärvi swarms in the greater Karelia-Kola metacraton, and refined the position of Karelia-Kola, with respect to the Superior and Wyoming cratons in the context of supercraton Superia (**Figure 2-17**). In order for both the Pääjärvi and Tohmajärvi swarms to fit approximately within the context of their counterparts in the Superior craton (Matachewan and Marathon

swarms, respectively), Karelia-Kola is positioned such that is south of the East half of the Superior craton and southeast of the Wyoming craton (**Figure 2-17**). The Pääjärvi dykes have two dominant trends including a local subset with a NE trend (Lashkujärvi) and a more regional subset with a NW trend, which are also tied to regional structures related to northeast extension (Skyttä *et al.*, 2019). This *ca.* 2450 Ma subset of Pääjärvi dykes with a NW trend points towards a magmatic centre located at the southeastern margin of the Superior craton but their direct trajectory skims slightly past the magmatic centre of their Superior counterpart of the Matachewan magmatic centre. While approximately consistent with regional structures in Karelia-Kola as well as with magmatism located at the southeast margin of the Superior craton, the fact that the Pääjärvi dykes have more than one orientation means that the few dated dykes that are part of the swarm may not necessarily represent the full regional dyke swarm pattern (e.g., subtle curvature of the dykes; **Figure 2-17**). In the configuration presented in **Figure 2-17**, the *ca.* 2400 Ma Uolevinlehto dykes and coeval du Chef dykes (Krogh, 1994) converge towards two plausible magmatic centres, located either at the southern margin of the Superior craton near the Matachewan magmatic centre, or further south along the adjoining margins between Karelia-Kola and Wyoming. Additionally, the imprecisely dated *ca.* 2077 $\pm$ 35 Ma Palomaa dykes trend towards the 2075-2076 Ma Fort Frances magmatic centre. The $\sim 30^\circ$ difference in trend used to differentiate between the Tohmajärvi and Palomaa swarms is also found in their counterparts in Superior—the Marathon and Fort Frances dyke swarms, respectively.

Another match that falls out of this configuration is between the nearly parallel *ca.* 1980 Ma Mine Centre dyke (Ernst *et al.*, 2006; M. Hamilton, pers. comm., 2019) of the Western Superior craton and the Paukkajanvaara dykes of Western Karelia. Furthermore,

with Karelia-Kola next to the Wyoming craton, the *ca.* 3.5 Ga Siurua Gneisses, Europe's oldest gneiss complex, is placed adjacent to some of the oldest rocks in the Wyoming craton at 3.2-3.8 Ga (e.g., *ca.* 3.3 Ga tonalitic gneisses in Tobacco Root Mountains; Mueller *et al.*, 2004; Krogh *et al.*, 2011).

Figure 2-17. *On the next page.* Latest Neoarchean- to early Paleoproterozoic Superia reconstruction joining Superior, Wyoming and Karelia-Kola cratons based on LIPs and their radiating dyke swarms, modified after Bleeker and Ernst (2006); see also Kilian *et al.* (2016a). Western Superior rotated 14° clockwise relative to eastern Superior (Evans and Halls, 2010). The 2000 km-diameter circle around the Matachewan magmatic centre indicates a typical size of an impinging plume head. Dashed lines from the Matachewan plume centre towards the Pääjärvi dykes illustrate the subtle curvature in trends. Extrapolated dyke trends are also drawn for the Tohmajärvi dykes towards the Marathon magmatic centre and for the Uolevinlehto and du Chef dykes towards their tentative magmatic centre. Present day north arrows indicate relative orientation of each craton. Minimum size estimates of Archean fragments represented by pink and purple shapes outlined by black dashed lines. Superia extent is light grey, approximate size and shape undefined. Abbreviations: BDM - Blue Draw Metagabbro, *ca.* 2450 Ma; MC - Mine Centre (M. Hamilton, pers. comm., 2019); Nb - Norbotten craton; Ko - Koitelainen. Superior dyke swarms modified from Buchan and Ernst (2004); Superior craton modified from Goodfellow (2007) and Horton *et al.*, (2017). Karelia-Kola basement geology and dyke swarms modified from Amelin *et al.* (1995); Mertanen *et al.* (1999); Lubnina *et al.* (2016); Huhma *et al.* (2018); GTK (2019). Wyoming dyke swarms and basement geology modified from Kilian (2016a) and Horton *et al.* (2017). Shape and relative size of each craton were independently confirmed using GPlates (Müller *et al.*, 2018).

Supercraton Superia (full shape and size of Superia undefined)

The orientation and position of Karelia-Kola with respect to the Superior craton presented in **Figure 2-17** can be reconciled with paleomagnetic studies by Mertanen *et al.* (1999) and Salminen *et al.* (2019). The D pole and D' virtual geomagnetic pole from 2.44 Ga mafic dykes by Mertanen *et al.* (1999) place Karelia-Kola either north (D pole) or south (D' pole) of Superior. The reconstruction by Bleeker and Ernst (2006) favoured the D' pole from Mertanen *et al.* (1999) and placed Karelia-Kola south of Superior. This position of Karelia-Kola with respect to Superior is in agreement with the 2.5 Ga paleomagnetic data from the Mistassini dyke swarm (Fahrig *et al.*, 1986; Buchan *et al.*, 1998), and at 2.4 Ga for which Pisarevsky *et al.* (2015) present a similar (but based on northern hemisphere poles) paleomagnetically permissible paleogeographic configuration. Similarly, Salminen *et al.* (2019) present a configuration based on a compilation of paleomagnetic studies of 2.3-2.7 Ga intrusive rocks in which Karelia-Kola is south of Superior.

Previous versions of the Superia reconstruction have also included the Hearne, Vaalbara (Pilbara and Kaapvaal), Zimbabwe, and Yilgarn cratons (Bleeker, 2004; Bleeker and Ernst, 2006; Ernst and Bleeker, 2010; Söderlund *et al.*, 2010; Pisarevsky *et al.*, 2015; Bleeker *et al.*, 2016; Gumsley *et al.*, 2017). Dyke geometries of the Kaminak swarm, which have been dated at 2500 Ma by Sandeman *et al.* (2013) and were previously dated between 2.45 and 2.50 Ga by Heaman (1994, 2004), were used to position the Hearne craton between Wyoming and Karelia-Kola by linking these swarms to the Mistassini and Matachewan magmatic centres (Ernst and Bleeker, 2010). Recent dates for the Kazan and Chipman dykes at *ca.* 2.11 Ga, however, support a connection to Marathon dykes and indicate that the position of the Hearne craton within Superia should be adjusted (Grohn *et al.*, 2016; Regan *et al.*, 2016). Work by Bleeker *et al.* (2016), Gumsley *et al.* (2017), and

Salminen *et al.* (2019) supports the positioning of Vaalbaara at the southwestern margin of Superior by linking multiple Paleoproterozoic dyke swarms with exact age matches. Yilgarn and Zimbabwe cratons have been tentatively implicated in the Superia reconstruction by Söderlund *et al.* (2010) by matching *ca.* 2411±2 and 2410±2 Ma Widgiemooltha dykes (Yilgarn craton) and 2408 Ma Sebang Poort dykes (Zimbabwe craton) to the contemporaneous Ringvassøy dykes from the West Troms Basement Complex of Northern Norway (Kullerud *et al.*, 2006). Additional geochronological and paleomagnetic data from the Erayinia dykes (Yilgarn craton) by Pisarevsky *et al.* (2015) further support a connection between the Yilgarn, Zimbabwe and Superior cratons. The establishment of *ca.* 2400 Ma Uolevinlehto swarm in Pudasjärvi and into the Taivalkoski and Kuhmo blocks expands the *ca.* 2400 Ma swarm system across Karelia and into Superior via the contemporaneous du Chef dykes. Their converging trends suggest a magmatic centre for these dykes either near the Matachewan magmatic centre, or farther south between Wyoming and Karelia-Kola. The latter location is preferred, as no *ca.* 2400 Ma dykes have been identified in the well-exposed and extensively studied southern Superior craton.

For each of the cratons included in this iteration of the Superia reconstruction (Superior, Wyoming and Karelia-Kola; **Figure 2-17**) the reader will note that their estimated full extent (lighter shades outlined in black dashed lines) are larger than the known/exposed areal extent (darker shades). The Lithoprobe seismic reflection and refraction studies across the accreted margins of the Superior craton and overlying orogenic belts (e.g., Grenville) reveal that Archean crust continues ~200 km below accreted orogenic belts and suggest that the Archean cratonic fragments as they exist now were much larger

prior to breakup and collision (Hammer *et al.*, 2010). Insight from these seismic surveys along the margins of Superior can be applied to other cratonic margins, such as Karelia-Kola, which we speculate continues into the Norbotten craton and then continues below the Caledonian Mountain range to the Norwegian coast.

2.6.6 Alternative reconstruction solutions

Consideration of the analytical uncertainties, particularly with respect to the paleomagnetic data, allows multiple solutions for the reconstruction and configuration of Archean cratons (see reviews by Reddy and Evans, 2009; Evans, 2013). The paleogeographic reconstruction by Pisarevsky *et al.* (2015), which includes the “Zimgarn” reconstruction of Smirnov *et al.* (2013), uses a similar fit between the Superior, Wyoming, and Karelia-Kola cratons as the Superia reconstruction of Bleeker and Ernst (2006) but prefers northern hemisphere poles (the opposite polarity option used for Superia) in their interpretation, and also incorporates the Yilgarn craton. An alternative solution to the Superia and Zimgarn reconstructions is that of a much more extensive but largely unconstrained supercontinent “Kenorland” (Williams *et al.*, 1991; Bleeker, 2003; Lubnina and Slabunov, 2011; Mitchell *et al.*, 2014).

2.6.7 Estimating collisional path during orogenesis

Archean cratons with rifted and/or fragmented margins are common amongst those that have seen later orogenesis. Just as swarms are applied regionally in continental reconstruction models, they can also be used locally to evaluate and quantify uplift and/or rotation of fragmented blocks with respect to their parental cratons (i.e., Pudasjärvi block with respect to Karelia-Kola craton). If the direction and amount of vertical axis rotation is

constrained, then an approximate path of collision can be surmised. A head-on collision normal to the margin of the craton may result in none or only minor rotation, whereas a significant degree of rotation of a coherent block may suggest oblique collision and transpression. Once determined, such a constraint on the trajectory of collision can be compared with other indicators such as sense of movement on regional faults and the apparent polar wander paths of the colliding cratons. In the case of the Pudasjärvi block and the $\sim 30^\circ$ clockwise rotation, this deformation is compatible with large-scale dextral transpression on the Archean–Paleoproterozoic boundary (Ladoga–Bothnian Bay deformation zone, or LBB of Gaál and Gorbatshev, 1987) during the *ca.* 1.92–1.79 Ga Svecofennian orogeny (Sorjonen-Ward, 2006; Lahtinen *et al.*, 2009).

2.7 Conclusions

U-Pb ID-TIMS geochronology and dyke trends are used to refine the classification of six mafic dyke swarms in the Pudasjärvi block, Western Karelia. The *ca.* 2450 Ma Pääjärvi swarm is defined by the 018°N -trending dyke 16SD_014 and the parallel 2444 ± 4 Ma Vengasvaara intrusion (Huhma *et al.*, 2018). The NW-trending *ca.* 2400 Ma Uolevinlehto swarm is introduced due to a distinctly younger estimated crystallization age of *ca.* 2400 ± 12 Ma of the Uolevinlehto dyke. The NW-trending $2135.2 \pm 3.6/-3.7$ Ma Myllykangas and 2128.9 ± 1.2 Ma Sipojuntti dyke ages confirm the existence of an older magmatic pulse related to the *ca.* 2135–2100 Ma Tohmajärvi swarm. The WNW-trending Palomaa swarm is distinguished from the Tohmajärvi swarm by a $\sim 30^\circ$ difference in trend and this distinction is supported by a preliminary Sm-Nd age of 2077 ± 35 Ma by Huhma *et*

al. (2018). Other swarms classified in the Pudasjärvi block are the NNW-trending *ca.* 1980 Ma Paukkajanvaara dykes and the undated “East-West” dykes that must be younger than 2450 Ma based on cross-cutting relationships, but are likely younger than *ca.* 2135 Ma.

Three geochemical groups including Group 1 Fe-tholeiitic basalts, Group 2 Mg-tholeiitic basalts and Group 3 MORB-type basalts are recognized within the dykes and layered mafic intrusions in the Pudasjärvi block and these geochemical groups have been characterized before in other parts of Western and Central Karelia

Structural restoration of the Pudasjärvi block is determined by aligning Tohmajärvi dykes in the Pudasjärvi block to dykes of the same age in the Kuhmo and Taivalkoski blocks. A clockwise $\sim 30^\circ$ vertical-axis rotation is consistent with large-scale dextral transpression on the Archean-Paleoproterozoic boundary (Ladoga-Bothnian Bay deformation zone, or LBB of Gaál and Gorbatshev, 1987) during the *ca.* 1.92-1.79 Ga Svecofennian orogeny (Sorjonen-Ward, 2006; Lahtinen *et al.*, 2009). In the restoration, the *ca.* 2450 Ma Pääjärvi dykes have a regional north-northwest trend and the Uolevinlehto dyke approximately aligns to the Taivalkoski (WD) and Viianki dykes in the Taivalkoski and Kuhmo blocks, respectively. The interpreted Paukkajanvaara dykes in the Pudasjärvi block are parallel to the *ca.* 1980-1963 Ma Kivikevätti (KD) and Paukkajanvaara localities. Correlations in the Taivalkoski and Kuhmo blocks to the Palomaa and East-West dykes, however, have not yet been identified.

Restoration of the Pudasjärvi block provide constraints on regional dyke swarm trends across the Karelia-Kola craton, and these trends are considered within the context of the Superia supercraton paleogeographic reconstruction. The best fit for Karelia-Kola is due south of the Superior craton and southeast of the Wyoming craton, a paleogeographic

position that is compatible with paleomagnetic data. The preferred magmatic centre for the converging *ca.* 2400 Ma Uolevinlehto and du Chef dykes is between the Karelia-Kola and Wyoming craton. This study and the refinement of the Superia supercraton reconstruction raises a number of questions for further research:

- (1) With better constrained block rotation of the Pudasjärvi block, can the Palomaa and East-West dykes be traced into the Taivalkoski and Kuhmo blocks?
- (2) What is the full distribution of the *ca.* 2400 Ma Uolevinlehto event and where is the magmatic centre? Does it include the Ringvassøy dykes in the Norbotten craton and do these perhaps correlate to contemporaneous events in Zimbabwe and Yilgarn?
- (3) What is the significance of the dual geochemistry, age pulses, and trends of the *ca.* 2135-2100 Ma Tohmajärvi magmatic event and how does this compare to the contemporaneous and geochemically diverse Marathon dykes?
- (4) Are the 1980 Ma Paukkajanvaara and Pechenga-Onega LIP (Karelia-Kola craton) and Mine Centre (Superior craton) dykes a valid match? If so, is this an indication of delayed break-up, until 1980 Ma or later, between the Karelia-Kola and Superior cratons?
- (5) Is there a genetic connection between the *ca.* 3.5 Ga Siurua gneiss in Karelia-Kola and the ancient core of the Wyoming craton?

Chapter 3

Structural controls on the trends of ca. 2.32 and 1.98 Ga dykes, Western Karelia

3.1 Abstract

Paleoproterozoic dyke swarms in the Taivalkoski block of Western Karelia are classified using U-Pb geochronology and field observations combined with the interpretation of aeromagnetic maps. This classification identifies curvilinear dyke swarm patterns for the *ca.* 2.32 Ga Taivalkoski-Kuito and *ca.* 1.98 Ga Paukkajanvaara swarms within the Taivalkoski block. Consistent curvature in both sets of dykes is interpreted to have occurred as a result of deformation related to sinistral displacement along the Oulujärvi shear zone.

This analysis is made possible by the addition of seven new U-Pb ages on mafic dykes in Western Karelia. The *ca.* 2.40 Ga Uolevinlehto swarm is dated at the Viianki locality (Kuhmo block) with an age of 2400.7 ± 1.8 Ma. The *ca.* 2.32 Ga Taivalkoski-Kuito swarm is dated at the Tulisaari dyke (Iisalmi block) at 2320 ± 13 Ma, Kakikangas dyke (Taivalkoski block) at 2329 ± 8.2 Ma, and Tilsanvaara dyke (Taivalkoski block) with a minimum age of 2292 ± 8.8 Ma. The Tilsanvaara locality includes a younger cross-cutting dyke and a preliminary analysis suggests that this younger cross-cutting dyke is 1972.8 ± 11.2 Ma and therefore belongs to the *ca.* 1.98 Ga Paukkajanvaara swarm. The *ca.* 1.98 Ga Paukkajanvaara swarm is also dated at the new Eno locality (Kuhmo block) and a

result of 1975.6 ± 9.7 Ma for the Kivikevätti dyke (Kuhmo block) is slightly younger than previous studies report. Whole-rock major and trace element data are presented for these dykes which have Fe-tholeiitic and high Mg-tholeiitic signatures, typical of Karelian dykes.

3.2 Introduction

Despite an abundance of mafic dykes that cut across the Archean basement of the Karelia-Kola craton, the distribution of regional Paleoproterozoic mafic dyke swarms remains poorly defined. Limited exposure, in addition to crustal reworking by the *ca.* 1.92-1.79 Ga Svecofennian orogeny, has meant that conventional dyke swarm mapping techniques, from field observations or by interpretation of aeromagnetic maps, do not or cannot provide sufficient information to independently inform dyke swarm assignments accurately.

The boundary between the Taivalkoski and Kuhmo blocks, defined by the Oulujärvi shear zone (OSZ), of Western Karelia is an example where there is an abundance of dykes with various trends that have not yet been assigned to regional swarms. Within the Kuhmo block, dyke swarms include the *ca.* 2.40 Uolevinlehto, *ca.* 2.32 Ga Taivalkoski-Kuito, *ca.* 2.11 Ga Tohmajärvi, and *ca.* 1.98 Ga Paukkajanvaara (Pekkarinen and Lukkarinen, 1991; Vuollo *et al.*, 1992; Stepanova *et al.*, 2017; Huhma *et al.*, 2018). The northern extent of these dyke swarms, however, is truncated by the OSZ. Previous interpretation of aeromagnetic maps indicates that the Taivalkoski block is host to a high density of dykes, but their subtly different trends compared to those in the adjacent Kuhmo block and sparse U-Pb geochronology have prevented their classification.

In this chapter, we present U-Pb geochronology on seven mafic dykes from the Taivalkoski, Kuhmo and Iisalmi blocks, and discuss relevant whole-rock major and trace element geochemistry for coeval dykes in Western Karelia. These new ages are used to confirm the regional trends of the EW-trending *ca.* 2.40 Ga Uolevinlehto, WNW-trending *ca.* 2.32 Ga Taivalkoski-Kuito and NW-trending *ca.* 1.98 Ga Paukkajanvaara dyke swarms across the Taivalkoski and Kuhmo blocks. After consideration of the plausible mechanisms causing the curvilinear trends observed in the *ca.* 2.32 Ga Taivalkoski-Kuito and *ca.* 1.98 Ga Paukkajanvaara swarms, we propose that the dykes were deformed by ductile deformation related to sinistral displacement along the OSZ.

3.3 Geological Setting

3.3.1 Archean Karelian Province

The Karelian Province is an Archean component in the Fennoscandian (Baltic) shield consisting largely of granitoid complexes and lesser greenstone belts, paragneisses and granulites (Gaál and Gorbatshev, 1987; Gorbatshev and Bogdanova, 1993; Sorjonen-Ward and Luukkonen, 2005; Slabunov *et al.*, 2006b; Hölttä *et al.*, 2008). The boundaries of the Karelian Province are marked to the northeast by the Belomorian mobile belt (Lapland-Kola) and to the southwest by the Svecofennian domain (Lahtinen *et al.*, 2008; Lahtinen, 2012). Regional differences in lithology, structure and age have resulted in the division of Karelia into three separate subprovinces: Vodlozero, Central and Western Karelia (Slabunov *et al.*, 2006b; Hölttä *et al.*, 2012).

Western Karelia covers northeastern Finland and continues into western Russia. Recognized for its fragmented crustal architecture, Western Karelia includes four distinct crustal blocks: Pudasjärvi, Iisalmi, Kuhmo and Taivalkoski (**Figure 3-1**; e.g., Vuollo and Huhma, 2005). At the core of the Pudasjärvi block is the 3.5 Ga Siurua gneiss, which is the oldest known rock in Fennoscandia (Mutanen and Huhma, 2003). The bulk of the Pudasjärvi block is composed of *ca.* 2.82-2.73 Ga TTGs (Trondjemite-Tonalite-Granodiorite) and these are bisected by the narrow NS-trending 2.82-2.80 Ga Oijärvi greenstone belt (Mutanen and Huhma, 2003; Hölttä *et al.*, 2012; Huhma *et al.*, 2012). Other subordinate rocks in the Pudasjärvi block include 2.74-2.73 Ga paragneisses, 2.70-2.62 granites, and the 2.70 Ga Ranua diorite-quartz intrusion (Mutanen and Huhma, 2003; Huhma *et al.*, 2012). The Iisalmi block is located south of the Pudasjärvi block and includes *ca.* 3.2-3.1 Ga tonalite gneisses with very small slivers of *ca.* 2.70 Ga diorite and Archean amphibolite (Hölttä *et al.*, 2000a, 2012; Paavola, 2003). To the east of the Pudasjärvi and Iisalmi blocks are the Taivalkoski and Kuhmo blocks, which are occasionally referred together as the Kianta block or Lentua complex (Heilimo *et al.*, 2011; Hölttä *et al.*, 2012). While the Taivalkoski block consists of massive *ca.* 2.70-2.87 Ga granitoid suites (Lauri *et al.*, 2006, 2011; Heilimo *et al.*, 2011), the Kuhmo block is significantly more complex. Variably deformed granitoids (TTGs, granites and sanukitoids) range in age from *ca.* 2.96 to 2.67 Ga (Luukkonen, 1985; Vaasjoki *et al.*, 1999; Käpyaho *et al.*, 2006; Heilimo *et al.*, 2011; Mikkola *et al.*, 2011). Within these granitoids is the narrow (15 km wide) *ca.* 2.80-2.95 Ga Tipasjärvi-Kuhmo-Suomussalmi greenstone belt that stretches 220 km across the Kuhmo block (Lehtonen *et al.*, 2017 and references therein).

Figure 3-1. Geochronology locations and trends of *ca.* 2.40 Ga, 2.32 Ga, and 1.98 Ga mafic dykes in Western Karelia. Details of locations A-V are listed in **Table 3-1**. Ages in bold and italicized text are those presented in this study. Bedrock geology, dykes and geochronology are compiled from Huhma *et al.* (1990, 2018), Mertanen *et al.* (1999), Hölttä *et al.* (2000); Vuollo and Huhma (2005), Hölttä *et al.* (2008), GTK (2012), Lahtinen (2012), Salminen *et al.* (2014), and Stepanova *et al.* (2015).

With the boundaries between blocks defined by major Paleoproterozoic shear zones and associated schist belts, the genetic relationships between blocks have yet to be fully established (Pesonen and Klein, 2015). This includes questions such as whether the Pudasjärvi block is an exotic terrane (Pesonen and Klein, 2015) and/or if the blocks were derived from a common landmass that has since been deformed (Vuollo *et al.*, 2000; Vuollo and Huhma, 2005; Hölttä *et al.*, 2012). The structures and schists belts defining the boundaries between the Archean blocks also cross-cut Paleoproterozoic mafic dyke swarms such that the dyke swarms cannot be followed unambiguously across these structures on the ground or by interpretation of aeromagnetic maps.

3.3.2 Paleoproterozoic mafic dykes

Hundreds of Paleoproterozoic mafic dykes between *ca.* 2.50 and 1.98 Ga intrude the Archean basement of Western Karelia (**Figure 3-1**; e.g., Mertanen *et al.*, 1999; Vuollo and Huhma, 2005; Bleeker *et al.*, 2008; Salminen *et al.*, 2014; Stepanova *et al.*, 2014, 2015, 2017). There are at least seven discrete mafic dyke swarm and sill events in Western Karelia including the *ca.* 2.50 Ga Shalskiy dyke, *ca.* 2.45 Ga Pääjärvi dykes and intrusions, *ca.* 2.40 Uolevinlehto dykes, *ca.* 2.32 Ga Taivalkoski-Kuito dykes, *ca.* 2.22 Ga Karjalitic sills, *ca.* 2.11 Ga Tohmajärvi-Pirtguba dykes, and *ca.* 1.98 Ga Paukkajanvaara dykes (Vuollo and Huhma, 2005; Bleeker *et al.*, 2008; Stepanova *et al.*, 2017; **Chapter 2**). Each of these intraplate mafic magmatic events are attributed to individual mantle plumes which lead to various degrees of continental rifting, however, some studies propose alternative mechanisms for rifting by delamination of the lower crust (Vuollo and Huhma, 2005; Stepanova *et al.*, 2014, 2015).

Table 3-1. Summary of *ca.* 2.40 Ga, *ca.* 2.32 Ga, and *ca.* 1.98 Ga dated mafic rocks, Western Karelia.

ID	Swarm/age Locality name/sample	Block	Latitude (°N)	Longitude (°E)	Age (Ma)	Geochronology Method and notes	Trend (°)	Dip (°)	Width (m)	Magn. susc. ($\times 10^{-3}$ SI units) Avg $\pm 1\sigma$ (n)	ϵ_{Nd}	Refs
<i>Uolevinlehto, ca. 2400 Ma</i>												
A	WD (15SD_073)	Taivalkoski	65.5785	28.2641	2407 $\pm$ 35	Sm-Nd whole-rock and mineral internal isochron	090	74S	10	~30	+1.6	1
B	Uolevinlehto (15SD_071)	Pudasjärvi	65.6902	27.0607	2400 $\pm$ 12	U-Pb, ID-TIMS single and multiple grain Bdy	325	~90	30	30 $\pm$ 3 (5)	+0.4	1, 2
C	Viianki (15SD_084)	Kuhmo	64.7731	30.0694	2400.7 $\pm$ 1.8	U-Pb, ID-TIMS multiple grain Bdy	080	~80N	25	25 $\pm$ 10 (5)	-1.2	1, 3, 4
D	Ca-250-1	Kuhmo	-	-	2404 $\pm$ 5	U-Pb, ID-TIMS multiple grain Bdy	095	-	50	-	-0.9	5
<i>Kuito-Taivalkoski, ca. 2330 Ma</i>												
E	Kallioniemi	Taivalkoski	65.5012	28.5752	2352 $\pm$ 62	Sm-Nd whole-rock and mineral internal isochron	NE	-	-	45 $\pm$ 5 (5)	+0.5	1, 4
F	XD4	Taivalkoski	66.1558	30.9896	2349 $\pm$ 24	Sm-Nd whole-rock and mineral internal isochron	W	~90	~10†	-	+1.0	1, 6
G	Törinkuru AD13 (15SD_072)	Taivalkoski	65.5249	28.0486	2333 $\pm$ 10	U-Pb TIMS, Multiple grain Bdy	280	90	25	42 $\pm$ 17 (12)	+1.7	1, 7
H	Runkaus volcanics	-	66.0542	25.4236	2332 $\pm$ 180 2332 $\pm$ 19/-3	Sm-Nd whole-rock and mineral internal isochron	-	-	-	-	0.6 $\pm$ 1.1	8
I	Taivalkoski AD10 (15SD_074)	Taivalkoski	65.5650	28.3335		U-Pb TIMS, Multiple grain Bdy	280	-	10	25 $\pm$ 5 (5)	+1.6	4
J	Kakikangas (15SD_075)	Taivalkoski	65.5183	28.6217	2329 $\pm$ 8.2	U-Pb TIMS, Multiple grain Bdy	240	-	50	40 $\pm$ 10 (5)	-	3
K	Humppi (15SD_080)	Iisalmi	63.3329	27.4272	2323 $\pm$ 13	Laser ablation MC-ICP-MS, Zrc	290	90	>70	25 $\pm$ 10 (10)	+1.3	1
L	Lohisärkkä	Kuhmo	65.0446	29.3366	2321 $\pm$ 21	U-Pb TIMS, Multiple grain Bdy	285†	-	200	-	+1.5	1
M	Tulisaari (15SD_083)	Iisalmi	63.5073	27.8859	2320 $\pm$ 13	U-Pb TIMS, Multiple grain Bdy	295	~90	200	40 $\pm$ 10 (10)	+1.6	1, 3, 9
N	Karhuvaara A1672	Kuhmo	64.4128	28.5901	2314 $\pm$ 6	U-Pb LA-ICPMS, Zrc	NW	-	>60	-	-	1
O	Ca-240-5	Kuhmo*	65.2170	30.2820	2311 $\pm$ 5	U-Pb ID-TIMS, Multiple grain Bdy	305†	-	50	-	+0.7	10
P	Ca-245-3	Kuhmo*	64.9211	30.4866	2309 $\pm$ 3	U-Pb ID-TIMS, Multiple grain Bdy	285†	-	45†	-	+0.5	10
Q	<i>Karkuvaara A365 (15SD_079)</i>	Taivalkoski	65.2736	27.7290	2306 $\pm$ 8	U-Pb TIMS, Multiple grain Bdy	-	-	-	-	+1.8	1
R	WNW Tilsanvaara (15SD_076)	Taivalkoski	65.4613	28.0752	2292 $\pm$ 8.8	U-Pb ID-TIMS, Multiple grain Bdy	275	90	-	~5-35	+1.0	1, 3

Table 3-1 continues on the next page

Notes: Localities A-V shown in **Figure 3-1**. Abbreviations: Bdy - baddeleyite; Zrc - Zircon; CA - Chemical abrasion; ID-TIMS - Isotope dilution thermal ionisation mass spectrometry. Dips reported with a "~" indicate those interpreted from contact-parallel parting planes and/or columnar jointing. † indicates value obtained from a map. Refs (references): 1. Huhma et al. (2018), 2. **Chapter 2**; 3. This study, 4. Vuollo and Huhma (2005), 5. Stepanova et al. (2017), 6. Mertanen et al. (1999), 7. Salminen et al. (2014), 8. Huhma et al. (1990), 9. Hölttä et al. (2000), Stepanova et al. (2015), 11. Vuollo et al. (1992).

Table 3-1. Continued.

ID	Swarm/age Locality name/sample	Block	Latitude (°N)	Longitude (°E)	Age (Ma)	Geochronology Method and notes	Trend (°)	Dip (°)	Width (m)	Magn. susc. ($\times 10^{-3}$ SI units) Avg $\pm 1\sigma$ (n)	ϵ_{Nd}	Refs
<i>Paukkajanvaara, ca. 1980 Ma</i>												
R	<i>NW Tilsanvaara (15SD_077)</i>	Taivalkoski	65.4613	28.0752	1972.8 $\pm$ 11.2	U-Pb ID-TIMS, CA-Zrc	295	90	10	~40-75	+0.2	1, 3
S	<i>Kivikevätti (15SD_085)</i>	Kuhmo	64.7549	29.7921	1975.6 $\pm$ 9.7	U-Pb ID-TIMS, CA-Zrc, multiple/single grain Bdy	330	83E	75	97 $\pm$ 8 (6)	+0.3	1, 3
T	Paukkajanvaara	Kuhmo	62.9312	30.0547	1965 $\pm$ 10	U-Pb ID-TIMS, Zrc	-	-	-	-	+1.8	11
U	<i>Eno dyke (15SD_088)</i>	Kuhmo	62.7440	30.4260	ca. 1980	U-Pb ID-TIMS, CA-Zrc	~305	-	>70	0.8	-	3
<i>Undated localities</i>												
V	<i>16SD_001</i>	Iisalmi	63.1039	27.7310			320	90	60			3

Notes: Localities A-V shown in **Figure 3-1**. Abbreviations: Bdy - baddeleyite; Zrc - Zircon; CA - Chemical abrasion; ID-TIMS - Isotope dilution thermal ionisation mass spectrometry. Dips reported with a "~" indicate those interpreted from contact-parallel parting planes and/or columnar jointing. † indicates value obtained from a map. Refs (references): 1. Huhma et al. (2018), 2. **Chapter 2**; 3. This study, 4. Vuollo and Huhma (2005), 5. Stepanova et al. (2017), 6. Mertanen et al. (1999), 7. Salminen et al. (2014), 8. Huhma et al. (1990), 9. Hölttä et al. (2000), Stepanova et al. (2015), 11. Vuollo et al. (1992).

Only a handful of dykes have been dated with ages of *ca.* 2.40 Ga in Karelia (**Table 3-1** and **Figure 3-1**). These include the subparallel EW-trending *ca.* 2378 Ma Viianki (Huhma *et al.*, 2018) and the 2404±5 Ma Ca-250-1 dyke in Russian Karelia (Stepanova *et al.*, 2017). In the Taivalkoski block, the WD dyke has a similar EW trend but is dated by a Sm-Nd whole-rock and mineral internal isochron at 2407±35 Ma (Vuollo and Huhma, 2005; Huhma *et al.*, 2018). In contrast, the Uolevinlehto dyke has a distinct 325° trend, but has a U-Pb baddeleyite age of *ca.* 2400±12 Ma, indicating that it is coeval (**Chapter 2**).

Thirteen *ca.* 2.32 Ga mafic dykes have been reported in the Kuhmo, Taivalkoski and Iisalmi blocks (**Table 3-1**). The majority of these (10) have NW trends and sub-vertical dips. Outliers include the NE-trending Kallioniemi and Kakikangas dykes which are proximal and subparallel to the Oulujärvi shear zone (OSZ; **Figure 3-1**), as well as the W-trending XD4 dyke of Mertanen *et al.* (1999). The dykes with uncharacteristic trends, however, have only been dated previously using Sm-Nd whole-rock and mineral internal isochron at the Kallioniemi (2352±62 Ma) and XD4 (2349±24 Ma) localities (**Table 3-1**; Mertanen *et al.*, 1999; Vuollo and Huhma, 2005).

The *ca.* 1.98 Ga Pechenga-Onega LIP is the final pulse of intraplate mafic magmatism before continental break-up of the Fennoscandian shield and opening of the Lapland-Kola and Svecofennian oceans (Peltonen *et al.*, 1996, 1998; Lubnina *et al.*, 2017). Ocean opening is marked by *ca.* 1.97-1.95 Ga ophiolites preserved within the Lapland-Kola and Svecofennian accreted terranes (Peltonen *et al.*, 1996, 1998; Lubnina *et al.*, 2017). A dense swarm of NW-trending *ca.* 1.98-1.96 Ga mafic dykes in the Kuhmo and Pudasjärvi blocks are associated with the Pechenga-Onega LIP, covering an areal extent of 600,000 km² (Vuollo and Huhma, 2005; Lubnina *et al.*, 2016b). In the Kuhmo block, the

dykes have prominent aeromagnetic signatures that can be traced for tens of kilometres but these signatures end abruptly at the Taivalkoski-Kuhmo border defined by the OSZ shear zone. Until this study, no U-Pb ages have been reported at *ca.* 1.98 Ga for dykes in the Taivalkoski block. The presence of *ca.* 1.98 Ga dykes in the Pudasjärvi blocks is inferred from aeromagnetic map interpretation, which show prominent NNW-trending lineaments cutting across the block. These lineaments are truncated at the southern and northern margins of the Taivalkoski block. Lacking prominent lineaments on aeromagnetic maps and U-Pb ages, no dykes of *ca.* 1.98 Ga affinity have been identified in the Iisalmi block.

3.3.3 Svecofennian deformation

Pervasive deformation along the southwestern margin of the Karelian craton is attributed to multiple episodes of arc and microcontinent collision from *ca.* 1.92 to 1.79 Ga which compose the Svecofennian orogeny (Lahtinen *et al.*, 2009; Torvela and Ehlers, 2010). The Svecofennian orogeny includes distinct accretionary events with variable convergence pathways (first from the northeast, then south to south-southeast transpression and finally convergence from the southeast, south and southwest (Torvela and Ehlers, 2010). Kärki *et al.* (1993) has characterized the resulting shear zones and schist belts in four stages of deformation. Stages D₁ and D₂ are related to folding and thrusting during convergence followed by stages D₃ and D₄, which record the development of the Savolappi and Finlandia ductile shear systems (Kärki *et al.*, 1993). These D₃ (Savolappi) and D₄ (Finlandia) ductile shear systems are responsible for the structures that truncate Paleoproterozoic dyke swarms and form the tectonic boundaries between the Pudasjärvi, Iisalmi, Taivalkoski and Kuhmo crustal blocks (**Figure 3-1** and **Figure 3-2**).

Figure 3-2. Curvature of the *ca.* 2.32 Ga Taivalkoski-Kuito and *ca.* 1.98 Ga Paukkajanvaara dykes. a. Geological sketch of the dykes, many of which are mapped from tracing aeromagnetic lineaments with high magnetic signatures. Associations and trends are ground-truthed at geochronology locations. b. Aeromagnetic map with the Oulujärvi shear zone and related structures (thick black lines) for reference. The grey-scale aeromagnetic units are reported in nanoteslas, nT. Map data (dykes, structures, aeromagnetic) modified from Vuollo and Huhma (2005), GTK (2012, 2016), Salminen *et al.* (2014). The red rectangle indicates the extent of Figure 3-14.

According to Kärki *et al.* (1993), the Savolampi shear system is defined by the N-S trending Hirvaskoski (HSZ), Pajala, and Russian North Karelia shear zones, each with a dextral sense of displacement. The HSZ divides the Pudasjärvi block from the Taivalkoski block by a series of steeply dipping parallel faults (**Figure 3-1**). The northern end of the HSZ dissipates as a horsetail splay in the Kuusamo schist belt north of the Taivalkoski block. The southern end is terminated by the younger D₄ Oulujärvi shear zone (OSZ; Kärki *et al.*, 1993). The late stage Finlandia (D₄) shear system is defined by the 250 km-long OSZ (**Figure 3-1**; Kärki *et al.*, 1993). The NE-trending subvertical shear zone marks the boundary between the Pudasjärvi and Iisalmi blocks, cuts through and displaces the stratigraphy of the Kainuu schist belt, and continues westward into the Archean basement, dividing the Taivalkoski and Kuhmo blocks (Kärki *et al.*, 1993). In the Archean basement of the Taivalkoski and Kuhmo block, the OSZ can be traced for approximately ~70 km where it ends as a brittle fault system (Kärki *et al.*, 1993). Additional minor brittle faults can be found along trend of the OSZ for at least another 70 km towards the Russian border.

3.4 Methods

3.4.1 U-Pb geochronology

About 10 kg of medium to coarse-grained gabbro was collected from the Viianki, Kakikangas, Tulisaari, Tilsanvaara, Kivikevätti, and Eno dykes. All procedures related to geochronology were completed at the Jack Satterly Geochronology Laboratory, University of Toronto. A steel jaw crusher and pulverizer were used to crush the rock from which

heavy mineral separates were prepared using a Wilfley table, magnetic separator and heavy liquids. Zircon and baddeleyite crystals were picked under a binocular microscope. Analyzed zircons were chemically abraded between one (Kivikevätti and Eno) and six (Tilsanvaara) hours following the procedures by Mattinson (2005). Each baddeleyite or zircon fraction was rinsed with HNO₃ before being dissolved in 0.10 ml of hydrofluoric acid (HF) and 0.02 ml of 8N nitric acid (HNO₃) at 195°C. During this sample loading step, a ²⁰⁵Pb-²³⁵U spike was added to the Teflon dissolution capsules. After dissolution (3-5 days) samples were dried down and re-dissolved in 3N HCl (Krogh, 1973). Then U and Pb were isolated from the solution using 50 µl anion exchange columns. The remaining elutions were dried down with phosphoric acid (H₃PO₄) and then loaded onto outgassed rhenium filaments with silica gel (Gerstenberger and Haase, 1997). Each sample was analyzed with a VG354 mass spectrometer using a Daly detector set on pulse counting mode. Decay constants are those of Jaffey *et al.* (1971) and the Excel add-in software Isoplot 3.00 (Ludwig, 2003) was used for age calculations and plotting. Results of standards (R33, OG1 and EARTHTIME 2.0 Ga synthetic solution of Black *et al.*, 2004; Condon *et al.*, 2015. and Stern *et al.*, 2019) analyzed over the duration of this study are presented in **Appendix C**.

3.4.2 Geochemistry

Whole-rock major and trace element geochemical data were acquired for 11 mafic dykes from Western Karelia (**Table 3-3**). At Carleton University, samples were crushed to <2 mm in a Mn-Steel jaw crusher and then pulverized in an agate mill. Powders weighing at least 5 g were sent to ALS Laboratories (www.alsglobal.com) for analysis of 13 major

oxides, and 42 minor and trace elements by inductively coupled plasma-atomic emission spectrometry (ICP-AES) and inductively coupled plasma-mass spectrometry (ICP-MS). Trace element samples were prepared with the lithium borate fusion method before dissolution and analysis by ICP-MS.

3.5 Results

Eleven mafic dykes were sampled in the Taivalkoski, Kuhmo and Iisalmi blocks in Western Karelia. With the exception of the Eno dyke and site 16SD_001, which was sampled at the Siilinjärvi carbonatite open pit, all other dykes have been the subject of previous studies (**Table 3-1** and references).

3.5.1 Field observations

3.5.1.1 Kakikangas, 15SD_075

The Kakikangas dyke site is a low-lying outcrop located off a gravel road. With magnetic susceptibility measurements of approximately 40×10^{-3} SI, the dyke has a prominent aeromagnetic signature and can be readily identified on an aeromagnetic map. The associated lineament identified on the aeromagnetic map indicates that the dyke has a trend of about 45° , which can be followed for up to 200 m (locality E; **Table 3-1**; **Figure 3-2**). Outcrop exposure suggests that the medium- to coarse-grained Kakikangas diabase dyke is at least 50 m wide.

3.5.1.2 Tilsanvaara dyke site, 15SD_076 and 15SD_077

Two dykes are found together at the Tilsanvaara locality in the Taivalkoski block (Huhma *et al.*, 2018). The dykes form a linear ridge and a 40-60 m high cliff. Both dykes and their intersection are visible on aeromagnetic maps, which show trends of 280° and 295°. The younger 295°-trending dyke (15SD_077) has a fine-grained texture and is chilled against the older 280°-trending coarse-grained dyke (15SD_076; locality R; **Table 3-1; Figure 3-1; Figure 3-3a**). Based on exposed outcrop, the younger dyke is about 10 m wide and has a sub-vertical dip.

3.5.1.3 Tulisaari, 15SD_083

Previously described and dated by Hölttä *et al.* (2000b) and Huhma *et al.* (2018), the Tulisaari dyke is located in the Iisalmi block (locality M; **Table 3-1; Figure 3-1**). The average texture of the dyke is fine- to medium-grained but coarser-grained sections are found near the presumed centre of the dyke. While contacts were not observed, the width of the dyke is estimated at 30 m based on exposed outcrop. Observation of parting planes suggest that the dyke is sub-vertical and has a 295° trend.

3.5.1.4 Viianki, 15SD_084

The Viianki dyke is a coarse-grained gabbro norite located in the Kuhmo block (locality C; **Table 3-1; Figure 3-1; Vuollo and Huhma, 2005; Huhma et al., 2018**). Intruding into Archean TTG gneiss, the contact of the Viianki dyke is irregular and lacks a chilled margin (**Figure 3-3**). Felsic rheomorphic dykes are commonly found within 0.5 metres of the

contact. The average trend of the 25 m-wide sub-vertical dyke is 080°, however, the trend of the dyke is interrupted by intermittent “steps” where the dyke jogs northward or southward by 10 to 15 m (**Figure 3-3c**).

3.5.1.5 Kivikevätti, 15SD_085

A quarry exploiting the Kivikevätti dyke for gravel exposes the 330°-trending dyke and the crackled texture of its chilled margin (locality S; **Table 3-1; Figure 3-1; Figure 3-3d**).

The contact between the TTG gneiss basement and the Kivikevätti dyke is dipping 83°E. Only one contact was exposed, but based on outcrop exposure, the dyke is estimated to be at least 75 m wide. The approximate location of the centre of the dyke includes abundant coarse grained patches and pegmatoidal pods.

3.5.1.6 Eno, 15SD_088

Found in a quarry, the Eno locality 15SD_088 is interpreted as a dyke based on the weak linear aeromagnetic signature that is associated with this site (locality U; **Table 3-1; Figure 3-3**). The linear magnetic high can be traced for 12 km and indicates that the dyke has a NW trend. While no contacts were found, the dyke is at least 50 m wide based on the exposure at the quarry.

Figure 3-3. Dyke contacts observed in the field. a. Tilsanvaara locality showing the contact relationship between the older (15SD_076) NW-trending dyke against the younger EW-trending dyke (15SD_077) which cuts across and is chilled against the older dyke. b. The irregular contact of the Viianki dyke (15SD_084) against Archean basement gneiss showing rheomorphic dykes intruding into the Viianki dyke. c. A schematic diagram illustrating the "stepping" nature of the dyke. d. Cracked texture of the chilled margin of the Kivikevätti dyke. e. Showing the vertical mafic dykes at the Siilinjärvi carbonatite pit.

3.5.1.7 Siilinjärvi dyke, 16SD_001

Dyke site 16SD_001 is exposed in the open pit at the Siilinjärvi carbonatite mine (locality V; **Figure 3-1**; **Figure 3-3e**). The dyke is vertical and 60 m wide with a NW trend. The contacts of the dyke are heavily sheared, and the average texture is fine-grained with rare medium-grained sections.

3.5.2 Petrographic characteristics

3.5.2.1 Uolevinlehto swarm, *ca.* 2.40 Ga

The *ca.* 2.40 Ga Viianki (15SD_084) gabbro norite dyke is nearly pristine with very little alteration (**Figure 3-4**). Major rock forming minerals in the Viianki dyke include plagioclase (60%), clinopyroxene (20%), orthopyroxene (5%), olivine (5%) and oxides (5%) forming an intergranular coarse-grained texture. Vuollo and Huhma (2005) have reported similar modal abundances for the major minerals in these rocks, and have also observed cumulate textures. Accessory phases (5%) in the Viianki dyke include Fe-Ti oxides (ilmenite), magnetite, pyrite, chalcopyrite, apatite, quartz, biotite, baddeleyite and zircon. Baddeleyite crystals are uniformly small (up to 50 μm long but average length is around 30 μm) and range from anhedral to euhedral with variable amounts of zircon overgrowth (**Figure 3-4f**).

3.5.2.2 Taivalkoski-Kuito swarm, *ca.* 2.32 Ga

Dykes belonging to the *ca.* 2.32 Ga swarm, including the 280°-trending Tilsanvaara, 240°-

trending Kakikangas and the 295°-trending Tulsaaari dykes, exhibit similar mineralogy with variable intensities of alteration (**Figure 3-5** to **Figure 3-8**). Major rock forming minerals in these diabase dykes are plagioclase (50%), clinopyroxene (20-25%), primary and secondary amphibole (25-35%) and minor Fe-Ti oxides (5%). Accessory phases (<5%) include apatite, chalcopyrite, pyrite, zircon and baddeleyite. Alteration is minor but includes sericitized plagioclase, and chlorite and amphibole after primary mafic minerals. While grain size varies from aphanitic at the margins to coarse near the centre of the dykes, textures are typically ophitic to sub-ophitic and less commonly intergranular. At the Tilsanvaara locality, the older 280°-trending dyke (15SD_076) is cut by thin 0.1-1.0 mm veinlets (**Figure 3-5a and b**; **Figure 3-6a and b**). Predominantly consisting of plagioclase and minor amphibole and epidote-group minerals, the source of these veinlets is traced to the younger 295°-trending dyke (15SD_077) that cross-cuts it (**Figure 3-5**).

Minerals used for dating the *ca.* 2.32 Ga dykes including baddeleyite and zircon are accessory phases varying significantly in terms of abundance. In the 280°-trending Tilsanvaara dyke (15SD_076), small prismatic zircon crystals are abundant and these are found in veinlets, as well as within the main matrix (**Figure 3-6**). Regardless of the host, these sub- to euhedral zircons have a homogeneous texture and very few inclusions. Baddeleyite in the Tilsanvaara sample is rare and limited to the matrix of the rock (no baddeleyite crystals were found in the veinlets). Hosted within amphibole or in the interstices between plagioclase and quartz crystals, baddeleyite is found in association with substantial zircon overgrowth (**Figure 3-6f**). In similarity to the zircons found in the matrix, the zircon overgrowth includes sub- to euhedral crystal aggregates with homogeneous textures. In the Kakikangas (15SD_075) and Tulsaaari (15SD_083) dykes, baddeleyite

grains are found in the matrix associated with quartz, hornblende, and plagioclase (**Figure 3-7** and **Figure 3-8**). Baddeleyite in the Kakikangas dyke (15SD_075) is sub to euhedral with minor zircon overgrowth (**Figure 3-7**). In contrast, the baddeleyite grains from the Tulisaari dyke have substantial zircon overgrowth (**Figure 3-8**).

3.5.2.3 Paukkajanvaara swarm, *ca.* 1.98 Ga

The 1.98 Ga dykes have similar mineralogy to those of the 2.32 Ga dykes primarily consisting of plagioclase (50%), clinopyroxene (30%), amphibole (10%) and biotite (5%) in ophitic to subophitic textures (**Figure 3-9**). Alteration is variable from pervasive in the Eno dyke to minor in the Kivikevätti dyke but generally consists of chlorite and amphibole alteration replacing primary mafic minerals and sericite alteration of plagioclase. Accessory phases (5%) include zircon and apatite (which is much more abundant compared to the *ca.* 2.32 Ga dykes) as well as opaque minerals such as ilmenite, magnetite, chalcopyrite, and pyrite. Unique to the Kivikevätti dyke, clinopyroxene occurs with simple twinning, and myrmekitic and graphic textures are found between grain boundaries and interstices between plagioclase, alkali feldspar and quartz.

Figure 3-4. Petrographic details of the Viianki dyke (15SD_084). a. Intergranular texture of plagioclase (Plag), clinopyroxene (Cpx) and hornblende (Hbl), plane polarized light. b. As in a, crossed nicols. c. Backscattered electron image showing baddeleyite and zircon hosted adjacent to hornblende (Hbl) and plagioclase (Plag). d. Baddeleyite with zircon overgrowth, Bdy, Zrc as in c. e. Anhedral baddeleyite grain within quartz, as in c. f. Picked baddeleyite grains.

Figure 3-5. Petrographic details of the WNW-trending (15SD_076) and NW-trending (15SD_077) cross-cutting dykes at the Tilsanvaara locality. Left panels: plane polarized light; right panels: crossed nicols. a-b. Intergranular texture of the older, WNW-trending dyke with plagioclase (Plag), hornblende (Hbl) and clinopyroxene. c-d. Contact between the older and coarse grained WNW-trending and younger fine-grained NW-trending dykes. A veinlet from the younger NW-trending dyke is shown intruding into the older, coarse-grained WNW-trending dyke. e-f. Subophitic-fine-grained texture of the younger NW-trending dyke (15SD_077).

Figure 3-6. Petrographic details of the zircon and baddeleyite found in the older WNW-trending (15SD_076) dyke at the Tilsanvaara locality. a. Photomicrograph showing the intergranular texture of the WNW-trending (15SD_076) dyke cut by a 0.1-0.5 mm veinlet, plane polarized light. The approximate extent of panel b is marked by the black rectangle. b. Backscattered electron image of the area indicated in a. showing the interface between the veinlet and host rock with sub- to euhedral zircon crystals in the veinlet. c. Backscattered electron image showing zircon hosted in hornblende (Hbl). Other minerals present include ilmenite (Ilm), chalcopyrite (Cpy), quartz (Qtz), and plagioclase (Plag). d. Backscattered electron image of zircon (Zrc) shown in c. e. Matrix-hosted baddeleyite (Bdy) and zircon crystals. f. Baddeleyite and zircon crystals in e. showing zircon as an overgrowth on baddeleyite. g. Photomicrograph of picked zircons. h. Photomicrograph of picked baddeleyite.

Figure 3-7. Petrographic details of the NE-trending Kakikangas dyke (15SD_075). a. Subophitic texture, plane polarized light. b. As in a, crossed nichols. c. Backscattered electron image showing baddeleyite hosted in alkali feldspar and hornblende. d. Baddeleyite, Bdy-1 as in c. with minor zircon overgrowth. e. Baddeleyite Bdy-2 as in c. f. Baddeleyite Bdy-3 as in c. g. Picked baddeleyite grains.

Figure 3-8. Petrographic details of the NW-trending Tulisaari dyke (15SD_083). a. Intergranular texture of plagioclase (Plag), oxides and secondary hornblende (Hbl), plane polarized light. b. As in a, crossed nicols. c. Backscattered electron image showing baddeleyite hosted adjacent to hornblende (Hbl), plagioclase (Plag) and quartz (Qtz). d. Baddeleyite, Bdy-1 as in c, with zircon (Zrc) overgrowth. e. Anhedronal zircon grain Bdy-2 as in c, within quartz. f. Baddeleyite Bdy-2 as in c, with zircon overgrowth. g. Picked baddeleyite grains.

Figure 3-9. Petrographic details of the NW-trending *ca.* 1.98 Ga Kivikevätti (15SD_085) and Eno (15SD_088) dykes. a. Subophitic texture of the Kivikevätti dyke with plagioclase (Plag), clinopyroxene with simple twins (Cpx), biotite (Bt), and oxides with minor chlorite alteration (Chl) on mafic minerals, plane polarized light. b. As in b, crossed nichols. Showing myrmekitic texture. c. Ophitic texture of the Eno dyke with secondary amphibole, plagioclase and accessory apatite (Ap). d. As in c, crossed nichols. e. Single grain baddeleyite (left) and zircon (right) fractions analyzed with U-Pb TIMS for the Kivikevätti dyke. f. Single grain zircon fractions analyzed with CA-TIMS for the Eno dyke.

3.5.3 Geochronology

3.5.3.1 Kakikangas dyke, 15SD_075

Mineral separation and picking yielded a small supply of ~50 transparent, honey-brown baddeleyite crystals and fragments from this medium-grained diabase. Subhedral crystals and fragments are generally lath-shaped and are on average 50 μm long and rarely 75 μm long (**Figure 3-6**). Four multigrain fractions involving four, nine, ten and eight grains/fraction were analyzed (**Table 3-2**) resulting in 1.5-3.1% discordant data (**Figure 3-10**). Discordance may be attributed to either mixing of two components (e.g., baddeleyite and zircon overgrowth) or variable lead loss (Heaman and LeCheminant, 1993; Rioux *et al.*, 2010; Söderlund *et al.*, 2013). Of the four discordant fractions, Bdy1 is the least discordant (1.5%) and is collinear with the more discordant fractions Bdy2-Bdy4, which overlap in error. A regression of these four fractions provides an upper intercept age of 2329.2 ± 8.2 Ma and a lower intercept of -40 ± 830 Ma. The upper intercept age of 2329.2 ± 8.2 Ma is considered a best estimate for the crystallization age of the 240° -trending Kakikangas dyke.

3.5.3.2 Tilsanvaara dyke, 15SD_076

The coarse-grained 280° -trending Tilsanvaara dyke yielded abundant prismatic zircon crystals and very few dark brown baddeleyite fragments (**Figure 3-6**). Four single grain baddeleyite fractions yielded 0.9-2.7% discordant and -2.5% reversely discordant results (**Table 3-2**; **Figure 3-10**). Of the fractions that are discordant, Bdy1-3 are collinear and a free regression calculated using these data provide an upper intercept age of 2299 ± 20 Ma

and a lower intercept of 936 ± 730 Ma. A regression through the least discordant fraction, Bdy1, anchored at 0 Ma yields an upper intercept of 2292 ± 8.8 Ma. This is considered a minimum age and best estimate for the crystallization age of the 280° -trending Tilsanvaara dyke. Two single grain zircon fractions provide concordant or near-concordant results with a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean age of 1972.8 ± 11.2 Ma (**Figure 3-10**). This result is considered to represent the timing of the intrusion of the younger cross-cutting 295° -trending Tilsanvaara dyke.

3.5.3.3 Tulisaari dyke, 15SD_083

Abundant baddeleyite crystals and fragments were separated from the medium-grained, 295° -trending Tulisaari dyke from the Iisalmi block. Baddeleyite crystals and fragments were clear to light brown, some of which having zircon overgrowth visible under the binocular microscope (**Figure 3-8**). Three baddeleyite grains without visible zircon overgrowth were selected for TIMS analyses and provided 1.6-6.6% discordant results (**Table 3-2; Figure 3-10c**), following similar Pb-loss patterns to the results of Hölttä *et al.* (2000a). A regression calculated through the least discordant fraction Bdy1 and the most precision fraction, Bdy4, provides an upper intercept age of $2336 + 31/-22$ Ma with a lower intercept at 517 ± 100 Ma. Calculation of a regression through the least discordant fraction, Bdy1 anchored at 0 Ma provides an upper intercept of 2320 ± 13 Ma and this is considered to represent a minimum crystallization age of the Tulisaari dyke.

3.5.3.4 Viianki, 15SD_084

Mineral separation of the coarsest grained portions of the 080°-trending Viianki dyke yielded many small baddeleyite crystals. Ranging in length from 20-60 μm , the baddeleyite grains were generally honey-brown and clear, platy crystals and fragments (**Figure 3-4**). Five multiple grain fractions (between seven to twelve grains) yielded variably discordant results (**Table 3-2; Figure 3-10d**). Fractions Bdy1-Bdy4, between 1.1-2.6% discordant, are collinear and a regression through these fractions provides an upper intercept of 2404 ± 29 Ma and a lower intercept of 126 ± 3000 Ma. A regression through the least discordant fraction, Bdy1 (1.1%) anchored at 0 Ma provides an upper intercept of 2400.7 ± 1.8 Ma. Fraction Bdy5 plots far to the left of the regression lines calculated using the less discordant fractions and is collinear with a multiple grain analysis on baddeleyite reported by Huhma *et al.* (2018), who reported a minimum age of the dyke at *ca.* 2395 Ma. The upper intercept age of 2404 ± 29 Ma is considered to represent the crystallization age of the Viianki dyke.

3.5.3.5 Kivikevätti, 15SD_085

The coarse-grained sample collected from Kivikevätti includes abundant zircon and some baddeleyite crystals and fragments (**Figure 3-9**). Zircon is transparent and occurs as thin platy crystals to 200 μm long, as well as smaller prismatic crystals with pyramidal terminations. The few baddeleyite crystals picked vary substantially in quality (zircon overgrowth), but some crystals were translucent and light brown with lengths up to 150 μm . Two single grain fractions of chemically abraded zircon and three single baddeleyite grains provided discordant results (**Table 3-2; Figure 3-10e**). The two single zircon

fractions are reversely discordant at -6.2 and -3.3%. The three single baddeleyite grains are discordant between 1.2 and 2.9%. Together, the five fractions are collinear and when calculated in a regression provide a lower intercept of -38 ± 610 Ma and an upper intercept of 1975.6 ± 9.7 Ma.

3.5.3.6 Eno, 15SD_088

Approximately 75 zircon crystals and fragments were picked from the coarse-grained Eno dyke sample (**Figure 3-9**). The elongate crystals measure up to 150 μm in length and 75 μm in width, and occur with poorly formed terminations and are heavily fractured. Following chemical abrasion, three single grain zircon fractions yielded nearly concordant results ranging from 0.3-1.9% discordant. A weighted mean of the $^{207}\text{Pb}/^{206}\text{Pb}$ ages from the three fractions provides a date at 1940 ± 3 Ma (**Figure 3-10f**). A weighted mean was calculated because the error ellipses for each fraction are not collinear and do not overlap (**Figure 3-10f**). If a regression line is calculated through the least discordant zircon fraction Zr1 and anchored at 1800 ± 100 Ma, the upper intercept is *ca.* 1980 Ma.

Table 3-2. U-Pb isotopic data for mafic dykes in Western Karelia.

Fraction (analysis number)	Weight (µg)	U (ppm)	Th U	Pbtot (pg)	PbC (pg)	²⁰⁶ Pb Meas.	²⁰⁷ Pb Meas.	2 σ	²⁰⁶ Pb ²³⁸ U	2 σ	Error Corr.	²⁰⁷ Pb ²⁰⁶ Pb	2 σ	²⁰⁶ Pb ²³⁸ U	2 σ	²⁰⁷ Pb ²³⁵ U	2 σ	²⁰⁷ Pb Age (Ma)	2 σ	Disc. (%)
<i>Kakikangas, 15SD_075</i>																				
Bdy1, 4 (sk38p241)	0.10	472	0.485	22.3	0.7	1889	8.7807	0.037	0.42868	0.00143	0.855	0.148557	0.000326	2299.8	6.5	2315.5	3.8	2329.4	3.8	1.5
Bdy2, 9 (sk38p184)	0.11	2420	0.107	13.6	0.8	1046	8.6682	0.052	0.42339	0.00172	0.803	0.148487	0.000544	2275.8	7.8	2303.7	5.5	2328.6	6.3	2.7
Bdy3, 10 (sk38p185)	0.09	283	0.729	13.9	0.6	1357	8.6710	0.049	0.42250	0.00194	0.850	0.148847	0.000448	2271.8	8.8	2304.0	5.2	2332.7	5.2	3.1
Bdy4, 8 (sk38p172)	0.10	593	0.056	25.0	2.3	729	8.5980	0.059	0.42101	0.00142	0.705	0.148118	0.000748	2265.0	6.5	2296.3	6.2	2324.3	8.7	3.0
<i>Tilsanvaara, 15SD_076</i>																				
Bdy1 (sk39p157)	0.05	326	0.146	7.0	0.5	881	8.4883	0.078	0.42334	0.00302	0.826	0.145421	0.000761	2275.6	13.7	2284.7	8.4	2292.8	9.0	0.9
Bdy2 (sk39p156)	0.05	289	0.239	6.2	0.4	921	8.2345	0.078	0.41362	0.00332	0.888	0.144388	0.000636	2231.5	15.1	2257.1	8.6	2280.5	7.6	2.5
Bdy3 (sk39p180)	0.10	174	0.346	7.6	1.0	490	8.0064	0.091	0.40785	0.00266	0.737	0.142377	0.001133	2205.1	12.2	2231.7	10.3	2256.3	13.8	2.7
Bdy4 (sk39p181)	0.10	171	0.182	7.2	1.1	426	7.8423	0.183	0.41492	0.00376	0.629	0.137081	0.002610	2237.4	17.1	2213.1	21.0	2190.6	33.0	-2.5
Zrc1 (sk38p196)	0.10	204	0.318	7.6	0.6	799	5.9819	0.058	0.35770	0.00222	0.716	0.121289	0.000830	1971.3	10.5	1973.2	8.5	1975.3	12.2	0.2
Zrc2 (sk38p197)	0.10	127	0.739	5.2	0.9	351	5.9214	0.124	0.35622	0.00343	0.669	0.120559	0.001943	1964.2	16.3	1964.4	18.1	1964.5	28.8	0.0
<i>Tulisaari, 15SD_083</i>																				
Bdy1 (sk39p158)	0.10	125	0.074	5.4	0.8	474	8.6796	0.114	0.42611	0.00398	0.812	0.147733	0.001144	2288.1	18.0	2304.9	11.9	2319.9	13.3	1.6
Bdy2, 3 (sk39p160)	0.50	172	0.087	36.1	1.3	1795	8.3792	0.032	0.41734	0.00107	0.779	0.145617	0.000357	2248.4	4.9	2272.9	3.5	2295.1	4.2	2.4
Bdy3 (sk39p184)	0.19	2051	0.137	15.9	1.5	686	7.9036	0.062	0.40109	0.00152	0.706	0.142916	0.000833	2174.1	7.0	2220.1	7.1	2262.8	10.1	4.6
Bdy4 (sk39p159)	0.50	129	0.094	25.5	1.2	1421	7.7261	0.035	0.39252	0.00118	0.744	0.142757	0.000437	2134.5	5.5	2199.6	4.1	2260.9	5.3	6.6

Table 3-2 continues on the next page

Notes: Fractions include single grain chemically abraded zircon (Zrc), or single or multiple grain baddeleyite (Bdy). The number of grains in a multiple grain fraction is indicated after the fraction number. Th/U is calculated from radiogenic ²⁰⁸Pb/²⁰⁶Pb ratio and ²⁰⁷Pb/²⁰⁶Pb age assuming concordance. Pbtot is the amount of Pb in picograms excluding blank, Pbcom is total amount of common Pb in picograms; assigned to the isotopic composition of laboratory blank. Meas. – Measured. ²⁰⁶Pb/²⁰⁴Pb corrected for fractionation and common Pb in the spike. 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. Uranium decay constants are those from Jaffey *et al.* (1971).

Table 3-2. Continued.

Fraction (analysis number)	Weight (μg)	U (ppm)	Th U	Pbtot (pg)	Pbcom (pg)	²⁰⁶ Pb ²⁰⁴ Pb Meas. 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. (%)
<i>Viianki, 15SD_084</i>																				
Bdy1, 7 (sk38p179)	1.00	145.9	0.129	66.7	0.8	5503	9.5312	0.028	0.44628	0.00112	0.929	0.154893	0.000169	2378.7	5.0	2390.6	2.7	2400.7	1.9	1.1
Bdy2, 10 (sk38p173)	0.30	1300	0.112	176.3	5.3	2115	9.4749	0.045	0.44300	0.00185	0.913	0.155121	0.000304	2364.1	8.3	2385.1	4.4	2403.2	3.3	1.9
Bdy3, 12 (sk38p174)	0.20	1862	0.149	169.5	1.1	9672	9.4767	0.025	0.44227	0.00100	0.915	0.155406	0.000163	2360.8	4.4	2385.3	2.4	2406.3	1.8	2.3
Bdy4, 7 (sk38p178)	0.10	1	0.131	58.1	1.7	2136	9.3788	0.031	0.43925	0.00100	0.829	0.154858	0.000297	2347.3	4.5	2375.8	3.1	2400.3	3.3	2.6
Bdy5, 7 (sk38p180)	0.10	2057	0.024	88.3	4.5	1280	9.0280	0.041	0.43120	0.00107	0.732	0.151849	0.000486	2311.1	4.8	2340.8	4.1	2366.9	5.5	2.8
<i>Kivikevätti, 15SD_085</i>																				
Zrc1 (sk38p186)	53	6	0.983	138.8	0.4	18276	6.3403	0.016	0.37986	0.00097	0.798	0.121054	0.000197	2075.7	4.6	2024.0	2.2	1971.8	2.9	-6.2
Zrc2 (sk38p187)	1.9	67	0.875	55.6	0.3	9098	6.2173	0.017	0.37107	0.00087	0.926	0.121517	0.000127	2034.5	4.1	2006.9	2.4	1978.6	1.9	-3.3
Bdy1, 4 (sk39p164)	4.5	25	0.068	39.4	2.5	1058	5.9217	0.034	0.35409	0.00088	0.669	0.121290	0.000549	1954.1	4.2	1964.4	5.0	1975.3	8.1	1.2
Bdy2, 5 (sk39p165)	0.5	403	0.027	68.4	4.9	937	5.8347	0.046	0.35028	0.00173	0.732	0.120809	0.000652	1935.9	8.3	1951.6	6.8	1968.2	9.6	1.9
Bdy3 (sk39p186)	61	1	0.041	24.5	2.0	815	5.8806	0.042	0.34998	0.00100	0.681	0.121864	0.000679	1934.5	4.8	1958.4	6.2	1983.7	9.9	2.9
<i>Eno, 15SD_088</i>																				
Zrc1 (sk38p183a)	7	7	0.372	19.1	0.6	1975	5.7285	0.00276	0.3497	1.12E-03	0.673	0.118807	0.000424	1933.2	5.4	1935.7	4.2	1938.4	6.4	0.3
Zrc2 (sk38p181)	6	12	1.401	34.0	0.7	2296	5.7110	0.00234	0.3465	1.08E-03	0.760	0.119534	0.000318	1917.9	5.2	1933.0	3.5	1949.3	4.8	1.9
Zrc3 (sk38p182)	1.1	6	0.994	29.6	0.7	2310	5.6548	0.00221	0.3458	1.02E-03	0.791	0.118597	0.000284	1914.6	4.9	1924.5	3.4	1935.2	4.3	1.2

Notes: Fractions include single grain chemically abraded zircon (Zrc), or single or multiple grain baddeleyite (Bdy). The number of grains in a multiple grain fraction is indicated after the fraction number. Th/U is calculated from radiogenic ²⁰⁸Pb/²⁰⁶Pb ratio and ²⁰⁷Pb/²⁰⁶Pb age assuming concordance. Pbtot is the amount of Pb in picograms excluding blank, Pbcom is total amount of common Pb in picograms; assigned to the isotopic composition of laboratory blank. Meas. – Measured. ²⁰⁶Pb/²⁰⁴Pb corrected for fractionation and common Pb in the spike. 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. Uranium decay constants are those from Jaffey *et al.* (1971).

Figure 3-10. *On the previous page.* U-Pb concordia diagrams for mafic dykes in Western Karelia. a. Kakikangas dyke sample 15SD_075. The discordia patterns on four multiple grain baddeleyite fractions are collinear and yield an upper intercept age of 2339.2 ± 8.2 Ma. b. Tilsanvaara dyke sample 15SD_076. The complex discordia patterns on the single grain baddeleyite fractions (Bdy1-Bdy3) provide an upper intercept age of 2299 ± 20 Ma (dashed regression line). A minimum age of 2292 ± 8.8 Ma is calculated from the least discordant datum (Bdy1) with the lower intercept anchored at 0 Ma (solid regression line). c. Tulisaari dyke sample 15SD_083. Single and multiple grain baddeleyite fractions yield scattered and discordant results. An upper intercept date determined using the least discordant Bdy1 fraction and more precise Bdy4 single grain fractions (dashed regression line) is $2336 + 31 / - 22$ Ma. A minimum age of 2320 ± 13 Ma is determined from the least discordant single grain fraction Bdy1 (solid regression line). Multiple grain air-abraded zircon fractions (dark grey ellipses) of Hölttä *et al.* (2000b) provide an upper intercept age of 2295.1 ± 4.6 Ma (grey stippled line) onto which fractions Bdy3 and Bdy4 of this study plot. d. Viianki dyke sample 15SD_084. The four least discordant multiple grain baddeleyite fractions yield an upper intercept age of 2404 ± 29 Ma. Regression of the least discordant baddeleyite fraction Bdy1 anchored at 0 Ma provides a minimum age of 2400.7 ± 1.8 Ma. The multiple grain fraction from Huhma *et al.* (2018) plots to the left of these regression lines and is collinear with the multiple grain baddeleyite fraction Bdy5 of this study, omitted from the age calculations. e. Kivikevätti dyke sample 15SD_085. Two reverse discordant zircon analyses and three discordant baddeleyite grains define an upper intercept of 1975.6 ± 9.7 Ma. These data are collinear with the four baddeleyite fractions from Huhma *et al.* (2018), reporting an age of 1980 ± 4 Ma for this dyke. f. Eno dyke, 15SD_088. Three chemically abraded single grain zircon fractions are near concordant but are not collinear. A weighted mean of their $^{207}\text{Pb}/^{206}\text{Pb}$ ages provides a minimum age of 1940 ± 3 Ma. A regression line through the least discordant fraction Zrc1 anchored at 1800 ± 100 Ma (Svecofennian deformation) provides an upper intercept *ca.* 1980 Ma.

3.5.4 Geochemistry

Whole-rock major and trace element geochemistry is presented for six mafic dykes, including the Kakikangas, Tulisaari, WNW- and NW-trending Tilsanvaara dykes, Kivikevätti and Eno dykes dated in this study (for geochemical data on the Viianki dyke (see **Chapter 2**), as well as a new dyke, 16SD_001, that cross-cuts the Siilinjärvi carbonatite open pit (**Table 3-3**). Additional geochemical analyses are presented on four *ca.* 2.32 Ga dykes that have been dated in previous studies: Törinkuru (AD13) as documented by Salminen *et al.* (2014), Taivalkoski AD10 of Vuollo and Huhma (2005),

Humppi and Karkuvaara after Huhma *et al.* (2018).

On total alkali-silica and Nb/Y versus Zr/Ti classification diagrams, the dykes plot in the basalt to basaltic andesite fields (**Figure 3-11a and b**). The dykes have Ti/V ratios between 20 and 50, which are consistent with mid-ocean ridge basalt (MORB) on a V versus Ti (ppm)/1000 diagram (**Figure 3-11c.**; Shervais, 1982). Nearly all dykes have ΔNb values ($\Delta\text{Nb} = 1.74 + \log(\text{Nb}/\text{Y}) - 1.92\log(\text{Zr}/\text{Y})$, as calculated by Fitton *et al.*, 1997) and for this reason plot below or at the margin of the Iceland plume array on a Zr/Y versus Nb/Y diagram (**Figure 3-11d**).

Sample 15SD_072A (AD13 dyke of Salminen *et al.*, 2014) and Eno 15SD_088 are exceptions as both are classified as andesite on the total alkali-silica and Nb/Y versus Zr/Ti diagrams, and have distinctly higher Ti/V ratios. These samples were taken from coarse-grained pegmatitic sections found near the centres of the dykes. Note that sample 15SD_072B, which is representative of the average medium-grained diabase found at the AD13 dyke, is classified as a basaltic andesite and plots within the cluster of other samples belonging to this study.

Multi-element patterns for the studied dykes (**Figure 3-12a and b**) are uniform and include flat to shallow heavy rare earth element (HREE) slopes corresponding to Tb/Yb ratios between 1.0 and 1.6 and moderate light rare earth element (LREE) slopes corresponding to La/Sm ratios between 1.4 and 2.1. Negative Nb anomalies occur for all samples, consistent with the samples plotting below the Iceland plume array on the Zr/Y versus Nb/Y diagram (**Figure 3-11d**).

Table 3-3. Whole-rock major and trace element geochemistry for *ca.* 2.40 Ga, 2.32 Ga, and 1.98 Ga mafic dykes, Western Karelia.

Sample	Locality	Notes	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	CaO	MgO	Na ₂ O	K ₂ O	Cr ₂ O ₃	TiO ₂	MnO	P ₂ O ₅	SrO	BaO	LOI	Total
		<i>Detection limit</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>
15SD_072A	Törinkuru AD13	Coarse grained centre	57.80	11.70	12.05	8.61	1.50	4.09	0.47	<0.01	1.35	0.13	0.67	0.01	0.01	1.38	99.77
15SD_072B	Törinkuru AD13	Average diabase texture	51.70	13.80	15.25	9.99	4.77	2.66	0.57	0.01	1.16	0.24	0.17	0.02	0.02	0.34	100.70
15SD_074	Taivalkoski AD10		52.30	12.30	16.95	8.15	3.97	2.63	0.93	0.01	2.08	0.22	0.20	0.02	0.04	-0.03	99.77
15SD_075	Kakikangas		52.00	12.60	16.90	8.38	4.20	2.61	0.81	0.01	2.08	0.22	0.20	0.02	0.03	-0.41	99.65
15SD_076A	WNW Tilsanvaara	1 m from contact	50.50	15.20	14.00	10.25	5.52	2.45	0.35	0.02	1.10	0.20	0.09	0.02	0.01	-0.30	99.41
15SD_077	NW Tilsanvaara		50.6	13.6	14.35	9.82	6.01	2.4	0.71	0.01	1.87	0.22	0.2	0.02	0.03	0.64	100.48
15SD_079	Karkuvaara		51.90	15.50	10.15	11.15	7.52	2.30	0.20	0.04	0.80	0.18	0.06	0.02	0.01	-0.01	99.82
15SD_080	Humppi		50.40	13.85	16.95	9.12	4.24	2.42	0.53	0.01	2.15	0.21	0.22	0.02	0.02	-0.29	99.85
15SD_083	Tulisaari		50.00	13.65	15.80	9.00	4.29	2.46	0.80	0.01	2.10	0.21	0.26	0.02	0.03	0.11	98.74
15SD_085B	Kivikevätti	1 m from chilled margin	47.6	12.65	17.35	8.67	4.82	2.38	1.02	0.01	2.84	0.26	0.44	0.02	0.05	-0.07	98.04
15SD_085C	Kivikevätti	Coarse grained centre	50.00	12.65	16.55	8.66	4.48	2.6	1.2	0.01	2.41	0.25	0.56	0.02	0.06	-0.01	99.44
15SD_088	Eno		58.1	11.95	11.70	8.52	3.1	4.17	0.23	<0.01	1.67	0.17	0.71	0.03	<0.01	0.29	100.64
16SD_001	Siilinjärvi dyke		49.30	14.70	16.05	8.60	2.63	3.12	0.58	<0.01	2.41	0.22	0.31	0.03	0.02	0.20	98.17
10-LT-05 Standard – Batch A			48.70	17.9	9.69	9.61	6.5	3.52	0.99	0.03	1.4	0.14	0.37	0.11	0.06	0.59	99.63
10-LT-05 Standard – Batch B			49.60	18.8	9.35	9.47	6.52	3.63	1.01	0.03	1.4	0.14	0.38	0.11	0.06	0.82	101.3
10-LT-05 Standard - Average			48.88	18.18	9.54	9.52	6.61	3.60	1.02	0.03	1.44	0.14	0.38	0.11	0.06	0.75	100.27
2σ			1.16	0.50	0.29	0.32	0.21	0.20	0.03	0.01	0.06	0.00	0.02	0.00	0.00	0.34	1.69

Table 3-3 continues on the next page.

Notes: Major elements in wt. % and trace elements in ppm. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B. All samples belong to batch A with the exception of 16SD_001, which belongs to batch B. The average composition of 10-LT-05 and 2σ values of 25 analyses spanning 6 years are also provided.

Table 3-3. Continued.

Sample	Locality	Ba	Ce	Cr	Cs	Dy	Er	Eu	Ga	Gd	Hf	Ho	La	Lu	Nb	Nd	Pr	Rb	Sm	Sn	Sr	Ta
	<i>Detection limit</i>	<i>0.5</i>	<i>0.5</i>	<i>10</i>	<i>0.01</i>	<i>0.05</i>	<i>0.03</i>	<i>0.03</i>	<i>0.1</i>	<i>0.05</i>	<i>0.2</i>	<i>0.01</i>	<i>0.5</i>	<i>0.01</i>	<i>0.2</i>	<i>0.1</i>	<i>0.03</i>	<i>0.2</i>	<i>0.03</i>	<i>1</i>	<i>0.1</i>	<i>0.1</i>
15SD_072A	Törinkuru AD13	91.5	96.4	<10	0.01	13.90	8.82	3.42	22.8	15.05	11.1	3.22	45.5	1.21	28.1	57.5	13.15	3.2	13.95	3	146.0	1.8
15SD_072B	Törinkuru AD13	182.0	33.0	60	0.23	5.82	3.61	1.50	20.4	5.80	4.0	1.28	14.8	0.50	8.8	19.1	4.34	15.7	5.25	1	175.0	0.5
15SD_074	Taivalkoski AD10	329.0	42.2	40	0.31	5.91	3.55	1.86	21.9	6.51	4.6	1.30	21.2	0.46	9.4	24.7	5.58	24.4	5.82	2	193.5	0.6
15SD_075	Kakikangas	267.0	38.2	60	0.22	6.12	3.83	1.95	23.0	6.40	4.7	1.41	18.7	0.48	9.7	22.8	5.08	21.3	5.91	2	176.5	0.6
15SD_076A	WNW Tilsanvaara	119.0	17.0	130	0.12	3.22	1.90	1.02	20.0	3.35	2.0	0.71	7.9	0.25	4.5	10.9	2.33	8.1	2.75	1	200.0	0.3
15SD_077	NW Tilsanvaara	265	35.7	60	0.61	5.26	3.29	1.67	21.1	5.68	3.7	1.12	17.2	0.41	11.7	20.6	4.92	21.1	5.2	1	196.5	0.8
15SD_079	Karkuvaara	64.3	9.6	270	0.39	2.60	1.66	0.87	19.2	2.64	1.4	0.62	4.3	0.23	2.7	7.1	1.44	7.3	1.93	1	137.5	0.2
15SD_080	Humppi	169.0	33.3	70	0.58	6.45	4.04	1.92	24.2	6.55	4.5	1.46	15.6	0.58	11.6	20.9	4.59	21.1	5.60	2	137.0	0.6
15SD_083	Tulisaari	241.0	41.7	70	0.53	8.01	4.92	2.01	22.3	7.67	5.3	1.78	19.8	0.66	11.9	25.8	5.59	26.4	6.27	2	172.0	0.7
15SD_085B	Kivikevätti	424	61.5	40	0.4	9.16	5.07	2.46	22.5	9.14	6.5	1.98	27.7	0.74	23.9	34.9	8	31	8.53	2	156	1.4
15SD_085C	Kivikevätti	526	77.7	60	0.39	11	6.56	2.95	22.4	10.95	7.7	2.31	37.7	0.92	26.1	42.5	10.3	39	10.65	2	164	1.6
15SD_088	Eno	33	131	<10	0.07	21.4	12.2	4.04	22.6	20.6	15	4.4	61.9	1.65	28.5	74.2	17.65	2.2	18.8	4	279	1.7
16SD_001	Siilinjärvi dyke	149.0	36.4	10	0.18	9.03	5.52	2.37	23.6	8.85	5.6	2.02	14.9	0.78	11.4	25.8	5.22	10.3	7.71	1	290.0	0.7
10-LT-05 Standard – Batch A		539.0	45.8	180	0.21	3.70	2.03	1.54	19.5	4.50	3.5	0.67	22.0	0.27	10.8	24.7	6.08	12.0	5.20	1	904.0	0.6
10-LT-05 Standard – Batch B		556.0	47.5	190	0.21	3.77	2.11	1.66	20.1	4.75	3.5	0.72	21.9	0.23	11.5	26.2	6.01	12.2	5.63	1	960.0	0.7
10-LT-05 Standard - Average		538.3	48.0	184	0.23	3.63	1.97	1.59	19.9	4.45	3.6	0.73	22.4	0.25	10.9	25.6	6.21	12.5	5.15	1.2	938.2	0.6
2σ		36.6	3.9	23	0.06	0.29	0.15	0.16	1.4	0.44	0.3	0.07	2.0	0.02	0.9	2.0	0.66	0.8	0.51	0.8	38.5	0.1

Table 3-3 continues on the next page.

Notes: Major elements in wt. % and trace elements in ppm. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B. All samples belong to batch A with the exception of 16SD_001, which belongs to batch B. The average composition of 10-LT-05 and 2σ values of 25 analyses spanning 6 years are also provided.

Table 3-3. Continued.

Sample	Locality	Tb	Th	Tm	U	V	W	Y	Yb	Zr	Ag	As	Cd	Co	Cu	Li	Mo	Ni	Pb	Sc	Tl	Zn
	<i>Detection limit</i>	<i>0.01</i>	<i>0.05</i>	<i>0.01</i>	<i>0.05</i>	<i>5</i>	<i>1</i>	<i>0.5</i>	<i>0.03</i>	<i>2</i>	<i>0.5</i>	<i>5</i>	<i>0.5</i>	<i>1</i>	<i>1</i>	<i>10</i>	<i>1</i>	<i>1</i>	<i>2</i>	<i>1</i>	<i>10</i>	<i>2</i>
15SD_072A	Törinkuru AD13	2.45	7.05	1.22	0.98	22	<1	78.2	7.97	438	<0.5	<5	<0.5	33	66	10	<1	10	14	21	<10	103
15SD_072B	Törinkuru AD13	0.91	2.49	0.50	0.44	219	<1	33.0	3.32	141	<0.5	<5	<0.5	46	201	10	1	46	5	35	10	142
15SD_074	Taivalkoski AD10	1.00	4.27	0.53	0.44	375	1	31.8	3.26	178	<0.5	<5	0.6	53	267	10	<1	53	6	33	<10	157
15SD_075	Kakikangas	1.01	3.49	0.57	0.52	390	1	33.4	3.42	172	<0.5	<5	0.5	53	272	10	<1	58	8	34	<10	156
15SD_076A	WNW Tilsanvaara	0.54	1.07	0.28	0.21	230	<1	17.5	1.84	72	<0.5	<5	0.6	51	179	10	<1	86	5	32	<10	111
15SD_077	NW Tilsanvaara	0.94	2.47	0.5	0.64	393	<1	28.5	2.95	143	<0.5	<5	0.6	48	78	10	<1	46	14	40	<10	129
15SD_079	Karkuvaara	0.43	0.54	0.24	0.15	233	<1	14.6	1.51	51	<0.5	<5	<0.5	41	144	10	<1	117	7	36	<10	84
15SD_080	Humppi	1.04	2.80	0.62	0.70	374	1	36.2	3.79	173	<0.5	<5	0.5	49	250	10	<1	59	4	33	<10	145
15SD_083	Tulisaari	1.28	3.42	0.73	0.57	322	<1	43.5	4.56	203	<0.5	<5	0.5	43	371	20	<1	54	6	34	<10	144
15SD_085B	Kivikevätti	1.48	3.54	0.79	1	527	1	50.4	5.13	262	<0.5	<5	1	51	58	20	2	34	6	46	<10	168
15SD_085C	Kivikevätti	1.83	4.97	0.96	1.2	396	1	57.2	6.05	305	<0.5	<5	<0.5	39	36	10	<1	29	7	41	<10	151
15SD_088	Eno	3.46	10.2	1.75	1.63	38	<1	108.5	10.65	573	<0.5	5	<0.5	26	9	10	<1	13	7	21	<10	139
16SD_001	Siilinjärvi dyke	1.53	1.82	0.80	0.43	338	<1	51.6	5.11	210	<0.5	<5	0.5	36	417	10	1	25	4	31	10	137
10-LT-05 Standard – Batch A		0.66	2.01	0.27	0.56	154	1	17.4	1.64	137	<0.5	<5	<0.5	39	41	10	<1	116	<2	17	<10	91
10-LT-05 Standard – Batch B		0.67	1.88	0.25	0.59	165	1	18.9	1.72	152	<0.5	<5	<0.5	43	45	10	1	124	<2	18	<10	100
10-LT-05 Standard		0.66	1.99	0.27	0.59	163.81	1	18.64	1.66	150.35	NA	7	1	39	42	10	1	115	4	17	10	92
2σ		0.06	0.22	0.04	0.1	21.7	0	1.48	0.21	14.05	NA	1	0	4	4	0	1	9	3	2	0	7

Notes: Major elements in wt. % and trace elements in ppm. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B. All samples belong to batch A with the exception of 16SD_001, which belongs to batch B. The average composition of 10-LT-05 and 2σ values of 25 analyses spanning 6 years are also provided.

Figure 3-11. Whole-rock major and trace element classification for dated dykes in this study and contemporaneous units. a. Total alkali silica diagram (Le Bas *et al.*, 1986). b. Zr/Ti versus Nb/Y (Pearce, 1996). c. V versus Ti/1000 (Shervais, 1982). d. Nb/Y versus Zr/Y (Fitton *et al.*, 1997; Luttinen, 2018). Data presented is a compilation from this study, Vuollo and Fedotov (2005); Stepanova *et al.* (2015, 2017), and **Chapter 2**.

3.6 Discussion

3.6.1 Geochronology and relation to regional dyke swarms

3.6.1.1 Uolevinlehto swarm, *ca.* 2.40 Ga

Mafic dykes with ages *ca.* 2.40 Ga have been reported in the Pudasjärvi, Taivalkoski and Russian blocks of Western Karelia (**Chapter 2; Figure 3-1**; Vuollo and Huhma, 2005; Stepanova *et al.*, 2017; Huhma *et al.*, 2018). In the Kuhmo and Taivalkoski block, these dykes have EW trends and span a minimum latitudinal range of about 90 km. The minimum age of the Viianki dyke at 2400.7 ± 1.8 Ma, located in the Kuhmo block, is consistent with this regional pattern. The contemporaneous *ca.* 2400 ± 12 Ma Uolevinlehto dyke located in the Pudasjärvi block, however, has a distinct NNW trend. As discussed in **Chapter 2**, the approximately $\sim 30^\circ$ difference in dyke trends in the Pudasjärvi block versus those in the Kuhmo and Taivalkoski block are attributed to Svecofennian deformation. This deformation caused the Pudasjärvi block to rotate $\sim 30^\circ$ clockwise about its vertical axis (**Chapter 2**).

3.6.1.2 Taivalkoski-Kuito swarm, *ca.* 2.32 Ga

Mafic dykes with ages *ca.* 2.32 Ga are widespread in Western Karelia having been identified in the Taivalkoski, Kuhmo and Iisalmi blocks (**Figure 3-1**; Mertanen *et al.*, 1999; Hölttä *et al.*, 2000b; Vuollo and Huhma, 2005; Salminen *et al.*, 2014). With the exception of three localities, dykes with *ca.* 2.32 Ga ages have NW trends.

In the Iisalmi block, the NW-trending Tulisaari dyke has a minimum age of

2320±13 Ma, results that are roughly 25 Ma older than those previously reported (Hölttä *et al.*, 2000b).

The WNW trend of the older dyke at the Tilsanvaara locality is included in this group of dykes because analyses of three single grain baddeleyite fractions provide a minimum age of 2292±8.8 Ma, which is within error of the general population of *ca.* 2.32 Ga dykes. Given that the growth of baddeleyite is limited to magmatic processes, these analyses are considered to represent the minimum crystallization age of the dyke. The significance of the younger zircon crystals that have also been analyzed from this dyke are discussed below in the section dedicated to *ca.* 1.98 Ga magmatism.

U-Pb geochronology has also confirmed that the NE-trending dykes at the boundary between the Kuhmo and Taivalkoski blocks are coeval. The single grain baddeleyite fractions from the Kakikangas locality provide an upper intercept age of 2329±8.2Ma, which is consistent with Sm-Nd dating on the subparallel Kallioniemi dyke at 2352±62 Ma (Vuollo and Huhma, 2005; Huhma *et al.*, 2018) and within error of many 2.32 Ga dykes across Karelia.

3.6.1.3 Paukkajanvaara swarm and Pechenga-Onega LIP, *ca.* 1.98 Ga

The U-Pb geochronology on the NW-trending *ca.* 1.98 Ga Kivikevätti and Eno dykes, as well as the zircons picked from the Tilsanvaara dyke, produced complicated results that require careful evaluation and discussion (**Figure 3-10b, e, and f**).

Baddeleyite analyses from the Kivikevätti dyke agree with previous age determinations by Huhma *et al.* (2018) as the three single grain baddeleyite fractions overlap with those of Huhma *et al.* (2018). A free regression of the three baddeleyite

fractions yields an upper intercept age of 1972 ± 30 Ma with a lower intercept of -549 ± 5200 Ma (not shown).

Two chemically abraded single zircon analyses for the Kivikevätti dyke yield reversely discordant results (**Figure 3-10e**). Reverse discordance occurs when U-Pb analyses plot above the concordia curve due to excess radiogenic ^{207}Pb and ^{206}Pb relative to ^{235}U and ^{238}U , respectively, such that the $^{206}\text{Pb}/^{238}\text{U}$ age is older than the $^{207}\text{Pb}/^{235}\text{U}$ age (Williams *et al.*, 1984; Mattinson *et al.*, 1996). Mattinson *et al.* (1996) suggests that reverse discordance is the result of natural processes related to the redistribution of radiogenic lead either internally by diffusion or movement (termed implantation) associated with alpha-recoil during U decay, or by the removal of U by external mechanisms (e.g., hydrothermal or metamorphic fluids, or chemical weathering). In this case, the cause of reverse discordance by an external source can be discounted because the Kivikevätti dyke is well preserved including fresh primary mafic magmatic minerals (**Figure 3-9**). Internal processes related to reverse discordance in zircon are especially important to consider when zircons are treated by chemical abrasion. The goal of chemical abrasion is to remove damaged zones of the zircon crystal that have lost lead (leading to discordance). The most damaged zones, however, are typically those with higher U concentrations which have therefore experienced more metamictization (damage to the crystal structure due to radiogenic decay) compared to low U zones. Given that chemical abrasion selects for well-preserved low U zones, reverse discordance in heavily metamict zircon crystals is not unexpected. For the case of the zircons analyzed from Kivikevätti, their reverse discordance is collinear with the discordant baddeleyite analyses from this study and Huhma *et al.* (2018), which further supports a case of reverse discordance due to the

internal redistribution of radiogenic Pb. Following these considerations, a free regression through the combined baddeleyite and zircon fractions provide an upper intercept age of 1975.6 ± 9.7 Ma, within error of the previous date by Huhma *et al.* (2018) at 1980 ± 4 Ma.

The three chemically abraded single grain zircon fractions from the Eno dyke include discordant and near-concordant results. The weighted mean of the $^{207}\text{Pb}/^{206}\text{Pb}$ ages from the fractions is calculated at 1940 ± 3 Ma. However, the meaning of this date is unclear given the known geological history of Karelia at *ca.* 1.9 Ga. Ocean opening is dated to as late as 1.95 Ga based on ophiolite studies related to the Pechenga-Onega LIP (Peltonen *et al.*, 1996, 1998) and within this context, the date of 1940 Ma on the Eno dyke is at least 10 Ma younger than expected. These younger results may be explained by early lead loss occurring within *ca.* 200 Ma of crystallization coinciding with the regional metamorphism related to the Svecofennian orogeny. A regression calculated through the least discordant Zr1 fraction and anchored at 1800 ± 100 Ma provides an upper intercept of *ca.* 1980 Ma, and this is proposed as the preliminary age of the Eno dyke. In fact, the geochronological record throughout Karelia has been overprinted by the Svecofennian orogeny as evidenced by many examples of Paleoproterozoic and Archean rocks with lower intercepts *ca.* 1.8-1.9 Ga (Vaasjoki, 2001a; Huhma *et al.*, 2018).

The two chemically abraded zircon crystals analyzed from the WNW-trending Tilsanvaara dyke yielded significantly younger dates than the baddeleyite grains (2292 ± 8.8 Ma) extracted from the same sample. Both zircon fractions are concordant and their weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ ages provides a date of 1972.8 ± 11.2 Ma (**Figure 3-10b**).

The sample of the Tilsanvaara dyke was collected approximately 1 m from a younger NW-trending cross-cutting dyke (**Figure 3-3a**), and samples of each dyke and

their shared contact were collected. Although not visible in hand sample, veinlets are found in the thin sections of samples collected at both the geochronology site and the contact with the younger dyke (**Figure 3-5**). Zircon crystals with comparable morphologies to those picked from the heavy mineral separate are found in veinlets, which can be traced to the contact with the younger dyke (**Figure 3-6**). Zircon is also found throughout the rock sample as overgrowths on baddeleyite and as individual crystals. While these zircon crystals occur with pyramidal crystal habits and have homogenous textures, lacking zonation typical of magmatic zircons, we note that Schaltegger (2007) has stressed that morphology is not diagnostic of origin. Instead occurrence and mineral associations should be used to determine the origin of zircon growth, in this case magmatic versus hydrothermal (Schaltegger, 2007). As zircon is found primarily in veinlets and as overgrowths in the Tilsanvaara dyke (15SD_076) sample, these are interpreted to be hydrothermal in origin and related to the younger cross-cutting NW-trending dyke. Based on these considerations, the weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1972.8 ± 11.2 Ma is tentatively attributed to the crystallization age of the younger NW-trending dyke, which intruded small veinlets into the older WNW-trending 2292 ± 8.8 Ma Tilsanvaara dyke.

3.6.2 Geochemistry

The geochemistry of basalt to basaltic andesite dykes dated at *ca.* 2.40, 2.32 and 1.98 Ga fall broadly into two geochemical groups including the Mg-tholeiites and Fe-tholeiites, which have been discussed and modelled by a number of previous studies (**Figure 3-11**; **Figure 3-12**; Vuollo *et al.*, 1992, 1995; Nykänen *et al.*, 1994; Stepanova *et al.*, 2015; Lubnina *et al.*, 2017).

Figure 3-12. Bivariate plots illustrating the difference between Mg-tholeiitic and Fe-tholeiitic geochemical groups. A. SiO₂ wt. % versus MgO. 'boninite' limits from Le Bas (2000) B. ΔNb, as calculated by (Fitton *et al.*, 1997), versus Th/Nb (after Luttinen, 2018). Data are a compilation from this study, Vuollo and Fedotov (2005); Stepanova *et al.* (2015, 2017), and **Chapter 2**.

Figure 3-13. Multi-element classification diagrams for dated dykes in this study and contemporaneous units, normalized to primitive mantle (PM) of Sun and McDonough (1989). a. The Paukkajanvaara swarm, *ca.* 1.98 Ga. The *ca.* 1.98 Ga NW-trending dyke is identified with a label (15SD_077). For comparison, the *ca.* 2.32 Ga WNW-trending Tilsanvaara dyke (15SD_076) is represented by the grey dashed line. b. The Taivalkoski-Kuito swarm, *ca.* 2.32 Ga. c. The Uolevinlehto swarm, *ca.* 2.40 Ga. Samples from this study are represented by black lines. Grey solid lines are samples from Vuollo and Fedotov (2005), and Stepanova *et al.* (2015).

The high Mg-tholeiitic geochemical group, often referred to as boninitic (Alapieti *et al.*, 1990; Vogel *et al.*, 1998; Srivastava, 2008; Pearce and Reagan, 2019), is characterized by high SiO₂ wt.% paired with high Mg numbers (**Figure 3-12**). While noted in various magmatic bodies at *ca.* 2.45 Ga and 1.9 Ga (Vuollo *et al.*, 1992; Vuollo and Huhma, 2005), the *ca.* 2400 Ma Viianki dyke (15SD_084) belonging to the Uolevinlehto

swarm is the only sample included in this study that has a high Mg-tholeiitic signature. In addition to elevated MgO values, the high Mg-tholeiites include pronounced negative Nb anomalies and slightly steeper LREE slopes on multi-element diagrams compared to those of the Fe-tholeiitic group (**Figure 3-13**). Both the flat HREE slope on the multi-element diagram and the Ti/V ratio of ~20 indicate that the high Mg-tholeiitic magmas were formed at shallow depths in the upper mantle. Geochemical characteristics the pronounced negative Nb anomalies (**Figure 3-13**) and negative ϵ_{Nd} values (**Table 3-1** and references) are suggestive of crustal contamination typical of magma generated in a continental setting. Local crustal contamination has also occurred, as indicated by the irregular dyke contacts with host rock and presence of rheomorphic dykes back-intruding into the Viianki dyke (**Figure 3-3**).

A number of studies have reported basalt to basaltic andesite dykes with Fe-tholeiitic geochemistry, and representatives of this geochemical group are identified in dykes from the *ca.* 2.40 Ga Uolevinlehto, 2.32 Ga Taivalkoski-Kuito and 1.98 Ga Paukkajanvaara swarms (Vuollo *et al.*, 1992, 1995; Nykänen *et al.*, 1994). These dykes plot within the continental flood basalt field of the V versus Ti diagram characteristic of a continental setting (**Figure 3-11c**). On a Nb/Y versus Zr/Y diagram, these samples plot at the lower boundary or below the Iceland plume array (**Figure 3-11d**). Their positive correlation between Nb/Y and Zr/Y is consistent with varying degrees of partial melting. Although the *ca.* 2.32 Ga Taivalkoski-Kuito dykes cover the full range of the sample population, the older 2.40 Ga Uolevinlehto dykes plot with proportionally lower Zr/Y and Nb/Y values compared to the younger *ca.* 1.98 Ga Paukkajanvaara dyke. This general trend is suggestive of decreasing percentage of partial melting with younger ages, but requires further

investigation with a more extensive database. On multi-element diagrams (**Figure 3-13**), the Fe-tholeiitic dykes have flat to shallow HREE and LREE slopes with negative Nb anomalies and variable enrichment in Th with respect to La. The flat HREE slopes are generally characteristic of melting at shallow upper mantle levels. The negative Nb anomalies, enrichment in Th and subtle enrichment in LREE are consistent with crustal contamination or with magmas that have been derived from a metasomatized subcontinental lithospheric mantle (Pearce, 2008). Additional evidence of crustal contamination is shown in the ΔNb (as calculated in Fitton *et al.*, 1997) versus Th/Nb diagram which demonstrates a strong negative correlation between these variables (**Figure 3-11**) and supports previous assimilation and fractional crystallization modelling by Nykänen *et al.* (1994). Additional modelling by Vuollo *et al.* (1992) and Nykänen *et al.* (1994) indicate these magmas were generated by 10-20% partial melting of lithospheric mantle. The subtle variation of flat to shallow HREE slopes may represent melting at different depths around the garnet-spinel transition in the lithospheric mantle (Nykänen *et al.*, 1994). Of the dykes that have been analyzed for Sm-Nd isotopes, the majority have ϵNd values from +0.2 to +1.8 with one outlier at -0.9 (**Table 3-1**). Results of previous modelling integrated with the Sm-Nd isotope record supports a common mantle source in which the magmas underwent lithospheric contamination during emplacement.

3.6.3 Curvilinear dyke swarm trends in the Taivalkoski block and extrapolation through Western Karelia

In order to understand the regional trends of dyke swarms in the Taivalkoski block, field observations and U-Pb geochronology are evaluated in consideration with aeromagnetic

maps published by the Geological Survey of Finland (GTK, 2016).

U-Pb geochronology confirms that the NE-trending 2329.2 ± 8.2 Ma Kakikangas dyke is contemporaneous with the WNW-trending 2299 ± 20 Ma (minimum age of 2292 ± 8.8 Ma) Tilsanvaara dyke. The 35° trend difference between the Kakikangas and Tilsanvaara dykes can be reconciled with examination of the aeromagnetic map (**Figure 3-2**). The NE-trending Kakikangas dyke and proximal subparallel dykes including Kallioniemi, can be followed on the aeromagnetic map, as they characteristically have high magnetic susceptibility and therefore occur as lineaments on aeromagnetic maps (**Table 3-1**). From northeast to southwest (**Figure 3-2**), the trend of the NE-trending dykes swings towards the west such that they align with the WNW-trending 2292 ± 8.8 Ma Tilsanvaara dyke and the 2333 ± 10 Ma Törinkuru (AD13) dyke of Salminen *et al.* (2014). Together, these features can be followed up to ~ 15 km from the western margin of the Taivalkoski block, where they disappear. There are a number of prominent linear features on the aeromagnetic map that have been previously interpreted as dykes by GTK (2012) within the Taivalkoski block. Those that run parallel to the curvature defined by the Kakikangas and Tilsanvaara dykes are assigned to the *ca.* 2.32 Ga Taivalkoski-Kuito swarm as well.

The northeast continuation of the Kakikangas and parallel dykes on the aeromagnetic map are limited as they end ~ 4 km along strike of the Kakikangas locality at a structure related to the OSZ. Although a set of NE-trending lineaments reappear on the aeromagnetic map ~ 10 km NE along strike, further extrapolation is recommended only after U-Pb geochronology can confirm their contemporaneity. It is noted that within the Kuhmo and Iisalmi blocks, *ca.* 2.32 Ga dykes are found with NW trends ($295^\circ \pm 10^\circ$) in the Kuhmo and Taivalkoski blocks but these are not traceable on the aeromagnetic maps to the

Taivalkoski-Kuhmo boundary.

The NW-trending dyke at the Tilsanvaara locality has a tentative age of 1972.8 ± 11.2 Ma. Lineaments with high aeromagnetic signatures and NW trends are abundant in the Taivalkoski block and there are a number of dykes that are subparallel to the NW-trending dyke at Tilsanvaara that are interpreted from aeromagnetic maps by GTK (2016). In similarity to the *ca.* 2.32 Ga Taivalkoski-Kuito dykes, the NW-trending dykes can be followed on the aeromagnetic maps because of their relatively high magnetic susceptibility. South of Tilsanvaara, the NW-trending lineaments can be followed to the OSZ, where their approach is marked by a swing in trend towards the east until they end abruptly. On the southern side of the OSZ, *ca.* 1.98 Ga dykes are mapped with NW-trends that are also associated with prominent aeromagnetic lineaments, which have been dated at the Kivikevätti, Eno and Paukkajanvaara sites (**Table 3-1**). The NW-trending dykes on the southern side of the OSZ are also curved locally near the shear zone. In this case, the dykes curve subtly towards the north and end at the shear zone (**Figure 3-2**).

3.6.4 A primary feature or result of deformation? Explanations for the curvature of dykes

A curvature or swing in dyke swarm trends is a relatively common occurrence. Some of the most spectacular examples include the *ca.* 1270 Ma Mackenzie or 2450-2480 Ma Matachewan dyke swarms in Canada (Fahrig, 1987; LeCheminant and Heaman, 1989; Bates and Halls, 1991; West and Ernst, 1991). As in these cases, among others, curvature of dyke swarms may either be primary, occurring at the time of emplacement, or caused by post-emplacement deformation.

Causes of primary curvature have been attributed to regional stress patterns (e.g., Mackenzie dykes; Ernst *et al.*, 1995; Hou *et al.*, 2010) or by dykes intersecting and intruding along faults (e.g., Mull dykes; MacDonald *et al.*, 2015). Mechanisms responsible for secondary curvature include brittle faulting, which may also be accompanied by lateral displacement of dykes (e.g., Matachewan dykes; Bates and Halls, 1991; West and Ernst, 1991) or by ductile deformation associated with shear zones (e.g., Florida dyke swarm; Halls *et al.*, 2001 and references therein).

The mechanism responsible for the curvature of the *ca.* 2.32 Ga Taivalkoski-Kuito and *ca.* 1.98 Ga Paukkajanvaara dyke swarms is interpreted to be secondary either by brittle or ductile deformation associated with the OSZ. Evidence in support of deformation includes the fact that the distinctly trending Taivalkoski-Kuito and Paukkajanvaara swarms show consistent curvature, swinging towards the east when approaching the OSZ from the north to the south (**Figure 3-2**). Additionally, the Paukkajanvaara dykes swing towards the north when approaching the OSZ from the south. This swing in dyke swarm trends is consistent with drag associated with the sinistral displacement along the OSZ. Furthermore, this localized occurrence of curvature rules out far-field influence from regional stress patterns. Intersection with a fault during emplacement is discounted because the curvature associated with the dykes continues well beyond the OSZ structures and because not all of the curvature is parallel to the OSZ structures that do exist.

The nature of the deformation resulting in the curvature of these dykes remains unresolved as either brittle or ductile deformation maybe plausible. If brittle deformation is the cause, then their displacement may be attributed to local (meter scale) brittle faulting, as observed in the Matachewan dykes by West and Ernst (1991). Alternatively, an example

of curvature of dyke swarms attributed to ductile deformation related to a shear zone is the *ca.* 1790 Ma Florida dyke swarm, located in the Rio de la Plata craton (Uruguay). Comparable in areal extent and degree of curvature to the Taivalkoski-Kuito and Paukkajanvaara swarms, the Florida dykes are associated with the dextral Sarandi del Yí shear zone (Halls *et al.*, 2001).

Figure 3-14. Aeromagnetic map of a selected area within the Taivalkoski block (see **Figure 3-2** for extent). Arrows point to lineaments with low magnetic signatures that may be representative of local faults related to the Oulujärvi shear zone. The grey-scale aeromagnetic units are reported in nanoteslas, nT. Aeromagnetic map from GTK (2016).

According to Bossi *et al.* (1993), despite drag folding associated with the shear zone, the tholeiitic basaltic andesite dykes are unmetamorphosed. This is similar to the petrographic characteristics of the Kakikangas dyke. While the Kakikangas dyke does

include alteration (chlorite and amphibole replacing pyroxene and sericitized plagioclase), the dykes have also preserved many of their primary textures (**Figure 3-7**). If ductile deformation is the cause of the $\sim 40^\circ$ swing in dyke trends within the Taivalkoski block, curvature caused purely by ductile deformation is expected to be accompanied by metamorphism unless the strain is accommodated by local, subsidiary faults. Current mapping in the vicinity of the OSZ, however, is only available at regional scales (e.g., Digital bedrock map, 1:200,000; GTK, 2012). But examination of the aeromagnetic map identifies linear features with low aeromagnetic signatures that cross-cut the dykes (**Figure 3-14**) which may represent local faults capable of accommodating strain imposed by the ductile shear system and limiting the dykes to only low grade metamorphism.

3.7 Conclusions

U-Pb geochronology is presented for seven mafic dykes in Western Karelia belonging to three distinct intraplate magmatic events including the *ca.* 2.40 Ga Uolevinlehto, *ca.* 2.32 Ga Taivalkoski-Kuito and *ca.* 1.98 Ga Paukkajanvaara dyke swarms.

A minimum age for the Viianki dyke is determined to be 2400.7 ± 1.8 Ma and is the best estimate for the timing of crystallization. This age confirms the presence of the EW-trending *ca.* 2400 Ma swarm in the Kuhmo block, which has previously been recognized in Russian Karelia by Stepanova *et al.* (2017) and in the Pudasjärvi block at the Uolevinlehto locality (**Chapter 2**). The Viianki dyke has a high Mg-tholeiitic signature, which is consistent with other dykes part of this swarm.

Three mafic dykes are dated with ages *ca.* 2.32 Ga, belonging to the Taivalkoski-

Kuito dyke swarm: (1) Iisalmi block: NW-trending Tulisaari dyke at 2320 ± 13 Ma; (2) Taivalkoski block: NW-trending Tilsanvaara dyke with a minimum age of 2292 ± 8.8 Ma, and (3) Taivalkoski block: NE-trending Kakikangas dyke at 2329.2 ± 8.2 Ma.

The *ca.* 1.98 Ga Paukkajanvaara dyke swarm is dated at three localities within the Kuhmo and Taivalkoski block. Addition of chemically abraded zircon and single grain baddeleyite analyses to the previously dated Kivikevätti dyke provide an estimated age of crystallization of 1975.6 ± 9.7 Ma. The chemically abraded zircons analyzed from the Eno dyke provide a minimum date of 1940 ± 3 Ma but this is too young given the known tectonic history of the area and the likely age of the dyke is proposed to be *ca.* 1980 Ma. The NNW-trending dyke at the Tilsanvaara locality is considered to have been dated by proxy at 1972.8 ± 11.2 Ma via chemically abraded hydrothermal zircons extracted from the older NW-trending dyke that it cross-cuts.

The geochemistry of the dykes part of the Taivalkoski-Kuito and Paukkajanvaara swarms are nearly identical, both bearing Fe-tholeiitic geochemical signatures. This geochemistry is suggestive that they were derived from a common source and produced under similar conditions (shallow upper mantle), with melting likely driven by separate mantle plume events. Their matching geochemical signatures means that dyke assignments in this area should rely on U-Pb geochronology or paleomagnetic investigations.

A regional analysis of the dyke trends from the dated *ca.* 2.32 and 1.98 Ga swarms paired with aeromagnetic maps finds that the curvature noted in both dyke swarms is consistent with the sinistral deformation associated with the Oulujärvi shear zone. Future paleomagnetic studies and more detailed structural geology mapping are recommended for further confirmation of this interpretation.

Chapter 4

Ni-Cu-PGE potential of the Nipissing sills as part of the ca. 2.2 Ga Ungava large igneous province

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4.1 Abstract

This report summarizes mineralization associated with the Nipissing Diabase sills in the broader Sudbury area and evaluates the geochemical link between these sills and potential feeder dykes that are part of the broader *ca.* 2.2 Ga Ungava large igneous province. New whole-rock major and trace element data are presented for the Nipissing sills and these are combined with previously published data. Sm-Nd isotope data are presented for the Maguire, Senneterre and Klotz dykes, and the Triangle Mountain sill, providing key constraints on their magmatic sources and history of crustal assimilation. To help constrain the temporal evolution of the magmatic system, a new U-Pb baddeleyite age has been determined for the Triangle Mountain sill at 2216.5 ± 2.1 Ma. We also report a previously obtained, precise, U-Pb zircon age for the Ni-Cu-PGE mineralized Shakespeare intrusion at $2217.0 +1.7/-1.5$ Ma. Together, whole-rock major and trace elements, Sm-Nd isotopes, and U-Pb geochronology are used to gain a more complete picture of how the mineralized components of the *ca.* 2.2 Ga Ungava large igneous province were generated.

4.2 Introduction

The *ca.* 2.2 Ga Ungava large igneous province (LIP) includes six discrete mafic dyke swarms and sub-swarms and one major sill province (**Figure 4-1**; Ernst and Bleeker, 2010). Converging on the northeastern margin of the Superior craton, the Senneterre (2214 Ma), Maguire (2229 Ma), and Klotz (2209 Ma) dykes define a radiating dyke swarm pattern that can be followed laterally for over 1000 km (Buchan *et al.*, 1998; Ernst and Bleeker, 2010). The Anuc (2220 Ma), Couture (2199 Ma), and Kogaluk Bay (2212 Ma) dykes are also recognized as part of the Ungava LIP but are volumetrically minor and localised within the Ungava Peninsula of northern Quebec (**Table 4-1**; **Figure 4-1**; Maurice *et al.*, 2009; Ernst and Bleeker, 2010; Maurice and Francis, 2010). The Nipissing Diabase sills are the most distal component of the Ungava LIP, and are considered to have been fed by the Senneterre dykes because of their contemporaneity (Buchan *et al.*, 1998). The Nipissing sills reach as far as 1500 km from the inferred magmatic centre. The sills are hosted within the Paleoproterozoic 2.5-2.3 Ga Huronian Supergroup (Bennett *et al.*, 1991; Young *et al.*, 2001) and are also found at or near the unconformity between the Huronian Supergroup and Archean basement rocks.

Long recognized as viable targets for magmatic Ni-Cu-PGE deposits, among other commodities such as hydrothermal Co-Ag-Au veins, the sills are also proposed to have contributed to the metal endowment of the Sudbury Igneous Complex (**Table 4-2**; **Figure 4-2**; Card and Pattison, 1973; Lightfoot and Naldrett, 1996; James *et al.*, 2002; Sproule *et al.*, 2007; Potter *et al.*, 2010). Here we compile and summarize present knowledge on mineralization associated with the Nipissing sills, including a brief overview of the recently mined Shakespeare deposit (Sproule *et al.*, 2007; Dasti, 2014). We include a refined U-Pb

baddeleyite age of one Nipissing sill (Triangle Mountain), for which previously published results (Noble and Lightfoot, 1992) suggested a younger magmatic pulse. The new data for the Triangle Mountain sill suggests an age of *ca.* 2216 to 2217 Ma, in line with most other U-Pb ages of Nipissing sills and allowing for a single, short-lived, magmatic event. We also report a previously obtained but unpublished U-Pb zircon age for the mineralized Shakespeare intrusion at 2217.0 +1.7/-1.5 Ma (Davis, 2002a, 2002b), which supports the conclusion of a short-lived magmatic event at *ca.* 2217 Ma. This zircon-based age is arguably the most precise among a dozen or so ages for Nipissing sills and related units (most of which are relying on baddeleyite) and could be further refined using modern chemical abrasion methods. New geochemical data for the Nipissing sills (Easton and Hrominchuk, 2002; Easton, 2006, 2007; Easton and Sykora, 2007; Gordon, 2016; Jobin-Bevans, 2016) and bedrock mapping in the Sudbury area make it possible to refine decades-old genetic models (e.g., Lightfoot and Naldrett, 1996; Jobin-Bevans, 2004) for this mineralized magmatic system.

Table 4-1. Summary of U-Pb ID-TIMS geochronology on mafic dykes and sills related to the Ungava large igneous province (LIP).

#	Locality	Long. (W)	Lat. (N)	Fraction details	Age (Ma)	References
<i>Dykes</i>						
1	Maguire (site 71)	72.540	58.090	multigrain bdy	2229+35/-20	Buchan <i>et al.</i> , 1998
2	Anuc	75.5151	59.1853	multigrain bdy	22220±1	Maurice <i>et al.</i> , 2009
3	Senneterre (site 30)	77.36	48.19	single/multigrain bdy	2214±12.4; 2216+8/-4	Buchan <i>et al.</i> , 1993, 1996
4	Kogaluk Bay	77.2326	59.1407	multigrain bdy	2212±3.5	Maurice <i>et al.</i> , 2009
5	Couture	75.4836	60.3462	multigrain bdy	2199±5	Maurice <i>et al.</i> , 2009
6	Klotz (site 73)	73.89	60.58	multigrain bdy and abraded zrc	2209.7±0.8	Buchan <i>et al.</i> , 1998
<i>Nipissing sills</i>						
8	Shakespeare ¹	81.8327	46.3500	TIMS, abraded zrc	2217±1.7	Davis, 2002b, 2002a; this study
9	Triangle Mtn. (BNB-16-31)	79.8578	47.5365	TIMS, multigrain bdy	2209.6±3.5; 2216±2.1	Noble and Lightfoot, 1992; this study
27	Castle Mine	80.7412	47.6793	multigrain bdy	2219.4+3.6/-3.5	Corfu and Andrews, 1986
29	Morin-Otter (12CM228)	83.5200	46.5800	multigrain bdy	2214±7.7	Gordon, 2016
32	Kerns	79.8418	47.5934	multigrain bdy	2217.2±4	Noble and Lightfoot, 1992
42	Clover leaf (BNB-13-048A)	80.9692	46.4373	zrc and bdy	2215±1 (minimum)	Bleeker <i>et al.</i> , 2015

Notes: Numbers (#) correspond to **Figure 4-1**. Coordinates in WGS1984; Long. – Longitude; Lat. – Latitude. All samples were processed by ID-TIMS methods unless otherwise indicated; ID-TIMS – isotope dilution thermal ionization mass spectrometry; TIMS – thermal ionization mass spectrometry; CA– chemical abrasion; bdy – baddeleyite; zrc – zircon. ¹Approximate location of the West Pit at the Shakespeare property (not an exact location).

Figure 4-1. Map of the *ca.* 2.2 Ga Ungava large igneous province. Dykes and sills are compiled from Buchan and Ernst (2004), Maurice and Francis (2010), Ontario Geological Survey (2011). Geology of the Superior craton is modified from Goodfellow (2007). Locations of published ages relevant to this study are indicated by white circles with numbers that correspond to **Table 4-1**. Black box indicates the extent of **Figure 4-2**.

Figure 4-2. Distribution of the Nipissing Diabase sills highlighting the locations of the Shakespeare intrusion (#8) and Triangle Mountain (#9) sill (modified from Ontario Geological Survey, 2011). Numbered symbols correspond to **Table 4-2**. Locations of Ni-Cu±PGE developed mines, prospects, and mineral reserves from Ontario Geological Survey (2018). Geology of the Superior craton is modified from Goodfellow (2007).

Table 4-2. Summary of Nipissing sills discussed in literature with comments on Ni-Cu-PGE mineralization.

#	Locality	Long. (W)	Lat. (N)	Mineralization comments ¹	Reference
<i>Mineralized Nipissing sills, Ni-Cu+/- PGE</i>					
7	Big Swan	81.7667	46.4000	Ni-Cu-PGE occurrence/prospect	Turcott, 2002; Sproule <i>et al.</i> , 2007
8	Shakespeare	81.8327	46.3500	Ni-Cu-PGE deposit	Sproule <i>et al.</i> , 2007; Dasti, 2014
9	Triangle Mountain, BNB-16-031	79.8578	47.5365	Cu, Ni-Co, Ag occurrence	Noble and Lightfoot, 1992; this study
10	Milner Lake	80.8155	47.6243	Developed Ag prospect (no reserves); Ni-Cu-Co prospects/occurrences	Conrod, 1989; Lightfoot and Naldrett, 1996a, 1996b
11	Emerald Lake	80.2263	46.9250	Cu occurrence	Lightfoot and Naldrett, 1996a, 1996b
12	Wanapitei Lake	8.6762	46.7619	Ni-Cu-PGE, Ag-Au and Au-Cu-Ni prospects	Lightfoot and Naldrett, 1996a, 1996b; Sproule <i>et al.</i> , 2007
13	Kukagami Lake/ Kelly	80.5035	46.7435	Ni-Cu-PGE and Au-Cu vein occurrence	Lightfoot <i>et al.</i> , 1993; Lightfoot and Naldrett, 1996a, 1996b; Sproule <i>et al.</i> , 2007
14	Davis-Kelly	80.4556	46.7065	Ni-Cu-PGE and Au-Cu occurrence; Au Prospect	Sproule <i>et al.</i> , 2007
15	Janes	80.4063	46.6780	Ni-Cu-PGE occurrence	Sproule <i>et al.</i> , 2007
16	Bonanza Lake	80.7168	46.6740	Past producing Au mine (no reserves); Au, Ni-Cu occurrences	Conrod, 1989; Lightfoot and Naldrett, 1996a, 1996b
17	Makada Lake	81.1475	46.3766	Ni-Cu occurrence	Sproule <i>et al.</i> , 2007
18	Basswood Lake	83.4403	46.3649	Cu occurrence/prospect,	Lightfoot <i>et al.</i> , 1993; Lightfoot and Naldrett, 1996a, 1996b
19	Louis Lake	81.3969	46.3156	Ni-Cu occurrence	Sproule <i>et al.</i> , 2007
20	Lac Panache	81.4105	46.2167	Cu-Ni occurrence	Sproule <i>et al.</i> , 2007
21	Casson Lake	81.5865	46.1504	Cu-Ni-Cr-(PGE) occurrence	Sproule <i>et al.</i> , 2007
<i>Mineralized Nipissing sills, Ag-Co veins</i>					
22	Englehart	79.7600	47.8343	Ag±Co-Cu-Au-Zn-Pb prospect/occurrence vein	Lightfoot and Naldrett, 1996a, 1996b
23	Portage Bay	79.7541	47.4887	Co-Ag vein occurrence	Conrod, 1989; Lightfoot and Naldrett, 1996a, 1996b
24	Cross Lake	79.6418	47.4265	Past producing Cu (vein) mine (no reserves), Cu±Ni-Au vein	Conrod, 1989; Lightfoot and Naldrett, 1996a, 1996b
25	Skunk Lake	80.1698	46.8990	Cu occurrence in qtz-carb veins	Simony, 1964; Lightfoot and Naldrett, 1996a, 1996b
26	Bruce mines	83.7808	46.2926	Prospect and past producer without reserves, vein Au-Cu-Ag	Lightfoot <i>et al.</i> , 1993; Lightfoot and Naldrett, 1996a, 1996b
27	Castle Mine/Miller lake	80.7412	47.6793	Past producing Ag±Ni-Co-Cu-Bi-Pb mine (no reserves)	Andrews <i>et al.</i> , 1986; Corfu and Andrews, 1986; Lightfoot and Naldrett, 1996a, 1996b
28	Duncan Lake	80.9810	47.6746	Ag-Au occurrence	Conrod, 1989; Lightfoot and Naldrett, 1996a, 1996b
29	Morin-Otter (12CM228)	83.5217	46.5877	Ni-Co-Cu-Ag±Bi-Au vein project	Gordon, 2016
<i>Other sills</i>					
30	HWY 144 (BNB-16-030)	81.5969	46.7621		Bleeker, work in progress, 2018
31	May Twp. sill (BNB-17-039)	82.0683	46.2170		Bleeker, work in progress, 2018
32	Kerns	79.8418	47.5934		Noble and Lightfoot, 1992; Lightfoot and Naldrett, 1996a, 1996b
33	Beaton Bay	80.7909	47.6193		Conrod, 1989; Lightfoot and Naldrett, 1996a, 1996b
34	Sharp Lake dyke (BNB-16-032)	79.7359	47.4149		This study
35	Slide Rock	80.0831	47.0775		Lightfoot and Naldrett, 1996a, 1996b
36	Obabika Lake	80.2321	47.0248		Simony, 1964; Lightfoot and Naldrett, 1996a, 1996b
37	Herridge Lake (BNB-16-033A)	79.8099	46.9854		This study
38	Narrows Island	80.0884	46.9144		Simony, 1964; Lightfoot and Naldrett, 1996a, 1996b
39	High Rock	80.0249	46.8990		Lightfoot and Naldrett, 1996a, 1996b
40	Morin-Otter Twp. (BNB-16-029)	81.0440	46.4205		This study
41	Elliot Lake	82.4544	46.3934		Lightfoot <i>et al.</i> , 1993
42	Clover leaf (BNB-16-028)	80.9700	46.4372		Bleeker <i>et al.</i> , 2015

Notes: Numbers (#) correspond to **Figure 4-2**. Coordinates in WGS1984. ¹Mineralization information from Ontario Geological Survey (2018). Abbreviations: Qtz-carb, Quartz-carbonate; Twp., township.

Although both the Nipissing sills and potential feeder dykes have been the subject of more recent local studies (Sproule *et al.*, 2007; Maurice and Francis, 2010), their genetic connection has not yet been tested through geochemistry nor has the overall magmatic system been considered in its entirety for the genesis and targeting of Ni-Cu-PGE mineralization.

4.2.1 Key scientific questions and progress

This study aims to conduct a survey of mineralization associated with the Nipissing sills, leading to a magmatic province-wide genetic model. In the process, we will address several questions about: a) the mineralization of the Nipissing sills in the Sudbury area; and b) genetic connections to the broader *ca.* 2.2 Ga Ungava LIP. Specific questions are:

- i) Do detailed geochemical data allow the genetic link between the Senneterre and Maguire dykes and the Nipissing sills, as proposed by Buchan *et al.* (1998)?
- ii) Relevant to the source of sulphur, is the negative ϵNd signature of the Nipissing sills reported by Lightfoot and Naldrett (1996) exclusively a result of in situ crustal assimilation, or was this signature acquired, at least partially, ‘up stream’ in the overall magmatic system, prior to final emplacement?
- iii) Using geochemical approaches for evaluating metallogenic potential and generation of LIPs (e.g., Pearce, 2008; Jowitt and Ernst, 2013; Pearce *et al.*, 2017) combined with geochronology, can we identify new patterns in the geochemistry that may aid in the regional targeting of the Nipissing sills?

- iv) How do these geochemical results and interpretations relate to the Superia supercraton paleogeographic reconstruction?

4.2.2 Characterisation of the Nipissing Diabase sills

The Nipissing sills were emplaced between *ca.* 2220 Ma and *ca.* 2210 Ma and are exposed from across the Ontario-Quebec border in the east to Sault Ste. Marie in the west, over an extent of more than 400 km. Sills are generally (sub)concordant to bedding of the Huronian Supergroup or follow the unconformity between the Archean basement and overlying sediments (e.g., **Figure 4-3**; Corfu and Andrews, 1986; Noble and Lightfoot, 1992; Buchan *et al.*, 1993, 1996; Lightfoot *et al.*, 1993; Bleeker *et al.*, 2015). In other cases, sills, sheets and irregularly shaped intrusions may be observed crosscutting Archean basement or stepping across Huronian stratigraphy (Lightfoot *et al.*, 1993).

Geometric and petrological characteristics of the Nipissing sills vary between localities (Hriskevich, 1968; Conrod, 1989; Lightfoot *et al.*, 1993). Intrusive bodies vary in thickness from tens to hundreds of metres and in strike length from one to ten kilometres or more. They vary in texture from homogeneous and undifferentiated lithologies to differentiated bodies with distinct lithological units (Lightfoot *et al.*, 1993; Lightfoot and Naldrett, 1996a). Commonly, differentiated sills include bounding units with upper and lower chilled margins of diabase and quartz diabase (Conrod, 1989; Lightfoot *et al.*, 1993; Lightfoot and Naldrett, 1996a). Overlying basal units, a typical differentiation series evolves from hypersthene gabbro to gabbro-norite, all of which is capped by a pegmatoidal varitextured gabbro (Conrod, 1989; Lightfoot *et al.*, 1993). In these upper units, angular to subangular rafts of host rock (e.g., Lorrain or Gowganda formations) and autoliths or

xenoliths of diorite are common (Conrod, 1989; Lightfoot *et al.*, 1993).

The differentiation series of the Nipissing sills and their geochemical evolution is attributed to coupled fractional crystallization and assimilation processes (Lightfoot and Naldrett, 1989, 1996a). The major element chemistry of the Nipissing sills varies with the degree of differentiation. Chilled margins and bounding quartz diabase units have major element chemistry typical of tholeiitic basalt (Lightfoot and Naldrett, 1989, 1996b, 1996a; Lightfoot *et al.*, 1993). In contrast, the upper differentiated units are characterized by more felsic and calc-alkaline signatures (Lightfoot *et al.*, 1993). The sills are enriched in light rare earth elements (LREE) and large ion lithophile elements (LILE), and are depleted in Nb, TiO₂ and P₂O₅, with respect to primitive mantle (Lightfoot *et al.*, 1993).

Lightfoot and Naldrett (1996a) conducted a Sm-Nd isotopic study across several differentiated sills (e.g., Kerns, Basswood Lake, Emerald Lake, Obabika, and Portage Bay) to better constrain the influence of assimilation of host rock on the overall composition of the Nipissing sills. Since the distinct lithological units within any given sill do not plot on a single isochron, and the most evolved lithologies trend towards host rock signatures (Lorrain and Gowganda formations), it was concluded that the more evolved magmas were contaminated by in situ crustal assimilation (Lightfoot and Naldrett, 1989, 1996a).

4.2.3 Geochemistry of the feeder systems

The geochemistry of Paleoproterozoic mafic dykes (2.5–2.0 Ga) of the Ungava Peninsula has been characterized using whole-rock major and trace element geochemistry by (Maurice and Francis, 2010). Their work included (but was not limited to) the Maguire, Anuc, Kogaluk Bay, Klotz, and Couture dyke swarms. Three distinct geochemical groups

were recognized, including the (1) high Fe-Ti, (2) low Fe-Ti, and (3) low REE groups. These groups were identified based on abundances of Fe_2O_3 and TiO_2 W % while tracking relationships with Fe, Ti and Al to Mg, and comparing relative enrichment of cations such as Na and trace element ratios, including Ti/V (Maurice and Francis, 2010). These three geochemical groups are not limited to particular dyke swarms or trends, and no secular evolution related to geochemistry has been recognized. Instead, the geochemical groups are found across various dyke swarms of different ages and trends (Maurice *et al.*, 2009; Maurice and Francis, 2010).

The high Fe-Ti group is defined by greater than 14 wt.% Fe_2O_3 , greater than 1.4 wt.% TiO_2 , and proportionally higher Na and Ti/V ratios (Maurice and Francis, 2010). There are two subgroups within the high Fe-Ti group. These are distinguished by the relative proportion of Al cations; split below (low Al) and above (high Al) 12 wt. % Al_2O_3 . The low Al subgroup is further distinguished by higher $\text{CaO}/\text{Al}_2\text{O}_3$ ratios and steeper heavy rare earth element (HREE) slopes on a multi-element plot. The high Fe-Ti (low Al) subgroup includes all samples belonging to the Kogaluk Bay and Anuc swarms as well as members of the Maguire swarm. The high Fe-Ti (high Al) subgroup does not include any dykes dated at 2.2 Ga but includes numerous unclassified dykes. A peridotite and pyroxenite mantle that incorporated crustal and alkaline components is the proposed source for this geochemical group (Maurice and Francis, 2010).

Characteristics distinguishing the low Fe-Ti group are greater than 11 wt.% Fe_2O_3 and less than 0.8 wt.% TiO_2 , positive correlations between Fe, Ti and Al to Mg, low Ti/V values, and relatively high Mg with respect to SiO_2 (Maurice and Francis, 2010). Due to their characteristically low HREE and high field strength element (HFSE) abundances, a

depleted refractory harzburgitic source is proposed for this geochemical signature (Maurice and Francis, 2010). This geochemical group includes members of the Maguire dykes.

The low REE group includes dykes belonging to the Klotz and Couture swarms. With a broad range of Fe₂O₃ and TiO₂ concentrations overlapping with the high Fe-Ti group, the low REE dykes are distinguished by lower Na₂O concentrations, Ti/V ratios of typical continental flood basalts and flat REE patterns. The quartz normative, low Mg characteristics of the low REE group are interpreted as resulting from fractionation and evolution of a melt extracted from a lherzolitic mantle source (Maurice and Francis, 2010).

4.2.4 Mineralization

4.2.4.1 General characteristics

Mineralization associated with the Nipissing sills occurs as magmatic Ni-Cu-PGE sulphides and secondary hydrothermal Co±Ag-Cu-Ni polymetallic veins (James *et al.*, 2002; Vaillancourt *et al.*, 2003; Sproule *et al.*, 2007; Potter *et al.*, 2010). Historic operations include about a dozen past producing mines, which exploited the sills for their hydrothermal cobalt-silver vein deposits located within the Coleman, Gillies Limit, South Lorrain and Plummer Additional townships (Potter *et al.*, 2010; Ontario Geological Survey, 2018). According to the Ontario Geological Survey *Mineral Deposit Inventory* (2018), the Nipissing sills include a number of developed prospects with (n = 1) and without (n = 3) reserves, prospects (n = 8), and occurrences (n = 98) reporting various combinations of Cu-Ni±PGE-Au-Ag (**Figure 4-2**). Several styles of magmatic Ni-Cu-PGE mineralization are reported in the Nipissing sills. These include disseminated Cu-Ni-Cr-

(PGE) pipes at Casson Lake, Ni-Cu-PGE sulphides disseminated internally at the Kelly and Davis-Kelly localities, massive to disseminated Ni-Cu-Co-PGE sulphides at the Louise and Waters localities, basal Ni-Cu-PGE sulphides with massive and disseminated textures at the Janes locality, and remobilized basal Ni-Cu-PGE at Rathburn Lake (**Table 4-2**; James *et al.*, 2002; Vaillancourt *et al.*, 2003; Sproule *et al.*, 2007; Dasti, 2014).

4.2.4.2 The Shakespeare intrusion

The Shakespeare intrusion represents the largest documented occurrence of Ni-Cu-PGE mineralization and one of the most recent developments exploiting a Nipissing sill (Dasti, 2014). It is located 70 km west of Sudbury, near the northern shore of Agnew Lake. The sill-like intrusion has a thickness that varies from 300 to 430 m, and a strike length of 14 km, dipping north between 40 and 80° (**Figure 4-3**; Sproule *et al.*, 2007). The rusty detrital pyrite (and uraninite) bearing formations of the lower Huronian Supergroup (Young *et al.*, 2001) which host the Shakespeare intrusion may have provided a local source of sulphur. The Shakespeare intrusion is identified as a Nipissing sill by its coeval U-Pb age at 2217 Ma and matching whole-rock major and trace element signatures (Sutcliffe *et al.*, 2002). Although production is currently on hold, the mine was in operation from 2010 to 2012. The most recent resource estimates of the intrusion include 11.8 Mt of probable ore and 5.44 Mt of indicated and inferred resources with 0.33% Ni, 0.35% Cu, and 0.87 g/t PGM and Au (Dasti, 2014).

Consistent with many Nipissing sills, assimilation and fractional crystallization control the differentiation series within the Shakespeare sill. Sproule *et al.* (2007) explain that the lithologies of the sill can be divided into an uneconomic, sparsely mineralized

Lower Group and an economic mineralized Upper Group. The Lower Group includes basal pyroxenite overlain by gabbro (**Figure 4-3**; Sproule et al., 2007; Dasti, 2014). Chilled against the Lower Group, the Upper Group hosts mineralization starting from the lower contact melagabbro, which differentiates upwards to a quartz gabbro and biotite-quartz diorite (Sproule et al., 2007; Dasti, 2014). Also restricted to the Upper Group are lithic fragments of angular to subrounded quartzite, as well as autoliths and xenoliths of diorite (Sproule et al., 2007; Dasti, 2014). The lithological units of the Shakespeare intrusion exhibit pervasive alteration of pyroxene to chlorite, minor sericite alteration of plagioclase, and variable sulphide recrystallization observed in the pyroxenite, gabbro, and biotite-quartz diorite (Sproule et al., 2007; Dasti, 2014). Chlorite and sericite alteration intensity varies with metamorphic grade from greenschist to amphibolite facies and is related to the 1.9 Ga Penokean Orogeny (Sproule et al., 2007).

Mineralization styles observed in the Upper Group involve ore mineral assemblages of pyrrhotite, chalcopyrite, and lesser pyrite (Sproule *et al.*, 2007; Dasti, 2014). Melagabbro belonging to the Upper Group is associated with sulphide abundances of 10 to 15% as disseminated, blebby (2–5 cm) to net-textured pyrrhotite, chalcopyrite, pentlandite, and gersdorffite (Sproule *et al.*, 2007; Dasti, 2014). Within the Lower Group, accessory (<1%) finely disseminated to blebby pyrrhotite and chalcopyrite are found at the contact between the basal pyroxenite and gabbro (Sproule *et al.*, 2007; Dasti, 2014).

Shakespeare igneous stratigraphy from diamond drill holes (DDH), hung from upper contact

Figure 4-3. Geological map of the Shakespeare intrusion (modified from Card and Palonen, 1976). Diamond drillhole logs with Ni-Cu-PGE assays from Dasti (2014). All drillholes have an azimuth of 147°. Drillhole U-03-119 has an inclination of 65.9° and length of 531 m. Drillhole U-03-122 has an inclination of 75° and length of 624 m.

4.3 Methods

4.3.1 Geochronology

Geochronological analyses were completed at the Jack Satterly Geochronology Laboratory at the University of Toronto. Standard separation techniques were used to separate baddeleyite and zircon from approximately 10 kg samples of the Triangle Mountain sill and Shakespeare intrusion (**Appendix B**). Results of standards (R33, OG1 and EARTHTIME 2.0 Ga synthetic solution of Black et al., 2004; Condon et al., 2015. and Stern et al., 2019) analyzed over the duration of this study are presented in **Appendix C**.

4.3.2 Whole-rock major and trace element geochemistry

In addition to the 26 geochemical Nipissing sill samples analyzed in this study, a further 765 whole-rock and/or trace element geochemical samples of Nipissing sills were compiled (Lightfoot and Naldrett, 1989, 1996b; Easton, 2006, 2007, 2013; Easton and Sykora, 2007; Ernst and Buchan, 2010; Gordon, 2016; Jobin-Bevans, 2016).

A total of 59 whole-rock major and trace element geochemical analyses of dykes from the Kogaluk Bay (2), Couture (2), Anuc (4), Klotz (15), Maguire (10), and Senneterre (14) dyke swarms plus two analyses from unclassified dykes are also included in the database (Ernst and Buchan, 2010; Maurice and Francis, 2010).

The compiled data set includes 843 analyses. Erroneous or incomplete analyses were removed based on the following criteria: 1) major element totals less than 97% (including loss on ignition); and 2) if any of the following trace elements were absent: Y, Zr, Nb, La, Sm, Tb, Yb or Th. To eliminate samples that may represent assimilation fractional crystallization (AFC) processes, dyke and sill samples with greater than 54 wt.

% SiO₂ were also eliminated. These selection criteria resulted in a filtered data set of 137 samples, comprising 48 dyke and 89 sill samples.

4.3.3 Sm-Nd isotope systematics

Sm-Nd isotope preparation and analyses were completed at the Isotope Geochemistry and Geochronology Research Centre (IGGRC) at Carleton University, Ottawa. For detailed protocols see **Appendix B**.

4.4 Results

4.4.1 Triangle Mountain sill

Mineral separation of coarse-grained to pegmatoidal gabbro near the top of the Triangle Mountain sill yielded more than a dozen baddeleyite crystals and crystal fragments. One larger single grain and three multiple grain fractions (2–11 crystal fragments) were analyzed and resulted in weakly discordant data points that show minor scatter due to variable Pb-loss behaviour (**Table 4-3; Figure 4-4**). In general, the data are similar to but extend the range of results reported earlier by (Noble and Lightfoot, 1992). In detail, our more precise data points better resolve the complex (early) Pb-loss behaviour often seen in baddeleyite (Davis and Sutcliffe, 1985; Bleeker, 2014), and the least discordant single grain analysis indicates an older age than obtained by the earlier study. The ²⁰⁷Pb/²⁰⁶Pb age of this least discordant single baddeleyite grain indicates a minimum crystallization age of 2214±2.0 Ma. A regression through the three results that define the right edge of the combined population (i.e., analyses least affected by early Pb-loss behaviour) yields an

upper intercept of 2216.5 ± 2.1 Ma, which we consider the best estimate for the crystallization age of the Triangle Mountain sill.

4.4.2 Shakespeare intrusion

A drill core sample from the upper portion of the Shakespeare intrusion, described by (Davis, 2002a, 2002b) as quartz diorite, yielded a limited but distinct population of euhedral, pale brown, long prismatic, and flat or platy zircon crystals devoid of cores. Apatite and rutile with a range of colours was also recovered but will not be discussed here (see report by Davis (2002a, 2002b) for a detailed discussion).

Zircon crystals were air-abraded and five single grains were analyzed, resulting in variable discordance (0.6–4.6%). They proved to be relatively high in uranium, which is typical for late-stage platy zircons in mafic magmatic rocks, resulting in discordance but allowing for relatively precise analyses (**Table 4-3; Figure 4-4**). All five analyses are collinear on a U-Pb concordia diagram, allowing for a precise upper intercept age of $2217.0 \pm 1.7/-1.5$ Ma, which is close to and within error of the $^{207}\text{Pb}/^{206}\text{Pb}$ age of the single least discordant (0.6%) analysis at 2216.4 ± 1.9 Ma. Given this near-concordance, the lack of scatter, and the lack of inheritance, this upper intercept age of $2217.0 \pm 1.7/-1.5$ Ma dates the crystallization of the Shakespeare intrusion and represents one of the more precise age estimates for the Nipissing magmatic event.

Figure 4-4. Geochronology of the Triangle Mountain sill (left; this study) and Shakespeare intrusion (right; after Davis, 2002b, 2002a). Above each Concordia diagram are photos of baddeleyite and zircon grains analyzed. Analysis numbers and results correspond to **Table 4-3**. Previous data on the Triangle Mountain sill from Noble and Lightfoot (1992) are shown for comparison. See text for discussion.

Table 4-3. U-Pb isotopic data of the Triangle Mountain sill and Shakespeare intrusion.

No.	Description (analysis number)	Weight. (µg)	U (ppm)	Th/U	Pbcom (pg)	$\frac{^{207}\text{Pb}}{^{204}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	2σ	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	2σ	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	2σ	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	2σ	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	2σ	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	2σ	Disc. %
													Age (Ma)		Age (Ma)		Age (Ma)		
<i>Triangle Mountain sill - BNB-14-031D-E</i>																			
1	1 bdy (sk37p194)	-	-	0.11	0.3	3550	0.40681	0.00078	7.792	0.018	0.13891	0.00012	2200.3	3.6	2207.2	2.1	2213.7	1.5	0.7
2	2 bdy frgs (sk37p195)	0.34	701	0.13	0.5	1677	0.40599	0.00082	7.755	0.019	0.13853	0.00012	2196.6	3.8	2203.0	2.2	2208.9	1.6	0.7
3	3 bdy (sk37p196)	1.1	682	0.08	1.7	1472	0.40297	0.00090	7.706	0.020	0.13869	0.00014	2182.7	4.1	2197.3	2.3	2210.9	1.8	1.5
4	11 bdy (sk37p197)	2.5	495	0.48	2.7	1630	0.40114	0.00090	7.633	0.020	0.13801	0.00013	2174.3	4.1	2188.8	2.3	2202.4	1.6	1.5
<i>Shakespeare sill - RHS02-1</i>																			
1	1 Ab zrc, eq pale brn (dwd4207)	0.7	697	0.40	0.18	9929	0.4077	0.0011	7.820	0.023	-	-	-	-	-	-	2216.4	1.9	0.6
2	1 Ab zrc, flat, pale brn (dwd4163)	1.2	1741	0.43	1.78	4231	0.4031	0.0013	7.727	0.026	-	-	-	-	-	-	2215.2	1.6	1.7
3	1 Ab zrc, eq, crk, pale brn (dwd4162)	1.9	1109	0.34	0.90	8373	0.4016	0.0020	7.688	0.039	-	-	-	-	-	-	2212.9	1.6	1.9
4	1 Ab zrc, eq pale brn (dwd4208)	0.2	679	0.36	0.39	1252	0.4002	0.0010	7.665	0.021	-	-	-	-	-	-	2213.8	2.0	2.3
5	1 Ab zrc, stubby, pale brn (dwd4164)	0.7	1348	0.46	1.20	2737	0.3902	0.0010	7.460	0.021	-	-	-	-	-	-	2210.6	1.9	4.6

Notes: Ab – abraded; bdy – baddeleyite grain; brn – brownish; crk – cracked; eq – equant; frgs – fragment; zrc – zircon grain
Th/U calculated from radiogenic $^{208}\text{Pb}/^{206}\text{Pb}$ ratio and $^{207}\text{Pb}/^{206}\text{Pb}$ age assuming concordance.

Pbtot is total amount of Pb in picograms excluding blank.

Pbcom is total amount of common Pb in picograms assuming the isotopic composition of laboratory blank:

Triangle Mountain sill (BNB-14-0.31D-E): $^{206}\text{Pb}/^{204}\text{Pb} = 18.49$; $^{207}\text{Pb}/^{204}\text{Pb} = 15.59$; $^{208}\text{Pb}/^{204}\text{Pb} = 39.36$ (errors of 0.4%)

Shakespeare sill (RHS02-1): $^{206}\text{Pb}/^{204}\text{Pb} = 18.221$; $^{207}\text{Pb}/^{204}\text{Pb} = 15.612$; $^{208}\text{Pb}/^{204}\text{Pb} = 39.360$ (errors of 2%).

Disc. - per cent discordance for the given $^{207}\text{Pb}/^{206}\text{Pb}$ age.

Uranium decay constants are from Jaffey *et al.* (1971).

4.4.3 Whole-rock major and trace element geochemistry

Whole-rock major and trace element data produced for this study are presented in Table 4-4. To illustrate the variability of samples, all samples with major elements ($n = 778$) are plotted on a total alkali-silica (TAS) plot (**Figure 4-5**; Le Bas *et al.*, 1986). The Nipissing sills show a wide range of SiO_2 wt. % content, from alkaline or subalkaline tholeiitic basalt to rhyolite compositions. The majority of dyke samples plot within the subalkaline basalt to basaltic andesite field. Results of the filtering procedure are shown on a second TAS plot (**Figure 4-6a**). The Nipissing sills show a dramatic decrease in range and are confined to the subalkaline field from basalt to basaltic andesite. Two Nipissing sill samples plot as outliers from the general population including one melagabbro from the Shakespeare intrusion (no. 6 from Sproule *et al.*, 2007) and a coarse-grained gabbro from Boon Township (sample 12RME-3404 from Easton, 2013). Results and discussion of data hereafter will pertain only to the filtered data set unless otherwise stated.

Figure 4-5. Total alkali versus silica plot (TAS; Le Bas *et al.*, 1986), plotting all available data for sills and dykes. Data from Lightfoot and Naldrett, 1989, 1996b; Easton, 2006; Easton and Sykora, 2007; Ernst and Buchan, 2010; Maurice and Francis, 2010; Gordon, 2016; Jobin-Bevans, 2016; this study.

Table 4-4. Whole-rock major and trace element geochemistry for 2.22 Ga Nipissing sills.

Sample	Longitude (X)	Latitude (Y)	Geochemical group	Notes <i>Detection limit</i>	SiO ₂ <i>0.01</i>	Al ₂ O ₃ <i>0.01</i>	Fe ₂ O ₃ <i>0.01</i>
BNB-16-028A	-80.97	46.44	Low Fe-Ti		51.80	7.14	10.70
BNB-16-028B	-80.97	46.44	Low Fe-Ti		51.90	7.04	10.25
BNB-13-048B	-80.97	46.44	Low Fe-Ti		51.70	7.58	10.30
BNB-16-029	-81.04	46.42	Low Fe-Ti		54.10	13.25	13.50
BNB-16-030A	-81.60	46.76	Low Fe-Ti		51.60	13.90	12.50
BNB-16-030B	-81.60	46.76	Low Fe-Ti		51.20	14.00	11.80
BNB-16-030C	-81.60	46.76	Low Fe-Ti		52.20	13.75	14.60
BNB-16-030D	-81.60	46.76	Low Fe-Ti		51.00	14.20	13.85
BNB-16-030E	-81.60	46.76	Low Fe-Ti		62.40	12.05	14.65
BNB-16-030F	-81.60	46.76	Low Fe-Ti		56.80	11.80	17.85
BNB-16-030G	-81.60	46.76	Low Fe-Ti		53.20	12.45	18.00
BNB-16-032A	-79.74	47.41	Low Fe-Ti		52.20	14.50	11.70
BNB-16-032B	-79.74	47.42	Low Fe-Ti		51.80	14.00	11.70
BNB-16-031A	-79.78	47.49	Low Fe-Ti	Triangle Mtn.	51.40	14.10	12.10
BNB-16-031B	-79.78	47.49	Low Fe-Ti	Triangle Mtn.	50.60	13.65	12.00
BNB-16-031C	-79.79	47.49	Low Fe-Ti	Triangle Mtn.	57.10	13.10	16.15
BNB-16-031D	-79.79	47.49	Low Fe-Ti	Triangle Mtn.	51.10	12.85	16.45
BNB-16-031E	-79.79	47.49	Low Fe-Ti	Triangle Mtn.	50.30	13.10	16.80
BNB-16-031F	-79.79	47.49	Low Fe-Ti	Triangle Mtn.	50.60	13.90	11.80
BNB-16-033A	-79.81	46.99	Low Fe-Ti	Herridge Lake	51.00	13.90	12.40
BNB-16-033B	-79.81	46.99	Low Fe-Ti	Herridge Lake	51.70	14.20	12.55
BNB-16-033C	-79.81	46.99	Low Fe-Ti	Herridge Lake	52.30	13.65	12.45
BNB-17-039A	-82.07	46.22	Low REE	May Township	50.30	13.50	14.40
BNB-17-039B	-82.07	46.22	Low REE	May Township	53.20	14.55	11.55
BNB-17-039C	-82.07	46.22	Low REE	May Township	66.60	14.75	5.20
BNB-17-039D	-82.07	46.22	Low REE	May Township	49.50	13.50	14.20
<i>Standards</i>							
10-LT-05					49.00	18.15	9.68
10-LT-05					49.30	18.65	9.79
10-LT-05- Average					48.88	18.18	9.54
2σ					1.16	0.50	0.29

Table 4-4 continues on the next page

Notes: Major elements in wt. % and trace elements in ppm. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B are presented. Superscripts A or B follow each sample name identify the corresponding batch. The average composition of 10-LT-05 and 2σ values of 25 analyses spanning 6 years are also provided.

Table 4-4. Continued.

Sample	CaO	MgO	Na ₂ O	K ₂ O	Cr ₂ O ₃	TiO ₂	MnO	P ₂ O ₅	SrO	BaO
<i>D.L</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>	<i>0.01</i>
BNB-16-028A	10.30	15.15	1.00	0.39	0.27	0.47	0.22	0.04	0.01	0.01
BNB-16-028B	9.58	15.15	0.96	0.54	0.28	0.47	0.22	0.04	0.01	0.01
BNB-13-048B	10.80	14.60	0.88	0.44	0.25	0.46	0.20	0.03	0.01	0.01
BNB-16-029	7.72	4.38	3.11	1.28	<0.01	1.06	0.20	0.10	0.01	0.03
BNB-16-030A	8.80	6.51	2.13	1.13	0.03	0.87	0.21	0.09	0.02	0.03
BNB-16-030B	8.27	6.48	2.92	0.97	0.01	0.68	0.19	0.07	0.02	0.02
BNB-16-030C	7.51	4.57	2.41	1.86	<0.01	1.18	0.23	0.10	0.02	0.03
BNB-16-030D	8.64	5.69	2.24	1.40	<0.01	1.04	0.21	0.08	0.02	0.02
BNB-16-030E	2.74	1.02	4.74	0.67	<0.01	1.05	0.12	0.38	0.01	0.02
BNB-16-030F	3.38	2.21	3.91	0.66	<0.01	1.39	0.15	0.21	0.02	0.02
BNB-16-030G	5.46	2.93	3.00	1.65	<0.01	1.88	0.25	0.15	0.02	0.03
BNB-16-032A	8.46	6.05	2.74	1.00	0.02	0.85	0.23	0.08	0.07	0.03
BNB-16-032B	7.88	6.27	3.85	0.76	0.02	0.86	0.22	0.08	0.03	0.02
BNB-16-031A	9.69	7.49	1.99	0.94	0.02	0.78	0.19	0.06	0.02	0.01
BNB-16-031B	9.41	7.98	1.88	1.21	0.02	0.76	0.19	0.06	0.02	0.02
BNB-16-031C	4.40	1.96	3.83	1.51	<0.01	1.41	0.26	0.19	0.02	0.04
BNB-16-031D	7.17	4.05	3.22	1.41	<0.01	1.56	0.24	0.10	0.02	0.02
BNB-16-031E	7.01	4.06	2.81	1.93	<0.01	1.55	0.25	0.09	0.01	0.03
BNB-16-031F	10.25	6.81	2.00	1.17	0.02	0.77	0.20	0.06	0.02	0.02
BNB-16-033A	10.40	6.38	2.23	0.84	<0.01	0.82	0.20	0.06	0.03	0.02
BNB-16-033B	10.20	5.84	2.31	0.59	<0.01	0.77	0.20	0.06	0.02	0.02
BNB-16-033C	9.88	5.72	2.48	0.75	<0.01	0.75	0.19	0.09	0.02	0.02
BNB-17-039A	9.46	7.24	2.50	0.53	0.04	1.51	0.17	0.18	0.01	0.02
BNB-17-039B	8.03	3.73	3.12	0.69	0.01	1.57	0.14	0.34	0.02	0.01
BNB-17-039C	4.65	2.26	4.97	0.44	<0.01	0.66	0.08	0.16	0.01	0.01
BNB-17-039D	9.49	6.98	2.53	0.54	0.03	1.55	0.18	0.17	0.01	0.01
<i>Standards</i>										
10-LT-05	9.57	6.53	3.61	1.01	0.03	1.44	0.14	0.37	0.11	0.06
10-LT-05	9.50	6.51	3.59	1.03	0.03	1.45	0.14	0.38	0.11	0.06
10-LT-05- Average	9.52	6.61	3.60	1.02	0.03	1.44	0.14	0.38	0.11	0.06
2σ	0.32	0.21	0.20	0.03	0.01	0.06	0.00	0.02	0.00	0.00

Table 4-4 continues on the next page

Notes: Major elements in wt. % and trace elements in ppm. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B are presented. Superscripts A or B follow each sample name identify the corresponding batch. The average composition of 10-LT-05 and 2σ values of 25 analyses spanning 6 years are also provided.

Table 4-4. Continued

Sample	LOI	Total	Ba	Ce	Cr	Cs	Dy	Er	Eu	Ga
<i>D.L</i>	<i>0.01</i>	<i>0.01</i>	<i>0.5</i>	<i>0.5</i>	<i>10</i>	<i>0.01</i>	<i>0.05</i>	<i>0.03</i>	<i>0.03</i>	<i>0.1</i>
BNB-16-028A	1.97	99.47	79.4	10.1	1930	1.03	1.91	1.03	0.51	9.4
BNB-16-028B	2.16	98.61	85.1	11.8	1970	1.46	1.93	1.08	0.47	9.5
BNB-13-048B	1.55	98.81	82.8	10.3	1810	1.49	1.90	1.04	0.49	9.5
BNB-16-029	1.37	100.11	262.0	24.1	10	0.23	4.28	2.54	1.01	18.2
BNB-16-030A	1.78	99.60	241.0	20.1	180	1.10	3.19	1.74	0.88	17.5
BNB-16-030B	1.83	98.46	178.0	16.0	50	0.63	2.61	1.55	0.85	16.7
BNB-16-030C	1.37	99.83	278.0	23.7	10	1.71	3.98	2.17	1.04	19.2
BNB-16-030D	1.86	100.25	219.0	19.1	10	1.20	3.16	1.78	0.93	18.8
BNB-16-030E	1.02	100.87	180.0	60.7	<10	0.51	10.05	5.69	2.27	24.0
BNB-16-030F	1.08	99.48	150.0	53.3	10	0.58	7.53	4.46	2.00	23.2
BNB-16-030G	0.83	99.85	279.0	35.5	<10	0.48	5.45	3.14	1.49	22.5
BNB-16-032A	2.11	100.04	226.0	23.3	140	0.97	3.16	1.77	0.94	17.1
BNB-16-032B	2.10	99.59	177.0	22.6	150	0.58	3.19	1.98	0.89	16.8
BNB-16-031A	1.58	100.37	116.0	13.9	110	0.78	2.75	1.65	0.75	15.8
BNB-16-031B	1.49	99.29	138.5	13.2	120	1.14	2.83	1.58	0.76	15.5
BNB-16-031C	1.31	101.28	310.0	51.1	10	2.15	6.38	3.83	1.89	23.3
BNB-16-031D	1.17	99.36	210.0	27.3	10	0.95	4.24	2.38	1.08	20.7
BNB-16-031E	1.22	99.16	278.0	23.1	10	0.82	3.87	2.11	1.15	20.5
BNB-16-031F	1.84	99.46	177.5	13.7	120	1.14	2.75	1.58	0.77	16.0
BNB-16-033A	0.89	99.17	138.5	16.7	30	1.06	2.90	1.57	0.80	17.1
BNB-16-033B	0.46	98.92	145.0	17.0	10	1.88	2.83	1.81	0.90	17.7
BNB-16-033C	1.06	99.36	176.5	21.9	30	1.01	3.73	2.22	1.10	17.6
BNB-17-039A	1.04	100.90	153.0	12.9	240	1.80	5.58	3.89	0.95	18.0
BNB-17-039B	1.28	98.24	125.5	47.8	50	1.36	7.32	4.52	2.16	21.9
BNB-17-039C	0.67	100.46	110.0	74.7	30	1.64	6.52	3.36	1.77	16.1
BNB-17-039D	1.05	99.74	117.5	13.5	240	1.79	5.68	3.87	1.05	19.1
<i>Standards</i>										
10-LT-05	0.66	100.36	546.0	47.3	180	0.20	3.68	2.01	1.47	19.1
10-LT-05	0.64	101.18	547.0	49.7	190	0.25	3.70	1.97	1.66	20.5
10-LT-05- Average	0.75	100.27	538.3	48.0	184	0.23	3.63	1.97	1.59	19.9
2 σ	0.34	1.69	36.6	3.9	23	0.06	0.29	0.15	0.16	1.4

Table 4-4 continues on the next page

Notes: Major elements in wt. % and trace elements in ppm. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B are presented. Superscripts A or B follow each sample name identify the corresponding batch. The average composition of 10-LT-05 and 2 σ values of 25 analyses spanning 6 years are also provided.

Table 4-4. Continued.

Sample	Gd	Hf	Ho	La	Lu	Nb	Nd	Pr	Rb	Sm
<i>D.L</i>	<i>0.05</i>	<i>0.2</i>	<i>0.01</i>	<i>0.5</i>	<i>0.01</i>	<i>0.2</i>	<i>0.1</i>	<i>0.03</i>	<i>0.2</i>	<i>0.03</i>
BNB-16-028A	1.78	1.0	0.39	4.6	0.14	1.9	6.1	1.33	13.3	1.65
BNB-16-028B	1.93	1.3	0.43	5.3	0.15	2.1	6.9	1.54	21.2	1.75
BNB-13-048B	1.72	1.0	0.40	4.8	0.16	1.8	6.2	1.33	17.7	1.64
BNB-16-029	3.58	2.6	0.91	11.4	0.39	4.3	13.4	3.10	45.8	3.35
BNB-16-030A	3.11	1.8	0.65	9.5	0.26	3.5	11.6	2.63	46.8	3.03
BNB-16-030B	2.80	1.6	0.58	7.3	0.20	2.7	9.1	2.06	39.3	2.29
BNB-16-030C	3.42	2.5	0.83	11.1	0.30	4.5	13.5	3.05	83.5	3.75
BNB-16-030D	2.81	1.9	0.63	8.7	0.23	3.3	11.1	2.38	61.4	3.02
BNB-16-030E	10.05	7.1	2.03	27.4	0.75	12.5	37.0	8.18	20.7	9.65
BNB-16-030F	7.50	4.9	1.65	25.0	0.62	8.9	29.7	6.76	23.5	7.26
BNB-16-030G	5.62	3.4	1.14	16.8	0.43	6.5	20.1	4.61	58.5	5.16
BNB-16-032A	3.01	2.2	0.67	11.1	0.24	4.1	12.6	2.86	42.0	2.86
BNB-16-032B	3.13	2.0	0.66	10.4	0.26	4.3	12.5	2.90	32.3	3.12
BNB-16-031A	2.64	1.7	0.61	6.0	0.25	2.7	8.3	1.81	34.3	2.34
BNB-16-031B	2.63	1.5	0.57	5.9	0.22	2.5	8.2	1.77	45.3	2.41
BNB-16-031C	6.79	4.7	1.35	24.2	0.50	8.9	27.8	6.32	41.0	6.42
BNB-16-031D	4.12	2.8	0.90	12.6	0.33	4.2	15.1	3.45	56.0	3.60
BNB-16-031E	3.75	2.5	0.78	10.8	0.29	4.1	13.2	2.94	72.1	3.54
BNB-16-031F	2.54	1.6	0.57	6.1	0.25	2.4	8.1	1.82	48.8	2.49
BNB-16-033A	2.76	1.7	0.59	7.7	0.21	2.9	9.5	2.12	22.5	2.57
BNB-16-033B	2.85	1.7	0.58	8.1	0.24	2.8	9.6	2.17	19.5	2.71
BNB-16-033C	3.60	2.1	0.77	10.1	0.30	3.7	13.1	2.91	23.5	3.47
BNB-17-039A	4.54	3.4	1.30	5.1	0.53	6.1	9.7	1.99	23.2	2.97
BNB-17-039B	8.30	6.4	1.59	22.8	0.72	10.8	28.8	6.52	20.1	6.77
BNB-17-039C	8.01	9.6	1.20	37.8	0.52	5.8	35.3	8.93	19.0	8.66
BNB-17-039D	4.66	3.6	1.31	5.3	0.58	6.5	9.9	2.03	22.6	3.19
<i>Standards</i>										
10-LT-05	4.54	3.7	0.75	21.7	0.24	10.5	25.6	6.15	12.1	5.35
10-LT-05	4.61	3.5	0.75	23.0	0.25	11.5	26.7	6.76	13.0	4.82
10-LT-05- Average	4.45	3.6	0.73	22.4	0.25	10.9	25.6	6.21	12.5	5.15
2 σ	0.44	0.3	0.07	2.0	0.02	0.9	2.0	0.66	0.8	0.51

Table 4-4 continues on the next page

Notes: Major elements in wt. % and trace elements in ppm. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B are presented. Superscripts A or B follow each sample name identify the corresponding batch. The average composition of 10-LT-05 and 2 σ values of 25 analyses spanning 6 years are also provided.

Table 4-4. Continued

Sample	Sn	Sr	Ta	Tb	Th	Tm	U	V	W	Y	Yb	Zr
<i>D.L</i>	1	0.1	0.1	0.01	0.05	0	0.05	5	1	0.5	0.03	2
BNB-16-028A	<1	90.5	0.1	0.31	1.25	0.16	0.40	249	1	10.1	0.95	37
BNB-16-028B	<1	67.5	0.1	0.31	1.57	0.16	0.64	244	1	10	1.10	51
BNB-13-048B	<1	81.6	0.1	0.27	1.18	0.17	0.37	263	1	10	0.99	36
BNB-16-029	1	102.5	0.3	0.68	2.95	0.4	0.92	392	<1	23.6	2.38	90
BNB-16-030A	1	169.0	0.2	0.53	2.09	0.25	0.63	260	1	16.9	1.70	68
BNB-16-030B	1	152.5	0.2	0.44	1.91	0.22	0.55	237	<1	14.4	1.40	61
BNB-16-030C	1	187.0	0.3	0.68	2.77	0.31	0.86	365	1	20.9	1.75	91
BNB-16-030D	2	188.5	0.2	0.51	2.1	0.25	0.67	400	1	16.7	1.59	70
BNB-16-030E	1	72.4	0.9	1.69	8.39	0.74	2.65	<5	1	52.9	4.69	262
BNB-16-030F	2	137.0	0.7	1.33	6.19	0.63	2.00	57	1	41.2	3.82	183
BNB-16-030G	1	150.0	0.4	0.97	4.35	0.45	1.32	484	1	29.4	2.65	128
BNB-16-032A	1	569.0	0.3	0.53	2.8	0.29	0.90	235	1	17	1.63	79
BNB-16-032B	1	270.0	0.3	0.55	2.89	0.25	1.00	233	<1	17.1	1.70	80
BNB-16-031A	<1	170.0	0.1	0.47	1.52	0.25	0.48	256	<1	15.5	1.53	57
BNB-16-031B	<1	157.5	0.2	0.48	1.44	0.22	0.47	249	<1	14.6	1.33	55
BNB-16-031C	2	142.5	0.6	1.08	6.61	0.52	1.99	73	1	34.6	3.42	174
BNB-16-031D	1	146.0	0.4	0.67	3.06	0.33	0.88	509	1	22.1	2.14	91
BNB-16-031E	1	129.5	0.3	0.62	2.84	0.33	0.85	555	1	20.4	1.97	90
BNB-16-031F	1	185.5	0.2	0.42	1.58	0.25	0.45	255	<1	15.1	1.56	56
BNB-16-033A	1	215.0	0.2	0.46	1.71	0.23	0.51	270	1	15.5	1.51	60
BNB-16-033B	1	193.5	0.3	0.46	2.02	0.25	0.57	255	1	15.5	1.63	61
BNB-16-033C	1	191.0	0.3	0.56	2.31	0.34	0.77	195	<1	19.9	2.05	76
BNB-17-039A	2	125.5	0.4	0.82	1.98	0.55	0.73	350	<1	36.2	3.60	126
BNB-17-039B	2	183.5	0.7	1.31	5.42	0.68	1.70	186	1	43.4	4.28	268
BNB-17-039C	1	174.0	1	1.21	12.5	0.5	3.21	104	<1	31.4	3.14	340
BNB-17-039D	2	134.5	0.4	0.82	2.18	0.57	0.73	362	<1	36.7	3.48	133
<i>Standards</i>												
10-LT-05	1	894.0	0.6	0.66	1.98	0.27	0.69	159	1	18.6	1.56	151
10-LT-05	2	948.0	0.6	0.68	2.02	0.29	0.57	173	<1	19.5	1.72	155
10-LT-05- Average	1	938.2	0.6	0.66	1.99	0.27	0.59	164	1	18.6	1.66	150
2σ	1	38.5	0.1	0.06	0.22	0.04	0.10	22	0	1.5	0.21	14

Table 4-4 continues on the next page

Notes: Major elements in wt. % and trace elements in ppm. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B are presented. Superscripts A or B follow each sample name identify the corresponding batch. The average composition of 10-LT-05 and 2σ values of 25 analyses spanning 6 years are also provided.

Table 4-4. Continued

Sample	Ag	As	Cd	Co	Cu	Li	Mo	Ni	Pb	Sc	Tl	Zn
<i>D.L</i>	<i>0.5</i>	<i>5</i>	<i>0.5</i>	<i>1</i>	<i>1</i>	<i>10</i>	<i>1</i>	<i>1</i>	<i>2</i>	<i>1</i>	<i>10</i>	<i>2</i>
BNB-16-028A	<0.5	<5	<0.5	58	51	10	1	328	6	41	<10	80
BNB-16-028B	<0.5	7	<0.5	50	38	10	<1	329	2	40	<10	78
BNB-13-048B	<0.5	<5	0.5	58	61	10	1	314	5	42	<10	81
BNB-16-029	<0.5	<5	<0.5	42	78	20	<1	36	18	40	10	95
BNB-16-030A	<0.5	6	<0.5	46	156	20	<1	108	8	34	<10	114
BNB-16-030B	<0.5	6	<0.5	43	165	20	<1	90	7	36	<10	129
BNB-16-030C	<0.5	8	0.5	46	177	20	<1	45	5	33	<10	111
BNB-16-030D	<0.5	6	<0.5	49	187	20	1	68	5	34	<10	120
BNB-16-030E	<0.5	<5	<0.5	16	41	10	1	<1	7	20	<10	52
BNB-16-030F	<0.5	<5	<0.5	39	42	10	1	1	6	31	<10	60
BNB-16-030G	<0.5	<5	0.5	52	95	20	1	11	8	34	<10	122
BNB-16-032A	<0.5	27	0.6	42	137	30	<1	99	7	32	<10	129
BNB-16-032B	<0.5	9	0.8	42	103	30	1	106	57	30	<10	170
BNB-16-031A	<0.5	6	<0.5	49	140	30	1	131	4	35	10	86
BNB-16-031B	<0.5	<5	0.5	50	129	30	1	158	3	35	<10	89
BNB-16-031C	<0.5	5	0.6	34	62	20	1	11	20	28	<10	182
BNB-16-031D	<0.5	13	0.8	54	177	20	<1	34	11	38	<10	112
BNB-16-031E	<0.5	9	0.6	56	188	20	1	36	7	37	<10	134
BNB-16-031F	<0.5	6	<0.5	43	129	20	1	95	8	33	10	97
BNB-16-033A	<0.5	<5	0.7	47	172	20	1	87	4	36	<10	89
BNB-16-033B	<0.5	<5	0.6	45	150	20	<1	65	12	34	<10	115
BNB-16-033C	<0.5	<5	<0.5	43	87	20	1	69	2	33	<10	90
BNB-17-039A	<0.5	<5	<0.5	42	106	20	<1	113	<2	42	<10	86
BNB-17-039B	<0.5	<5	<0.5	33	99	10	<1	44	4	30	<10	81
BNB-17-039C	<0.5	<5	<0.5	16	26	10	<1	23	7	23	<10	39
BNB-17-039D	<0.5	5	<0.5	43	146	10	1	110	<2	44	10	88
<i>Standards</i>												
10-LT-05	<0.5	<5	0.6	38	42	10	2	114	6	18	<10	94
10-LT-05	<0.5	<5	<0.5	42	43	10	1	124	4	18	<10	100
10-LT-05- Average		7	0.6	39	42	10	1	115	4	17	10	92
2σ		1	0.1	4	4	0	1	9	3	2	0	7

Notes: Major elements in wt. % and trace elements in ppm. Sample 10-LT-05 is an internal basalt standard used at Carleton University; analytical results for batches A and B are presented. Superscripts A or B follow each sample name identify the corresponding batch. The average composition of 10-LT-05 and 2σ values of 25 analyses spanning 6 years are also provided.

Nearly all dykes and sills classify as basalt on a Zr/Ti versus Nb/Y plot (**Figure 4-6b**), with only a few exceptions plotting in the andesite/basaltic andesite field. Most samples have Ti/V ratios between 20 and 50, plotting within the continental basalt field of Shervais (1982), although three Nipissing sills, one Klotz dyke, and one Maguire dyke plot within the ocean island basalt (OIB) field, with Ti/V ratios over 50 (**Figure 4-6c**). Conversely, there are other Nipissing sills and some Maguire and Senneterre dykes that have Ti/V ratios below 20.

Figure 4-6. Major and trace element classification of dykes and sills belonging to the filtered geochemical subset of the Ungava large igneous province. a) Total alkali vs. silica plot (Le Bas *et al.*, 1986); b) Zr/Ti vs. Nb/Y classification plot for basalt (Pearce and Norry, 1979); and c) V vs. Ti (ppm)/1000 (Shervais, 1982). Data sources as in **Figure 4-5**.

Five distinct multi-element patterns are recognized within the *ca.* 2.2 Ga dykes and sills (**Figure 4-7**). These patterns correspond to the geochemical groups defined by Maurice and Francis (2010), including the high Fe-Ti (low and high Al subgroups), low Fe-Ti and low REE groups.

The high Fe-Ti group is characterized by Ti/V ratios greater than 15 and Nb/Yb_{PM} ratios greater than 1.3, steep LREE slopes (La/Sm_{PM} ratios between 1.5 and 3; normalized to primitive mantle from Sun and McDonough, 1989), a negative Nb anomaly, and Th enrichment (**Figure 4-6** and **Figure 4-7a** and **b**). Two subgroups within the high Fe-Ti group are distinguished by their HREE slopes (Tb/Yb_{PM} ratios). These directly correspond to the low and high Al subgroups defined by Maurice and Francis (2010). With steep HREE slopes (Tb/Yb_{PM} ratios >1.5), the low Al subgroup includes all dykes from the Anuc and Kogaluk Bay swarms as well as members of the Maguire and Senneterre swarms. The high Al subgroup, in contrast, has nearly flat HREE slopes (Tb/Yb_{PM} ratios <1.5). These include two unclassified dykes from Maurice and Francis (2010) as well as members of the Nipissing sills from the Otter, Joubin and Boon townships near Elliot Lake.

The low Fe-Ti group includes the majority of Nipissing sills, and Senneterre and Maguire dykes (70% of the filtered data set; **Figure 4-7c**). Multi-element patterns of the low Fe-Ti group include pronounced negative Nb anomalies, enrichment in Th with Nb/Yb_{PM} ratios of approximately 1, steep LREE slopes (La/Sm_{PM} 1.5–4), and flat HREE slopes. Sills belonging to this group tend to have more pronounced negative Nb anomalies, greater enrichment in Th, and slightly higher La/Sm_{PM} ratios compared to dykes.

Figure 4-7. Multi-element plots of the geochemical groups identified among the dykes and sills belonging to the Ungava LIP. Trace element concentrations are normalized to primitive mantle values of Sun and McDonough (1989). Where available, $\epsilon_{Nd_{T=2.2Ga}}$ are matched to corresponding samples. a) High Fe-Ti (low aluminum subgroup); b) high Fe-Ti (high aluminum subgroup); c) low Fe-Ti group; d) low REE (-ΔNb subgroup); and e) low REE (+ΔNb subgroup). Trace element data sources as listed in **Figure 4-5**. Isotopic data are from this study, and compiled from Lightfoot and Naldrett (1996b) and Maurice and Francis (2010).

With nearly flat LREE and HREE slopes on a multi-element plot, the low REE group has the most MORB-like pattern of all dykes and sills that are part of the Ungava LIP (**Figure 4-7d** and **e**). Modified from Maurice and Francis (2010), the low REE group is divided into two subgroups by Th and Nb abundance. The low REE ($-\Delta\text{Nb}$) subgroup is slightly enriched in Th and has a negative Nb anomaly (**Figure 4-7d**). This subgroup includes members of the Klotz dykes and Nipissing sills. The low REE ($+\Delta\text{Nb}$) subgroup is distinguished from all other groups as it lacks both a negative Nb anomaly and Th enrichment (ΔNb as calculated by Fitton *et al.*, 1997; **Figure 4-7e**). All of the Couture and some of the Klotz belong to this low REE ($+\Delta\text{Nb}$) subgroup.

4.4.4 Isotope geochemistry

Sm-Nd isotope analyses were completed for two Maguire dykes, one Senneterre and one Klotz dyke, as well as two samples from the Triangle Mountain sill (**Table 4-5**). The two Maguire dykes have distinct $\varepsilon\text{Nd}_{T=2229\text{Ma}}$ values at +0.4 and -3.3. The Senneterre dyke has an $\varepsilon\text{Nd}_{T=2214\text{Ma}}$ value at -2.1, matching one of the Maguire dykes. The Klotz dyke has an $\varepsilon\text{Nd}_{T=2209\text{Ma}}$ of +3.1, which is distinct from both Maguire and Senneterre dyke sample (**Figure 4-8**). The samples analyzed from the Triangle Mountain sill include a fine-grained gabbro (52.4 wt. % SiO_2 and $\text{Mg\#} = 59$) located near the basal contact of the sill, and a course-grained pegmatoidal varitextured gabbro located near the top of the sill (52.9 wt. % SiO_2 and $\text{Mg\#} = 35$). The samples produced $\varepsilon\text{Nd}_{T=2214\text{Ma}}$ values of -0.4 and -1.8, respectively, overlapping within analytical error (error margin of ± 0.5 εNd units) with the Senneterre dyke and one of the Maguire dykes.

Table 4-5. Sm-Nd isotope data for Triangle Mountain (Nipissing sill), Maguire, Senneterre, and Klotz dykes.

Sample	Locality	Long. (W)	Lat. (N)	Geochemical group	Nd (ppm)	Sm (ppm)	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}_i$	$\epsilon\text{Nd}_{(i)}$
BNB-16-031B	Triangle Mtn.	79.7800	47.4946	Low Fe-Ti	9.74	2.72	0.512208	0.1687	0.509748	-0.4
BNB-16-031D	Triangle Mtn.	79.7942	47.4903	Low Fe-Ti	15.66	3.94	0.511895	0.1522	0.509675	-1.8
BXA91-7605	Maguire	75.044	57.212	High Fe-Ti, low Al	22.81	5.79	0.512024	0.1535	0.50977	0.4
BXA91-7101	Maguire	72.540	58.090	Low Fe-Ti	8.23	1.87	0.511600	0.1377	0.509579	-3.3
TM34	Senneterre	81.794	47.665	Low Fe-Ti	8.91	2.35	0.511984	0.1593	0.509659	-2.1
BXA91-7302	Klotz	73.910	60.582	Low REE (+ ΔNb)	11.99	3.6	0.512580	0.1817	0.509935	3.1

Notes: Coordinates in WGS1984.

$^{143}\text{Nd}/^{144}\text{Nd}_m$ is the measured present-day value, $^{143}\text{Nd}/^{144}\text{Nd}_i$ is the initial ratio and ϵNd_i based on crystallization ages of 2216 Ma for the Triangle Mountain sill, 2229 Ma for Maguire, 2214 Ma for Senneterre, and 2209 Ma for the Klotz dyke.

Isotope ratios are normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$.

2 σ uncertainties of $^{143}\text{Nd}/^{144}\text{Nd}_m$ are 0.000011 based on reproducibility of the in-house Nd Standard.

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

Figure 4-8. $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ versus SiO_2 wt.% with chondrite uniform reservoir (CHUR) reference line. Sources as in Figure 4-7.

4.5 Discussion

Magmatism of the Ungava LIP can be broadly categorized into three pulses: early (2229–2220 Ma), middle (2217–2214 Ma), and late (2209–2199 Ma) (**Table 4-1**). The early pulse is affiliated with swarms of limited areal extent (Anuc and possibly some members of the Maguire dykes) and represents volumetrically minor magmatism. The middle pulse includes most of the Maguire and Senneterre dykes, which define the radiating pattern of the Ungava LIP, as well as the local Kogaluk Bay dykes. Nipissing sills are also part of the middle pulse, which is confirmed by the new crystallization age of the Triangle Mountain sill at 2216.5 ± 2.1 Ma. Together, the Maguire, Senneterre, and Kogaluk Bay dykes and the Nipissing sills represent the most significant phase of the Ungava LIP volumetrically. The Klotz and Couture dykes belong to the late pulse. Though the Couture dykes are of limited area, the Klotz dykes are geographically more extensive and possibly expand the radiating pattern of the Ungava LIP to the north.

4.5.1 Feeders to the Nipissing sills

Three geochemical signatures are recognized within the Nipissing sills. These include the high Fe-Ti (high Al), low Fe-Ti and low REE ($-\Delta\text{Nb}$) groups (see **Figure 4-7b, c, and d**), each of which match different dyke swarms located within the Ungava Peninsula.

The Otter, Joubin, and Boon townships are host to the Nipissing sills with high Fe-Ti (high Al subgroup) geochemistry (5 of 89 samples). Unlike the low Fe-Ti and low REE ($-\Delta\text{Nb}$) groups, the high Fe-Ti (high Al) subgroup does not have exact matches to dykes of the Ungava LIP. Instead, these sills match geochemistry of dykes not yet assigned to a particular swarm.

The majority of Nipissing sills (76 of 89 samples) belong to the low Fe-Ti geochemical group, which also includes most of Maguire (7 of 9 samples) and Senneterre (12 of 14 samples) dykes. Matches in whole-rock major and trace element patterns (**Figure 4-7c**), in addition to Nd isotopic ratios (**Figure 4-8**), support their genetic relationship as proposed by (Buchan *et al.*, 1998).

Although less abundant, other geochemical signatures recognized within the Nipissing sills further support a link to the dykes of the Ungava LIP. Nipissing sills located within May, Baldwin, and Otter townships, for example, are less enriched in LREEs compared to those belonging to the low Fe-Ti group. These sills (8 of 89 samples) are a closer match to the Klotz dykes of the low REE ($-\Delta\text{Nb}$) subgroup.

4.5.2 A preliminary tectono-magmatic interpretation

At least three distinct mantle end members are recognized as sources of Ungava LIP magmas feeding dykes and sills. These include the high Fe-Ti (low and high Al) and low REE groups. The high Fe-Ti (low Al) group has $\epsilon\text{Nd}_{(t=2.2\text{ Ga})}$ values between -0.2 and +0.4 (**Figure 4-8**; Maurice and Francis, 2010). Because these dykes are enriched in LREE, have high Tb/Yb_{PM} ratios, and have a negative Nb anomaly, deeper depths of melting can be inferred but their initial isotopic mantle-source signatures are obscured by the incorporation of a crustal component. Conversely, major and trace element patterns (e.g., flat HREEs, and negative correlation between Al₂O₃ and Mg#, as described by Maurice and Francis, 2010) indicate that magmas of the high Fe-Ti (high Al) subgroup were generated at shallower depths and from a distinct mantle source compared to the high Fe-Ti (low Al) subgroup.

The low REE (+ Δ Nb) subgroup has $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values between +3.1 and +4.1 (**Figure 4-8**; Maurice and Francis, 2010; this study). Paired with the nearly flat trace element patterns, their geochemical signature is characteristic of magmas produced by higher degrees of partial melting of upper mantle at shallow depths. The low REE (- Δ Nb) subgroup is considered to have been derived from the low REE (+ Δ Nb) subgroup but was contaminated prior to final emplacement. Indicators of crustal contamination include a negative Nb anomaly, subtle Th and LREE enrichments, and a single $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ value of +1.4 (Maurice and Francis, 2010).

In similarity to the low REE group, flat HREE patterns of the low Fe-Ti group are indicative of shallow melting. A crustal contamination signature suggested by enrichment in LREEs and Th, and prominent negative Nb anomalies are accompanied by $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values of -2.1 and -3.3 in the Senneterre and Maguire dykes. Assimilation-fractional crystallization (AFC) processes continued following emplacement of the Nipissing sills, indicated by progressively lower $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ with increasing SiO_2 wt. % (**Figure 4-8**). Note that the degree of AFC for each individual Nipissing sill is slightly different given that the processes progressed in situ. For example, the Triangle Mountain sill shows very little change from the most primitive to the more evolved sections in terms of $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values, which range from -0.4 to -0.8. In contrast, the Kerns sill shows a wider range of $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values from -2.6 to -5.6. Whether the low Fe-Ti group was generated from a distinct source or is a heavily contaminated phase derived from the low REE group requires further investigation.

Figure 4-9. Discrimination plots for metallogenic potential, after Jowitt and Ernst (2013). a) Ni vs. MgO wt.%; dashed line marks the within plate basalt trend from Keays and Lightfoot (2007); and b) Cu/Zr_{PM} vs. Th/Yb_{PM}. References as in **Figure 4-5**. Normalization values for Th, Yb, and Zr from Sun and McDonough (1989), and normalization values for Cu from McDonough and Sun (1995).

4.5.3 Ni-Cu-PGE potential of the Nipissing sills

Jowitt and Ernst (2013) characterize LIPs with known Ni-Cu-PGE mineralization as having a basaltic composition with Ti/V ratios of less than 50, HREEs similar to primitive mantle, and enrichment in LREEs (monitored by La/Sm_{PM} and Th/Yb_{PM} ratios greater than 1). A magmatic system with these characteristics, however, does not explicitly indicate a fertile system. The magmatic system must also have been generated from a source undepleted in chalcophile elements and have undergone S-saturation (Jowitt and Ernst, 2013). The Ni versus MgO wt. % plot can be used to evaluate S-saturation (**Figure 4-9a**; Jowitt and Ernst, 2013). Samples with Ni concentrations less than the within-plate basalt

array are inferred to have lost Ni to an immiscible sulphide melt; a deficit that occurs by early S-saturation (Jowitt and Ernst, 2013). With the exception of the samples from the mineralized Shakespeare deposit and two samples from the high Fe-Ti (low Al) group, all samples plot either on or below the within-plate basalt array, suggestive of an early sulphur saturation event.

Ratios between chalcophile and lithophile elements of comparable incompatibilities offer a means of evaluating chalcophile element concentrations of a magma without influence from fractional crystallization processes (Jowitt and Ernst, 2013). **Figure 4-9b** compares ratios of $\text{Cu}/\text{Zr}_{\text{PM}}$ against the crustal contamination proxy $\text{Th}/\text{Yb}_{\text{PM}}$. A magmatic group is considered prospective for Ni-Cu-PGEs when it includes signatures that are both undepleted (fertile) and depleted (loss of chalcophile elements to an immiscible sulphide magma; Jowitt and Ernst, 2013). Prospective geochemical groups include the low Fe-Ti and low REE groups. The low Fe-Ti group, which includes the Shakespeare deposit and most Nipissing sills, shows the widest variation in $\text{Cu}/\text{Zr}_{\text{PM}}$ and high $\text{Th}/\text{Yb}_{\text{PM}}$ values, consistent with high degrees of crustal assimilation (**Figure 4-9b**) and local sulphur saturation. The low REE group with $+\Delta\text{Nb}$ has $\text{Cu}/\text{Zr}_{\text{PM}}$ values above primitive mantle, indicates that these are slightly enriched in chalcophile elements. Nipissing sills belonging to the low REE group consistently have $-\Delta\text{Nb}$ signatures that follow a $\text{Cu}/\text{Zr}_{\text{PM}}$ depletion trend with $\text{Th}/\text{Yb}_{\text{PM}}$ and for this reason, may be prospective.

Samples belonging to the high Fe-Ti (low Al and high Al) group are considered less prospective. Although the high Fe-Ti (low Al) group includes enriched and depleted chalcophile signatures, their steeper HREE slopes are uncharacteristic of mineralized intrusions. The high Fe-Ti (high Al) subgroup is not considered prospective because all

samples (sills and dykes) are depleted in chalcophile elements.

4.6 Implications for the Superia supercraton paleogeographic reconstruction

The Superia supercraton paleogeographic reconstruction proposed by Bleeker and Ernst (2006) predicted that the Nipissing sills directly fed the Karjalitic sills located in the Karelia-Kola craton (Vuollo and Phrainen, 1992; Vuollo and Huhma, 2005). Sm-Nd isotope analyses by Hanski *et al.* (2010) report whole-rock $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values ranging from +2.1 to -0.5 for the Karjalitic sills. Based on this evaluation of nine mafic sills and dykes, Hanski *et al.* (2010) found that the Karjalitic sill population is isotopically homogeneous and involved only minor in situ crustal contamination. The preliminary results from this study indicate that the Maguire and Senneterre dykes of the Low Fe-Ti group (the geochemically compatible feeder dykes to the Nipissing sills) have lower $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values (-3.3 and -2.1, respectively) compared to the Karjalitic sills analyzed by Hanski *et al.* (2010). If the Karjalitic sills were the continuation of the Nipissing sills then their $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values should either be equivalent or less than the Nipissing sills (with lower values associated with progressive crustal assimilation during emplacement). Therefore, the Karjalitic sills are not the direct continuation of the Nipissing sills (see also Davey *et al.*, 2016). Instead, the Karjalitic sills may have been fed from a separate branch of dykes from the Ungava LIP. For example, a separate branch may correspond to the high Fe-Ti (low Al) subgroup which includes $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values between +0.4 and -0.2; values that are comparable to the Karjalitic sills. Further geochemical investigations, however, are required to confirm their compatibility. Additionally, a regional significance of the high Fe-Ti (low Al) dykes is unclear given their limited extent in the Ungava Peninsula.

The overall finding that the Karjalitic sills are not the ‘downstream’ equivalent of the Nipissing sills has implications for the Superia supercraton paleogeographic reconstruction. The tight fit between the Kola Peninsula (Karelia-Kola craton) and the southeastern margin of the Superior craton proposed by Bleeker and Ernst (2006) can be relaxed to allow for more flexibility in the positioning of the Karelia-Kola within the context of the Superia supercraton paleogeographic reconstruction.

4.7 Summary

The geochemistry of dykes and sills belonging to the *ca.* 2.2 Ga Ungava LIP are compiled to gain a better understanding of the overall magmatic system and Ni-Cu-PGE prospectivity. Geochemical groups described by Maurice and Francis (2010), including the high Fe-Ti (low and high Al), low Fe-Ti and low REE groups, are recognized within the dykes and sills belonging to the Ungava LIP.

The high Fe-Ti (low Al) group includes representatives of the Anuc, Kogaluk Bay, and some Maguire and Senneterre dykes. No Nipissing sill matches this geochemical signature. Trace element characteristics indicate that the group was generated from deeper depths (compared to the low REE and low Fe-Ti groups) and incorporated a crustal component that may have influenced the $+0.4 \text{ } \epsilon\text{Nd}_{T=2.2\text{Ga}}$ signature.

The high Fe-Ti (high Al) group has geochemical characteristics of melting at shallow depths but from a more enriched source compared to the low Fe-Ti and low REE groups. Ubiquitous depletion of chalcophile elements indicates that the Nipissing sills belonging to this group are not prospective for Ni-Cu-PGE. No dykes with U-Pb ages dated

at 2.2 Ga belong to this geochemical group.

Low Fe-Ti chemistry is the most abundant geochemical group and includes the majority of samples belonging to the Senneterre and Maguire dykes and Nipissing sills. Heavy rare earth element patterns indicate that these magmas were generated at shallow mantle depths and LREE enrichment is ascribed to crustal contamination. Samarium-neodymium isotopes indicate that the Maguire and Senneterre dykes feeding the Nipissing sills incorporated a crustal component prior to emplacement as evidenced by low $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values at -2.1 and -3.3. The mineralized Shakespeare deposit belongs to this geochemical group and the group as a whole is considered prospective.

A new crystallization age for the Triangle Mountain sill, part of the low Fe-Ti group, is determined to be 2216.5 ± 2.1 Ma by U-Pb thermal ionization mass spectrometry (TIMS) of baddeleyite. Values of $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ of between -0.4 and -1.8 for the Triangle Mountain sill indicate that in situ crustal assimilation occurred but likely not to the same degree as other sills (e.g., Obabika and Kerns sills: Lightfoot and Naldrett, 1996a).

The low REE group is considered to have been generated by melting of the upper mantle at shallow depths. The group is divided into subgroups based on Th and Nb abundance. The $+\Delta\text{Nb}$ subgroup lacks indicators of crustal contamination in terms of trace element patterns (absence of a negative Nb anomaly, and no Th enrichment). Samarium-neodymium isotopes with $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values between +3.1 and +4.1 further support sourcing from uncontaminated upper mantle. This sub-group includes the Couture and Klotz dykes. The $-\Delta\text{Nb}$ sub-group, on the other hand, is distinguished by subtle enrichment in Th and LREEs as well as a negative Nb anomaly. The single $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ value of +1.4 is indicative of input from a crustal component, especially when paired with its multi-element

signature. The $-\Delta\text{Nb}$ subgroup includes members of the Klotz dykes and some Nipissing sills, and is considered prospective for Ni-Cu-PGE.

With respect to the paleogeographic reconstruction of the Superia supercraton by Ernst and Bleeker (2010), we find that the Nipissing sills did not directly feed the Karjalitic sills in the Karelia-Kola craton. The close fit between the Superior and Karelia-Kola craton previously restricted by this correlation can be relaxed which will allow for more flexibility in the positioning of the Karelia-Kola craton within the paleogeographic reconstruction.

4.8 Next steps

The analysis of a second batch of six samples for Sm-Nd isotopes is in progress, which will expand the isotope patterns identified in the Senneterre, Maguire, and Klotz dykes. These new data combined with the existing database will help determine a more detailed model of the evolution of the magmatic system to further constrain the crustal components recognized in the various geochemical groups, and whether the low REE and low Fe-Ti groups are related within a single evolving magmatic system. Furthermore, the spatial distribution of the geochemical groups and their age relationships will also be evaluated.

Future geochronology is suggested for the Nipissing sills belonging to the high Fe-Ti (high Al) and low REE geochemical groups. Precise ages for these specific geochemical groups can test their association with the Ungava event and help to constrain the overall secular evolution of the magmatic system (for example, do the low REE ($-\Delta\text{Nb}$) Nipissing sills match the age of the Klotz dykes?).

Chapter 5

Evidence for a single large igneous province at ca. 2.11 Ga across supercraton Superia

5.1 Abstract

The Superia paleogeographic reconstruction is largely based on the shared large igneous province (LIP) record to join the Superior, Hearne, Wyoming and Karelia-Kola Archean cratons from >2.5-2.0 Ga in a supercraton. The methodology relies on matching U-Pb geochronology (i.e., magmatic barcodes), paleolatitudes from paleomagnetic studies, and radiating geometry of mafic dyke swarms belonging to LIPs. If contemporaneous dyke swarms and other LIP units from these Archean cratons are indeed fragments of the same LIP, then each matching dyke swarm should have common geochemical signatures diagnostic of a shared petrogenesis.

Confirming contemporaneity to the Marathon dykes of the Superior craton, a new U-Pb age at 2114.4 ± 3 Ma from chemically abraded zircons analyzed by thermal ionization mass spectrometry is determined for the Nieminen dyke, Western Karelia. To test for a shared petrogenetic history, a geochemical analysis is presented on *ca.* 2.11 Ga mafic magmatism from the Marathon and Kapuskasing dykes (i.e., Marathon LIP, Superior craton), Kazan dykes and Griffin sills (Hearne craton), Bear Mountain dykes (Wyoming craton) and Tohmajärvi-Pirtguba dykes, Misi sills, and Jouttiaapa and Hirsimaa mafic volcanic rocks (all from the Karelia-Kola craton).

The 2.11 Ga mafic units can be geochemically related by bulk mixing of a mantle plume derived melt, represented by Griffin sills, with enriched upper mantle melts characterized by a subset of the Tohmajärvi-Pirtguba dykes. These end-members and their mixtures then underwent variable degrees of assimilation and fractional crystallization (AFC) as they ponded at the interface between crust and mantle prior to final emplacement.

5.2 Introduction

Supercraton Superia is a paleogeographic reconstruction of at least one of Earth's earliest large ancestral landmasses that involves reworked Archean crustal fragments from Canada's Superior and Hearne cratons, as well as the Wyoming (United States of America) and Karelia-Kola (Finland and Russia) cratons (Bleeker and Ernst, 2006; Ernst and Bleeker, 2010). Similar geological features, such as comparable assembly histories, Precambrian basement age and rock type, supracrustal assemblages, and mafic magmatism have long been recognized and discussed in the literature for these cratons (Williams *et al.*, 1991; Heaman, 1997; Bickford *et al.*, 2007; Chamberlain and Mueller, 2007). Key to the Superia supercraton paleogeographic reconstruction is the recognition that multiple pre-breakup Paleoproterozoic large igneous provinces (LIPs) from 2.5 to 2.0 Ga, including their feeder systems (dykes and sills), and rarely preserved flood basalt components match in time across these cratons (Bleeker and Ernst, 2006).

While a single age match may simply be a coincidence, the chance that multiple LIPs correlate through time is less likely random and probably more indicative of a shared history. LIPs and especially their plumbing system components (particularly regional

mafic dyke swarms) provide temporal and spatial constraints used to test paleogeographic reconstructions. Conventional approaches include paleomagnetic, geochronological and dyke swarm geometry studies. For example, the paleomagnetic study by Kilian *et al.* (2016b) shows that Wyoming and Superior were closely linked until at least 2.15 Ga and likely until 2.0 Ga. Paleomagnetic details constraining paleolatitude and paleorientation of the Karelia-Kola and Hearne cratons, however, are not well constrained. There are no key poles available for the Hearne craton between 2.5-2.0 Ga (Buchan *et al.*, 2000). In the Karelia-Kola craton, the D versus D' paleomagnetic poles measured from 2.45 Ga mafic intrusions by Mertanen *et al.* (1999) are still debated as to which is representative of the primary magmatic signal (Pesonen *et al.*, 2012). For this reason, the paleogeographic reconstruction presented in **Chapter 2** proposes a position for the Karelia-Kola craton south of the eastern Superior craton and southeast of the Wyoming craton (**Figure 5-1**) – a best fit that is a compromise between *ca.* 2.4-2.0 Ga U-Pb geochronology, dyke trends and existing paleomagnetic data (Mertanen *et al.*, 1999; Bleeker and Ernst, 2006; Pisarevsky *et al.*, 2015; Salminen *et al.*, 2019). For reference, this paleogeographic reconstruction keeps the eastern Superior craton in its present position against which the Hearne, Karelia-Kola and Wyoming cratons are repositioned into their proposed locations within the reconstruction. Constraints on the positioning of the pervasively deformed Hearne craton are limited. With few paleomagnetic studies showing similar latitudes as the Superior and Wyoming cratons, correlation between the Hearne and Superior cratons is primarily based on matching magmatic barcode records (Ernst and Bleeker, 2010).

The most significant age matches used for the Superia supercraton reconstruction are at 2.50, 2.45, 2.40, 2.22, 2.11, and 2.00 Ga, with many of the corresponding magmatic

events providing the key poles and geometric constraints used for the previous reconstructions (e.g., Mertanen *et al.*, 1999; Bleeker and Ernst, 2006; Halls *et al.*, 2008; Salminen *et al.*, 2019). If these contemporaneous LIPs were produced by the same magmatic event then their counterparts from each craton should have comparable geochemistry, which may or may not include more than one geochemical signature.

The purpose of this study is to test the Superia supercraton paleogeographic reconstruction using geochronology and geochemistry of *ca.* 2.11 Ga mafic dykes and sills — a time representing some of the most prolific mafic magmatism, including multiple pulses, on the various crustal fragments of Superior, Hearne, Wyoming and Karelia-Kola cratons that make up Superia (**Figure 5-1**). To further expand upon the geochronological record in Karelia-Kola, we report a U-Pb chemical abrasion isotope dilution thermal ionization mass spectrometry (CA-ID-TIMS) age of 2114.4 ± 3 Ma for the Nieminen dyke from Western Karelia. A geochemical comparison of whole-rock major, trace element and Nd isotope data are compiled from the literature and new data are presented for the contemporaneous Marathon and Kapuskasing dykes (Superior), Kazan dykes and Griffin sills (Hearne), Bear Mountain dykes (Wyoming), and Tohmajärvi-Pirtguba dykes, Misi gabbro sills, and Jouttiaapa and Hirsimaa mafic volcanic rocks (all from Karelia-Kola). The geochemical analysis identifies multiple signatures within dyke swarms on individual cratons (in which ages are confirmed by U-Pb geochronology; see also Halls *et al.*, 2008) that match to dykes and sills in the other cratons. These chemical variations are explained by mantle plume melts that mixed with asthenosphere and then underwent variable degrees of assimilation and fractional crystallization while ponding at the base of the crust prior to final emplacement.

Geological features

- 2100-2135 Ma dykes and sills
- ca. 2215 Ma Nipissing sills
- ca. 2215 Ma Karjalitic sills
- Other Paleoproterozoic swarms
- Paleoproterozoic cover sequences

Geochemical groups

- ▲ Group 1A
- Group 1B
- ◇ Group 2
- ◇ Group 3
- Dated samples without geochemistry

Figure 5-1. *On the previous page.* Superia supercraton paleogeographic reconstruction from the latest Archean to Paleoproterozoic breakup modified after Bleeker and Ernst (2006), Ernst and Bleeker (2010), and **Chapter 2**. Dykes and sills represented by blue lines and shapes are those dated at *ca.* 2100-2135 Ma. Locations of dated samples with their geochemical group are shown for each craton with further age details and references as in **Table 5-1**. The age of the Nieminen dyke, Western Karelia (dated in this study) is in bold text. Western Superior is rotated 14° clockwise with respect to eastern Superior (Evans and Halls, 2010). North arrows on each craton show present day orientations. Archean fragment sizes are minimum estimates indicated by pink and purple shapes with dashed outlines. The full extent of supercraton Superia is unknown but it is represented in light grey. Superior dyke swarms are from Buchan and Ernst (2004); Superior craton modified from Goodfellow (2007) and Horton *et al.* (2017). Karelia-Kola basement geology and dyke swarms modified from Amelin *et al.* (1995); Mertanen *et al.* (1999); GTK (2012); Lubnina *et al.* (2016); Huhma *et al.* (2018). Wyoming dyke swarms and basement geology modified from Kilian (2016a) and Horton *et al.* (2017). Inset: Top graph shows the probability distribution for ages between 2100 and 2135 Ma from each craton individually; bottom graph shows the probability distribution and histogram for all ages between 2100 and 2135 Ma together. Probability distribution plots and histogram calculated using software of Ludwig (2003).

5.3 Geological Setting

5.3.1 Superior craton

Built on the 3.0 Ga North Caribou superterrane cratonic nucleus, the Superior craton is an amalgamation of independently grown Mesoarchean microcontinents and Neoarchean oceanic crust that accreted during the Neoarchean (Percival, 2007a, 2007b). After *ca.* 2.6 Ga cratonization, the Archean crust of the Superior craton was intruded by a number of regionally extensive Paleoproterozoic mafic dyke swarms which ultimately resulted in rifting events leading to the breakup of supercraton Superia (Bleeker, 2003). Of the 2.50 to 2.00 Ga dyke swarm record in the Superior craton, the 2.11 Ga Marathon dyke swarm is associated with the initiation of rifting at the southern margin of the Superior craton, and its successor, the 2.08 Ga Fort Frances swarm farther west, is considered the continuation of magmatic activity, further driving breakup and drifting of the Wyoming

and Karelia-Kola cratons (**Table 5-1**; **Figure 5-2**; Halls *et al.*, 2008).

Halls *et al.* (2008) proposed the formal name of the Marathon LIP to include the *ca.* 2.11 Ga Kapuskasing and Marathon dykes, and *ca.* 2.08 Ga Fort Frances dykes, which together form a 180° fanning dyke swarm pattern (**Table 5-1** and **Figure 5-2**). The Kapuskasing dykes form the branch of ENE-trending dykes east of Lake Superior and are contemporaneous with the NS-trending Marathon dykes located north of Lake Superior. While the Kapuskasing and Marathon dykes are the focus of this paper, hereafter referred to as the Marathon dykes, it is noted that the Marathon LIP defined by Halls *et al.* (2008) also included the 25 Ma younger *ca.* 2.08 Ga Fort Frances dykes. These form the western most branch of the radiating dyke swarm pattern, but because of their distinctly younger age, their geochemistry is not included in the present study.

The Marathon dykes have been the subject of previous paleomagnetic, geochronological and geochemical studies (see **Table 5-1** for a summary of U-Pb geochronology; Buchan *et al.*, 1996; Hamilton *et al.*, 2002; Halls *et al.*, 2008; Bleeker *et al.*, 2015). Paleomagnetic investigations by Buchan *et al.* (1996) and Halls *et al.* (2008) found that the emplacement of the Marathon dykes record a reversal in Earth's magnetic field in which the older dykes have a normal polarity and the younger dykes are reversed. Early geochemical studies found that the quartz tholeiitic Marathon dykes record a progression from enriched to depleted signatures through time and attributed their geochemical evolution to a magmatic system that progressively assimilated less crust as the rift system developed (Halls *et al.*, 2008).

Figure 5-2. Distribution of *ca.* 2.11 Ga Marathon Dykes in Superior craton showing sample locations with their assigned geochemical group and U-Pb age locations (**Table 5-1**). Samples compiled from Halls *et al.* (2008), Ernst and Bleeker (2010), Bleeker *et al.* (2015), and this study. Dykes from Buchan and Ernst (2004) Buchan and Ernst (2004), and Superior craton (dark pink) modified after (Goodfellow, 2007).

Table 5-1. Summary of U-Pb geochronology on 2.11 Ga mafic rocks in the Superior, Hearne, Wyoming and Karelia-Kola cratons.

ID	Craton Locality name/sample	Type	Latitude (°N)	Longitude (°E)	Age (Ma)	U-PB Dating Method and notes	Polarity	Trend (°)	Dip (°)	Width (m)	Reference
<i>Superior</i>											
A	LL20-3	D	50.5600	-87.1500	2125.7±1.3	Single grain Bdy	N	050	~90	>65	Halls et al., 2008
B	Site 22	D	49.18	-86.91	2121±14	Multiple grain Bdy	N	000	-	50	Buchan et al., 1996
C	183/BNB-14-019	D	48.0336	-82.7356	2121±8	Single grain Bdy		-	~90	-	Halls et al., 2008
D	BNB-12-022 ("Northern dyke")	D	46.2921	-82.8406	2116±5	Preliminary analysis, Bdy		283	~90	30	Bleeker et al., 2015
E	DP1-5	D	50.3170	-87.5200	2112±9	Single grain Bdy	R	045	~90	60	Halls et al., 2008
F	DP2-2	D	49.8035	-87.5750	2109±1.6	Single grain Bdy		015	~90	38	Halls et al., 2008
G	MK7-8	D	48.1200	-84.4100	2106±3.5	Single grain Bdy	R	045	~90	50	Halls et al., 2008
H	BNB-12-028	D	46.3484	-83.1879	2105±5	Preliminary analysis, Bdy		290	~90	50	Bleeker et al., 2015
I	TK48-19	D	47.7700	-84.3800	2104.6±1.8	Single grain Bdy	R	048	~90	24	Halls et al., 2008
J	Site 4 Jellicoe area/ MA-5	D	49.70	-87.46	2101±1.9	Multiple and single grain Bdy	R	000	85°W	76	Buchan et al., 1996; Hamilton et al., 2002
K	Site 19 Manitouwadge area /BXA91-21	D	49.13	-85.75	2101±1.6	Multiple grain Bdy	R	000-010	-	50	Buchan et al., 1996; Hamilton et al., 2002
<i>Hearne</i>											
-	Hurwitz Gabbro/91T9-4	S	-	-	2111.2±0.6	Multiple grain Bdy	-	-	-	-	Heaman and LeCheminant, 1993
<i>Wyoming</i>											
L	Laramie Range/LR4	D	42.3478	-105.7044	2115.4±9.8	SIMS Bd, Zrc		058	-	30	Kilian, 2015
M	FM-3	D	42.2395	-107.1016	2113±15	Multiple grain Bdy		NE	-	30	Bowers and Chamberlain 2006
N	Granite Mountains/T10GM11	D	42.6517	-107.3270	2111 ±22	SIMS Bdy	R	085	-	16.5	Kilian, 2015
O	Wind River Range/95SP-2	D	42.4063	-108.9480	2106±13	SIMS Bd, Zrc			-		Kilian, 2015

Table 5-1 continues on the next page

Notes: Localities A-D1 correspond to figures from **Figure 5-1** to **Figure 5-4**. Abbreviations: Type: D: Dyke, S: Sill, I: Intrusion, MV: Mafic volcanic rock; DDH - Sample from diamond drill hole; Bdy - baddeleyite; Zrc - Zircon; Ttn - titanite; CA - chemically abraded. Coordinates with two decimals limited by original publication. All geochronology methods are by ID-TIMS unless indicated otherwise. ~ Dips are within 10° of vertical. † Dip is reported as subvertical.

Table 5-1. Continued.

ID	Craton	Type	Latitude	Longitude	Age	U-PB Dating	Trend	Dip	Width	Reference
	Locality name/sample		(°N)	(°E)	(Ma)	Method and notes	(°)	(°)	(m)	
<i>Karelia-Kola</i>										
P	Pittiövaara Sodankylä/A0841	D	67.4357	26.382259	2135±31	Multiple grain Zrc	-	-	-	Räsänen and Huhma, 2001
Q	Myllykangas/16SD_019	D	65.5407	25.3306	2135.2+3.6/-3.7	Single grain CA-Zrc and Bdy	350	-	30	Chapter 2
R	Koppakumpu diabase/A1214	D	66.2611	24.9114	2130±4	SHRIMP, Zrc	-	-	-	Huhma et al., 2018
S	Sipojuntti/15SD_092	D	65.7310	24.8864	2128.9±1.2	Single grain Bdy	320	-	75	Chapter 2 ; see also Perttunen and Vaasjoki, 2001
T	Pirtguba/Ca-255	D	64.9790	30.7280	2126±5	SHRIMP-II, Zrc	NNW (325)	90†	70	Stepanova et al., 2014
U	Pitkännuotionpulju Gabbro/A066	S	67.83345	26.826878	2124±5	Multiple grain Zrc		-		Manninen et al., 2001
V	Matkavaara Albitite/A1684	S	66.5675	26.9578	2123±7	Ion Microprobe, Zrc; U-Pb TIMS, Ttn	-	-	-	Niiranen et al., 2005
W	Kaakamovaara Tornio/A410 A986	S	66.1346	24.2041	2117±7	Multiple grain Zrc	-	-	-	Perttunen and Vaasjoki, 2001
X	Misi gabbro/A077/A078/94001981	S	66.6187	26.6800	2117±4	Multiple grain, air abraded Zrc	-	-	-	Perttunen and Vaasjoki, 2001; Niiranen et al., 2003
Y	C-2717-2a	D	65.9222	33.3928	2114.9±16	Ion Microprobe NORDSIM, Zrc	-	-	30	Stepanova and Stepanov, 2010
Z	Kylälampi/A817	I	67.5969	27.6301	2114±5	Multiple grain, air abraded Zrc	-		-	Räsänen and Huhma, 2001
Z	Nieminen/15SD_082	D	63.3773	27.8036	2114.4±3	Single grain CA-Zrc	280	-	>30	This study; see also Toivola et al., 1991; Huhma et al., 2018
B1	Piipolanmäki/465 and Kutsu/A427	D	62.4440 62.2900	30.2870 30.5750	2113±4	Multiple grain Zrc	NW	90†	-	Pekkarinen and Lukkarinen, 1991
C1	Hirsimaa Formation	MV	66.1990	25.1912	2106±8	SHRIMP, Zrc	-	-	-	Karhu et al., 2007
D1	Oravaara Basalt	MV	62.8220	29.2480	2105±16	Multiple grain Zrc	-	-	-	Huhma, 1986; see also Nykänen et al., 1994

Notes: Localities A-D1 correspond to figures from **Figure 5-1** to **Figure 5-4**. Abbreviations: Type: D: Dyke, S: Sill, I: Intrusion, MV: Mafic volcanic rock; DDH - Sample from diamond drill hole; Bdy - baddeleyite; Zrc - Zircon; Ttn - titanite; CA - chemically abraded. Coordinates with two decimals limited by original publication. All geochronology methods are by ID-TIMS unless indicated otherwise. ~ Dips are within 10° of vertical. † Dip is reported as subvertical.

5.3.2 Hearne Craton

The Hearne craton is embedded within the Churchill Province of continental North America. Fused to the Rae craton via the 1.9 Ga Snowbird tectonic zone (Berman *et al.*, 2007), the Hearne craton consists mainly of 2.74-2.67 Ga metavolcanic and metasedimentary supracrustal rocks (Aspler and Chiarenzelli, 1996). Mafic magmatism dated at 2.11 Ga has been reported in the Griffin sills from the deformed Hurwitz basin (Heaman and LeCheminant, 1993; Aspler *et al.*, 2002) and, by the Kazan and Chipman dykes (Grohn *et al.*, 2016; Regan *et al.*, 2016). Geochemistry of the Griffin sills has been studied in detail by Aspler *et al.*, (2002). The enriched, alkaline signatures and $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values between +0.7 and +1.0 are consistent with magmas sourced from enriched asthenosphere at depths and pressures within the garnet stability field. These geochemical characteristics are consistent with magmatism related to a mantle plume (Aspler *et al.*, 2002). Geochemistry for the Kazan dykes is included in the database compiled by Ernst and Buchan (2010) and the highly deformed and metamorphosed Chipman dykes have been previously studied by Flowers *et al.* (2006).

5.3.3 Wyoming craton

In contrast to the well exposed Superior craton, most of the Precambrian basement in the Wyoming craton is covered by Phanerozoic sedimentary rocks (Love and Christiansen, 1985; Sims *et al.*, 2001). Exposures of Precambrian crust can be found in the Teton, Wind River, and Laramie Mountain ranges, among a number of smaller and less extensive ranges, all of which were uplifted by thrusting during the Cretaceous Laramide orogeny (Hoy and Ridgway, 1997; Fan and Carrapa, 2014). Basement rocks in these blocks include

Paleoarchean metasedimentary units, Meso to Neoarchean metaplutonic suites and rarely Eoarchean rocks (i.e., 3.5 Ga trondhjemite gneiss; Chamberlain and Mueller, 2007).

Figure 5-3. Distribution of *ca.* 2.11 Ga Bear Mountain dykes in Wyoming showing sample locations with their assigned geochemical group and ages where available; age references as in **Table 5-1**. Dyke lines and sample locations modified from (Kilian, 2015). Precambrian geology of Wyoming modified from Horton *et al.* (2017), Archean in dark pink, younger cover in light pink.

Paleoproterozoic dyke swarms occur in high concentrations throughout the exposed basement blocks but the limited exposure has resulted in regional dyke swarm patterns remaining poorly defined (Kilian, 2015). Despite these challenges, paleomagnetic and geochronological studies, including dykes at 2.11 Ga, have been conducted on many Paleoproterozoic dykes across the Wyoming craton (Redden *et al.*, 1990; Cox *et al.*, 2000; Bowers and Chamberlain, 2006; Dahl *et al.*, 2006; Hark, 2008; Kilian *et al.*, 2016a). The *ca.* 2.11 Ga Bear Mountain dyke swarm located in the Wind River, Granite, Bear and

Laramie Ranges (**Figure 5-3**) recorded the same magnetic field reversal found in the contemporaneous Marathon dykes but no previous geochemical studies have been conducted (Buchan *et al.*, 1996; Bowers and Chamberlain, 2006; Kilian, 2015).

5.3.4 Karelia-Kola craton

The composite Karelia-Kola craton (“greater Karelia” of Bleeker and Ernst, 2006) is an aggregation of independently formed *ca.* 3.5-2.7 Ga microcontinents and interspersed *ca.* 2.8 Ga greenstone belts that assembled during the 2.72 Ga Belomorian orogeny (Mutanen and Huhma, 2003; Slabunov *et al.*, 2006b; Hölttä *et al.*, 2008, 2012; Lahtinen *et al.*, 2011; Mints, 2015b, 2015a; Mints *et al.*, 2015). Karelia-Kola differs from the Superior and Wyoming cratons in that the 1.92-1.79 Ga Svecofennian orogeny caused major reworking of crustal blocks, particularly on the western margin of Karelia (Kärki *et al.*, 1993; Vuollo *et al.*, 2000; Torvela and Ehlers, 2010; **Chapter 2; Chapter 3**) Proterozoic sedimentary cover of the Karelian Supergroup is also a major component of the Precambrian record in Karelia-Kola, covering most of Finnish Lapland.

The extensive Paleoproterozoic mafic magmatic record in from 2.50 to 1.98 Ga includes mafic dykes, sills and metavolcanic rocks at *ca.* 2.11 Ga in the Karelia-Kola craton (**Figure 5-4**; see **Table 5-1** for a summary of U-Pb geochronology and references). Dykes include the dense NW-trending *ca.* 2135-2113 Ma Tohmajärvi swarm, which spans across various crustal blocks of Finnish Karelia as well as the parallel Pirtguba swarm in Russian Karelia (Toivola, 1988; Pekkarinen and Lukkarinen, 1991; Toivola *et al.*, 1991; Vuollo *et al.*, 1992; Nykänen *et al.*, 1994; Räsänen and Huhma, 2001; Vuollo and Huhma, 2005; Stepanova and Stepanov, 2010; Stepanova *et al.*, 2014, 2015). Recent U-Pb zircon and

baddeleyite ID-TIMS ages presented in Chapter 2 of the Sipojuntti dyke (2128 Ma; see also Perttunen and Vaasjoki, 2001) and Myllykangas dyke (2135 Ma) confirm the presence of an older pulse in the Pudasjärvi block of Western Karelia (**Figure 5-5a and b**). In Lapland, the Misi gabbro sills, which are hosted in the Peräpohja schist belt, are 2117 Ma and associated with local skarn deposits (Manninen and Huhma, 2001; Räsänen and Huhma, 2001; Niiranen *et al.*, 2003, 2005). Mafic volcanic rocks at approximately 2.11 Ga are documented in the Peräpohja schist belt (Perttunen and Hanski, 2003) and include the mafic tuff of the Hirsimaa formation dated at 2106 ± 8 Ma (Karhu *et al.*, 2007; Huhma *et al.*, 2018) and the amygdaloidal basalt from the Jouttiaapa formation approximately dated by Sm-Nd at 2090 ± 70 Ma (Huhma *et al.*, 1990). The southernmost extent of 2.11 Ga mafic magmatism is marked by the Oravaara metabasalts, dated by Huhma (1986) at 2105 ± 16 Ma from U-Pb analyses on zircon (Nykänen *et al.*, 1994).

Previous geochemical studies of the 2.11 Ga mafic dykes and sills in Finland and Russian Karelia have identified at least two distinct geochemical patterns: MORB-type patterns in the Pirtguba dykes and Fe-tholeiitic dykes with high La/Yb ratios in the Tohmajärvi dykes and Misi gabbro sills (Vuollo *et al.*, 1992; Nykänen *et al.*, 1994; Niiranen *et al.*, 2003, 2005; Stepanova *et al.*, 2014). While the MORB-type dykes are considered to have been produced below thin lithosphere, those with an Fe-tholeiitic chemistry have been attributed melting at deeper depths and assimilation of crust related to magma generation below thicker lithosphere (Nykänen *et al.*, 1994; Stepanova *et al.*, 2014). Some whole-rock major and trace element geochemistry and Nd isotope studies have been completed for the Hirsimaa, Jouttiaapa ($\epsilon\text{Nd}_{T=2.09\text{Ga}} = +5.1$ to $+3.2$) and Oravaara ($\epsilon\text{Nd}_{T=2.1\text{Ga}} = +2.6$) metabasalts and these have reported both MORB and OIB-like patterns

(Huhma, 1986; Huhma *et al.*, 1990; Perttunen and Hanski, 2003).

Figure 5-4. Distribution of *ca.* 2.11 Ga Tohmajärvi-Pirtguba dykes, sills and mafic volcanic rocks in Karelia-Kola showing sample locations with their assigned geochemical group and ages where available; age references as in **Table 5-1**. Dykes, bedrock geology and sample locations modified from Perttunen and Hanski (2003); Hölttä *et al.* (2008); GTK (2012); Lahtinen (2012); Stepanova *et al.* (2014); GTK (2018).

5.4 Methods

5.4.1 U-Pb geochronology

Approximately 10 kg of medium to coarse grained gabbro was collected from the Nieminen dyke from the Iisalmi block of Western Karelia. After a thorough scrubbing to remove organic and loose material, a heavy, non-magnetic mineral concentrate was prepared by first crushing in a jaw crusher and ring mill, and then using a Wilfley table, magnetic separator and methylene iodide. Zircon crystals were handpicked from the heavy mineral separate under a binocular microscope. Following Mattinson (2005), the zircons were annealed in a 1000°C furnace and then subjected to 6 hours of chemical abrasion. U-Pb analysis by the ID-TIMS method was completed at the Jack Satterly Geochronology Laboratory, University of Toronto (Krogh, 1973). During sample loading, a 1.14 mg spike of $^{202-205}\text{Pb}$ - $^{233-235}\text{U}$ (ET2535 tracer from the EARTHTIME project as described by Condon et al., 2015) was added to the Teflon dissolution capsules. Each single zircon crystal was dissolved in 0.10 ml of hydrofluoric acid (HF) and 0.02 ml of 8N nitric acid (HNO_3) at 195°C. Following dissolution, samples were dried down and dissolved again in 0.15 ml of 3N hydrochloric acid (HCl). In-house made 50 μl anion exchange columns were used to extract U and Pb. Phosphoric acid (H_3PO_4) was added to the U-Pb elutions before being dried down on a hot plate. Samples were loaded with silica gel onto outgassed rhenium filaments. U and Pb were analyzed in static mode with a VG M354 mass spectrometer using multiple Faraday cups. Ellipses and age errors in the text, tables and concordia diagrams reflect 2σ uncertainties. The Excel add-in Isoplot (Ludwig, 2003) was used for plotting of all data. Results of standards (R33, OG1 and EARTHTIME 2.0 Ga synthetic solution of Black et al., 2004; Condon et al., 2015. and Stern et al., 2019) analyzed over

the duration of this study are presented in **Appendix C**.

5.4.2 Geochemistry

Whole-rock major and trace element geochemical data were collected for mafic dykes from the Superior (12 dykes), Karelia (3 dykes) and Wyoming (20 dykes) cratons (**Table 5-3**). At Carleton University, samples were crushed to <2mm in a Mn-steel jaw crusher and then pulverized in an agate mill. Pulverized samples weighing at least 5 g were sent to ALS Laboratories (www.alsglobal.com) for analysis of 13 major oxides, and 42 minor and trace elements by inductively coupled plasma-atomic emission spectrometry (ICP-AES) and inductively coupled plasma-mass spectrometry (ICP-MS). Trace element samples were prepared with the lithium borate fusion method prior to dissolution and analysis by ICP-MS.

In addition to the geochemical data produced from this study, the overall investigation has incorporated geochemical data from Huhma *et al.* (1990), Aspler *et al.* (2002), Niiranen *et al.* (2003, 2005), Halls *et al.* (2008), Ernst and Buchan (2010), Stepanova and Stepanov (2010), Stepanova *et al.* (2014, 2015), and GTK (2018). See also Toivola (1988), Toivola *et al.* (1991), Nykänen *et al.* (1994), and Vuollo and Huhma (2005) for additional geochemical data that were not incorporated into the database due to insufficient trace element data available or because only averages were presented in the data tables. Geochemical samples were filtered such that only samples with major element totals greater than 97% were used.

5.4.3 Sm-Nd isotopes

Sm-Nd isotope analyses were completed for 27 dykes belonging to 2.11 Ga units from the Karelia (3), Superior (13), and Wyoming (11) cratons and whenever possible, dykes with U-Pb ages were prioritized over those without a known age (**Table 5-4**). Pulverized samples weighing between 29 and 120 mg were spiked with a ^{148}Nd - ^{149}Sm solution and dissolved in a mixture of 16M HNO_3 and ~29M HF for 5 days on a hot plate at 110°C. The dissolved samples were then dried down and redissolved in a mixture of 6M HCl and 8M HNO_3 . The following day, the sample was dried down again and redissolved in 2.5M HCl and then loaded into 14 mL Bio-Rad borosilicate glass chromatography columns filled with 3 mL of Dowex AG50W-X8 cation resin. The columns were washed with 23mL of 2.5M HCl through the columns before the Rare earth elements (REE) were collected using 9 mL of 6M HCl. This rare earth element elution was dried down on a hot plate set at 110°C overnight. The REE residue was then dissolved in 0.26 M HCl and loaded into Eichrom Ln Resin chromatographic columns containing Teflon powder coated with HDEHP [di(2-ethylhexyl)] orthophosphoric acid (Richard et al., 1976). The columns were washed with 6.5 mL of 0.26M HCl prior to the collection of Nd elutions by passing 5mL of 0.26M HCl through the columns. Sm elutions were then collected using 0.5M HCl. Both the Nd and Sm elutions were left to dry down on a hot plate set at 110°C overnight with 5 μL of 0.3M H_3PO_4 . Approximately 3 μL of H_3PO_4 was added to the sample for loading and approximately 50% of the sample was loaded onto an outgassed Re filament part of a double filament assembly. Sm and Nd analyses were completed using a Thermo-Finnigan Triton TI thermal ionization mass spectrometer with filaments heated to 1500 and 1800°C. Isotope ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$.

In addition to the sample measurements, a USGS standard (BCR-2) was run yielding Nd = 28.14 ppm, Sm = 6.477, and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512648$. Over a three year period additional analyses of the 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, 2σ). Furthermore, an IGGRC in-house Nd standard measured routinely over a period of three years yield $^{143}\text{Nd}/^{144}\text{Nd} = 0.511831 \pm 0.000011$ (2σ), an equivalent value of $^{143}\text{Nd}/^{144}\text{Nd} = 0.511861 \pm 0.000009$ (2σ) for the La Jolla Nd standard. Uncertainties on Sm and Nd concentrations are estimated at less than 0.2%.

5.5 Results

5.5.1 Geochronology of the Nieminen dyke, Western Karelia

Mineral separation of medium- to coarse grained gabbro from the Nieminen dyke yielded abundant clear, colourless, tabular or lath shaped gem-quality zircon crystals and fragments up to 300 μm long. Three single grain zircon fragments approximately 200 μm long were analyzed (**Table 5-2**). The three analyses yielded concordant and overlapping results and the weighted mean of the three $^{207}\text{Pb}/^{206}\text{Pb}$ ages is 2114.4 ± 3 Ma which represents the crystallization age of the dyke (**Figure 5-5d**). These data plot to the right of the regression line within error of the age 2111 ± 5 Ma by Huhma et al. (2018) from 8 fractions of air-abraded zircon (**Figure 5-5c and d**; see also Toivola *et al.*, 1991).

Figure 5-5. U-Pb concordia diagrams for selected *ca.* 2.11 Ga Tohmajärvi-Pirtguba dykes. A. The 2135±3.6/-3.7 Ma Myllykangas dyke dated with chemically abraded zircons and single grain baddeleyite (**Chapter 2**). B. The 2128±1.2 Ma Sipojuntti dyke dated with single grain baddeleyite. C. Air abraded zircon results with an upper intercept of 2111±5 Ma of the Nieminen dyke from (Huhma *et al.*, 2018). The extent of panel d is indicated by the black rectangle. D. Three concordant chemically abraded single zircon analyses from the Nieminen dyke, this study, with a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2114.4±3 Ma.

Table 5-2. U-Pb isotopic data for the Nieminen dyke, Western Karelia.

Fraction (analysis number)	Weight (μg)	U (ppm)	Th U	Pbtot (pg)	Pbcom (pg)	^{206}Pb ^{204}Pb Meas. Meas.	^{207}Pb ^{235}U	2 σ	^{206}Pb ^{238}U	2 σ	Error Corr.
<i>Nieminen dyke, 15SD_082</i>											
Zrc1 (sk37p229)	4.1	42	0.76	77.2	0.5	9090	7.017	0.018	0.3878	0.0008	0.919
Zrc2 (sk37p228)	3.1	48	0.68	66.4	0.2	17270	7.018	0.017	0.3876	0.0008	0.923
Zrc3 (sk37p227)	3.7	38	0.67	62.6	0.2	14687	7.001	0.017	0.3872	0.0008	0.940

Table 5-2. Continued.

Fraction (analysis number)	^{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. (%)
<i>Nieminen dyke, 15SD_082</i>									
Zrc1 (sk37p229)	0.13121	0.00013	2112.8	3.8	2113.5	2.2	2114.2	1.8	0.1
Zrc2 (sk37p228)	0.13131	0.00013	2111.9	3.7	2113.8	2.2	2115.6	1.7	0.2
Zrc3 (sk37p227)	0.13114	0.00011	2109.6	3.6	2111.5	2.1	2113.3	1.5	0.2

Notes: Fractions are single grain chemically abraded zircon analyses. Zrc – zircon

Th/U calculated from radiogenic $^{208}\text{Pb}/^{206}\text{Pb}$ ratio and $^{207}\text{Pb}/^{206}\text{Pb}$ age assuming concordance.

Pbtot is amount of Pb in picograms excluding blank. Pbcom is total amount of common Pb in picograms assigned to the isotopic composition of laboratory blank.

Meas. indicates the measured ratio.

$^{206}\text{Pb}/^{204}\text{Pb}$ corrected for fractionation and common Pb in the spike.

Correction for ^{230}Th disequilibrium in $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ 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 $^{207}\text{Pb}/^{206}\text{Pb}$ age.

Uranium decay constants are from Jaffey *et al.* (1971).

5.5.2 Geochemistry

5.5.2.1 Whole-rock major and trace element geochemistry

All selected dykes from Karelia, Superior and Wyoming are classified as tholeiitic basalt to basaltic andesite on a total alkali silica diagram (**Figure 5-6**). The subalkaline/tholeiitic rocks are further classified on a Jensen ternary diagram with a range of compositions from high Fe-tholeiitic to high Mg-tholeiitic basalts (**Figure 5-6**). The tholeiitic basaltic classification is consistent with the Nb/Y versus Zr/Ti classification diagram which indicate basalt to basaltic andesite compositions (**Figure 5-9a**). Dykes from each craton have consistent SiO₂ wt. % and, with the exception of CaO, no correlations are present when compared to Al₂O₃, TiO₂, P₂O₅, and Fe_{tot} (**Figure 5-7**). Major element oxides Al₂O₃, CaO, TiO₂, P₂O₅, Na₂O, K₂O, and Fe_{tot} show negative correlations to Mg number (which ranges from 32-61; **Figure 5-8**). Additionally, a weak correlation between CaO/Al₂O₃ and Mg number is also present (**Figure 5-8**). With the exception of four samples (out of 38) all dykes have Ti/V ratios between 20 and 50, plotting in the continental flood basalt field after Shervais (1982; **Figure 5-9b**). Three samples have lower Ti/V ratios and plot in the arc basalt field and one sample has a Ti/V ratio that is greater than 50 and plots within the ocean island basalt field (**Figure 5-9b**; Shervais, 1982).

Figure 5-6 Major element classification of the *ca.* 2.11 Ga dykes from Superior, Wyoming, and Karelia-Kola. Results presented in **Table 5-3**. Top: Total alkali silica classification diagram (Le Bas et al., 1986). Bottom: Al-Fe_{tot}+Ti-Mg ternary classification diagram (Jensen, 1976).

Figure 5-7. Major element oxides Al₂O₃, CaO, TiO₂, P₂O₅, SiO₂ and Fe_{tot} against SiO₂ for the *ca.* 2.11 Ga dykes from Superior, Wyoming, and Karelia-Kola. Two-sigma error is based on reproducibility of the LT-10 basalt standard (see methods and **Table 5-3**). Errors are not shown on the Y-axis when 2σ is less than 0.2 (or less than the thickness of the X-axis error bar).

Figure 5-8. Major element oxides Al_2O_3 , CaO , $\text{CaO}/\text{Al}_2\text{O}_3$, P_2O_5 , Na_2O , K_2O , TiO_2 and FeO_{tot} against Mg number for the *ca.* 2.11 Ga dykes from Superior, Wyoming, and Karelia-Kola. Two-sigma error is based on reproducibility of the 10-LT-05 basalt standard (see methods and **Table 5-3**). Errors not shown when 2σ is less than 0.2.

Figure 5-9. Whole-rock major and trace element characteristics of sampled dykes from Superior, Wyoming, and Karelia-Kola. Results presented in **Table 5-3**. a. Zr/Ti versus Nb/Y classification plot (Pearce, 1996). b. V versus Ti (Shervais, 1982). c. $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ versus $1/\text{Nd}$.

Three multi-element patterns (normalized to primitive mantle) are recognized within the 2.11 Ga dykes from the Karelia, Superior and Wyoming cratons (**Figure 5-10**). The flat MORB-type signature of pattern 1 is unique to a subset of the Tohmajärvi-Pirtguba dykes found in the Karelia-Kola craton (**Figure 5-10a**). The geochemistry of pattern 2 is distinguished by negative La/Sm slopes (ratios >1) and Nb abundances approximately equal to Th and La when normalized to primitive mantle. Although rare, dykes with pattern 2 geochemistry are located in the Superior and Wyoming cratons (**Figure 5-10c** and **e**). Pattern 3 geochemical signature is the most common and found in the Karelia-Kola, Superior, and Wyoming cratons. The multi-element patterns associated with pattern 3 include flat to shallow heavy rare earth element (HREE) slopes ($\text{Tb/Yb}_{\text{PM}} \sim 1$) and negative light rare earth element (LREE) slopes ($\text{La/Sm}_{\text{PM}} > 1.4$). Distinguishing characteristics of pattern 3 include a negative Nb anomaly and Th concentrations greater than La (**Figure 5-10b, d, and f**).

Figure 5-10. Multi-element diagrams normalized to Primitive Mantle for sampled dykes from Superior, Wyoming and Karelia-Kola. a and b. Patterns 1 and 3 from Karelia-Kola. c and d. Patterns 2 and 3 from Superior. e and f. Patterns 2 and 3 from Wyoming. Blue lines highlight dykes that have U-Pb ages. Black lines represent undated dykes. Primitive mantle values from Sun and McDonough (1989).

Table 5-3. Whole-rock major and trace element geochemistry for 2.11 Ga dykes from the Superior, Wyoming and Karelia cratons.

Sample	Geochemical group	Latitude (Y)	Longitude (X) <i>Detection limit.</i>	SiO ₂ <i>0.01</i>	Al ₂ O ₃ <i>0.01</i>	Fe ₂ O ₃ <i>0.01</i>	CaO <i>0.01</i>	MgO <i>0.01</i>	Na ₂ O <i>0.01</i>	K ₂ O <i>0.01</i>	Cr ₂ O ₃ <i>0.01</i>	TiO ₂ <i>0.01</i>
<i>Superior</i>												
BNB-12-021A ^{1A}	3	46.2892	-82.8352	50.20	13.85	11.8	10.65	6.88	1.75	0.51	0.01	1.05
BNB-12-021B ^{1A}	3	46.2892	-82.8352	52.00	15.35	11.7	10.65	6.60	1.60	0.78	0.01	1.09
BNB-12-022 ^A	3	46.2921	-82.8406	51.00	12.85	17.55	8.01	4.77	2.47	1.21	0.01	2.02
BNB-12-023A ^A	3	46.2491	-82.8782	48.70	13.70	14.45	10.95	6.89	1.95	0.75	0.02	1.43
BNB-12-023B ^A	3	46.2491	-82.8782	48.90	13.85	14.15	10.50	6.93	2.11	0.86	0.03	1.41
BNB-12-024 ^A	3	46.2480	-82.8791	48.80	14.05	13.8	10.30	6.91	2.11	1.20	0.03	1.41
BNB-12-027 ^A	3	46.4700	-82.5432	48.50	14.20	11.45	10.70	7.36	2.09	0.66	0.03	0.95
BNB-12-028A ^A	3	46.3484	-83.1879	50.20	13.15	16.1	8.23	4.83	2.14	2.09	0.01	1.89
BNB-12-028B ^A	3	46.3484	-83.1879	49.40	13.00	15.7	7.80	5.39	3.19	0.99	0.01	1.88
BNB-14-012A ^A	3	48.2202	-82.5940	47.60	13.00	18.4	9.67	5.89	2.27	0.49	0.01	1.90
BNB-14-019C ^{2A}	3	48.0336	-82.7356	50.00	12.80	15.45	10.70	6.57	2.14	0.39	0.02	1.60
BNB-14-019D ^{2A}	3	48.0336	-82.7356	49.80	12.65	15.05	10.90	6.97	2.03	0.37	0.03	1.50
BNB-14-019E ^{2A}	3	48.0336	-82.7356	48.70	13.05	14.3	10.35	6.42	2.69	0.64	0.02	1.48
LL20-3	2	50.5600	-87.1500	49.40	16.05	13.80	9.16	3.60	2.58	1.00	0.01	1.92
TK48-19	3	47.7700	-84.3800	47.80	11.20	18.85	8.61	5.48	2.18	1.05	0.02	2.91
DP1-5	3	50.3170	-87.5200	49.60	14.05	13.05	10.90	6.99	2.19	0.48	0.02	1.12
MK7-8	3	48.1200	-84.4100	50.80	14.30	13.30	10.90	7.30	2.01	0.42	0.03	1.06
DP2-2	3	49.8035	-87.5750	49.50	13.05	16.00	9.70	5.51	2.23	0.75	0.01	1.59
<i>Wyoming</i>												
95SP-2	2	42.4063	-108.9480	48.30	14.30	16.75	9.91	5.71	2.47	0.72	0.01	2.44
95SP-2*				48.00	14.20	16.30	9.88	5.72	2.45	0.73	0.01	2.31
BM1-4	1B	42.2395	-107.1016	47.10	13.60	17.50	9.36	5.84	2.39	0.61	0.01	2.82
BM3	2	42.2588	-107.1030	49.40	13.20	15.55	9.09	4.97	2.33	0.53	0.01	2.07
SM3	3	42.1935	-106.8267	50.10	14.20	13.10	9.32	5.56	2.23	1.11	0.02	1.53
SM4	3	42.1866	-106.8269	50.00	14.65	14.95	9.93	4.98	2.11	0.85	0.01	1.58
SM6-3	3	42.1817	-106.8248	51.60	12.65	14.80	6.61	4.50	3.66	1.13	0.01	1.94
SM7-3	3	42.1831	-106.8322	49.30	14.05	14.15	9.53	6.17	1.81	0.90	0.02	1.31
GM5-3	3	42.5453	-107.5693	52.50	13.10	9.45	10.75	7.63	1.61	0.48	0.06	1.01
GM7-1	3	42.6447	-107.3618	50.40	13.50	16.30	7.99	5.85	2.62	1.32	0.03	1.87
GM8	3	42.4414	-107.2083	50.00	13.55	12.95	10.75	6.61	2.14	0.42	0.02	1.26
GM11	3	42.6517	-107.3270	50.40	14.20	12.50	9.10	6.06	2.90	1.25	0.02	1.47
LR2	3	42.4241	-105.7012	49.40	15.05	11.50	11.15	6.93	1.58	0.73	0.02	0.74
LR5-6	3	42.2106	-105.6503	49.90	12.95	14.10	7.98	5.54	3.31	0.97	0.02	1.78
LR6-1	3	42.2094	-105.6488	51.50	13.00	14.65	6.43	4.42	4.65	0.83	0.01	2.21
LR8-7	2	42.5328	-105.7487	49.10	11.60	16.40	7.76	6.56	2.72	0.84	0.01	1.99
LR16-2	3	42.5113	-105.7814	48.70	13.25	15.10	9.63	6.18	1.92	0.74	0.02	1.68
<i>Dykes subparallel to 2.1 Ga dykes</i>												
LR11-6	3	42.5328	-105.7487	48.90	12.40	16.50	8.00	4.35	2.26	1.35	0.01	2.38
GM15-3	3	42.5759	-107.6122	54.60	13.50	11.10	7.81	4.72	3.48	0.68	0.01	1.22
GM16	3	42.5776	-107.6072	53.40	13.65	11.40	7.76	5.40	2.89	1.57	0.01	1.15
LR1-8	3	42.4247	-105.7040	49.70	13.85	11.75	11.50	7.68	1.06	0.52	0.03	0.82
<i>Karelia</i>												
15SD_082 ^B	1A	63.3773	27.8036	50.70	12.95	16.10	10.35	6.38	2.12	0.15	0.01	1.12
15SD_092 ^B	3	65.7310	24.8864	46.40	12.55	14.05	9.13	8.59	2.69	0.70	0.08	2.29
16SD_019A ^C	1A	65.5407	25.3306	49.40	13.15	15.35	10.70	6.82	1.96	0.27	0.03	1.04
10-LT-05 Standard – Batch A				49.30	18.65	9.79	9.5	6.51	3.59	1.03	0.03	1.45
10-LT-05 Standard – Batch B				49.00	18.15	9.68	9.57	6.53	3.61	1.01	0.03	1.44
10-LT-05 Standard – Batch C				48.70	17.9	9.69	9.61	6.5	3.52	0.99	0.03	1.42
10-LT-05 Standard – Average				48.88	18.18	9.54	9.52	6.61	3.60	1.02	0.03	1.44
2σ				1.16	0.50	0.29	0.32	0.21	0.20	0.03	0.01	0.06

Table 5-3 continues on the next page

Notes: Major elements in wt. %, trace elements in ppm. Sample 10-LT-05 is a basalt standard by Carleton University; results for batches A, B and C are shown. Superscripts A, B or C identify batches. 10-LT-05 average and 2σ values of 25 analyses spanning 6 years are shown. *duplicate. 1. “Granary Lake dyke”; 2. Site 183 from Halls *et al.* (2008).

Table 5-3. Continued.

Sample	MnO	P ₂ O ₅	SrO	BaO	LOI	Total	Ba	Ce	Cr	Cs	Dy	Er	Eu	Ga	Gd	Hf
Detection limit	0.01	0.01	0.01	0.01	0.01	0.01	0.5	0.5	10	0.01	0.05	0.03	0.03	0.1	0.05	0.2
<i>Superior</i>																
BNB-12-021A ¹	0.18	0.07	0.01	0.01	1.46	98.43	80.7	16.2	60	1.08	3.46	2.06	1.07	18.4	3.41	2.1
BNB-12-021B ¹	0.18	0.08	0.02	0.02	1.91	101.99	146.5	18.5	80	1.00	3.36	2.10	0.93	17.9	3.39	2.1
BNB-12-022	0.22	0.29	0.02	0.04	1.39	101.85	373.0	71.6	40	2.67	6.43	4.13	2.08	21.0	7.15	5.0
BNB-12-023A	0.25	0.12	0.02	0.01	1.53	100.77	86.0	18.5	190	3.56	4.52	2.57	1.48	20.0	4.72	2.6
BNB-12-023B	0.22	0.11	0.02	0.01	1.70	100.80	94.7	17.7	190	2.50	4.60	2.37	1.35	19.7	4.44	2.6
BNB-12-024	0.22	0.09	0.02	0.02	1.78	100.74	144.5	16.1	190	0.76	4.03	2.57	1.25	18.6	4.10	2.2
BNB-12-027	0.19	0.08	0.02	0.01	2.25	98.49	102.5	16.4	230	1.16	3.08	2.04	0.89	14.6	2.73	1.5
BNB-12-028A	0.22	0.29	0.02	0.06	1.00	100.23	493.0	70.1	40	6.52	5.86	3.62	2.08	20.2	6.56	4.3
BNB-12-028B	0.22	0.25	0.02	0.03	1.54	99.42	288.0	56.4	50	1.07	5.12	3.22	1.89	18.5	5.91	4.0
BNB-14-012A	0.26	0.17	0.02	0.03	-0.16	99.55	234.0	33.6	60	0.30	5.12	3.16	1.58	20.1	4.97	2.8
BNB-14-019C ²	0.25	0.16	0.01	0.02	-0.15	99.96	178.5	26.8	150	0.33	5.11	3.15	1.55	17.2	4.79	2.6
BNB-14-019D ²	0.24	0.14	0.01	0.02	-0.02	99.69	161.0	25.3	180	0.30	4.86	3.16	1.45	17.1	4.66	2.5
BNB-14-019E ²	0.23	0.14	0.02	0.02	1.98	100.04	196.0	27.1	160	0.56	4.76	3.04	1.43	16.7	4.70	2.4
LL20-3	0.18	0.18	0.03	0.03	1.26	99.20	305.0	45.9	50	0.98	5.23	2.89	1.54	21.8	5.31	4.1
TK48-19	0.24	0.45	0.02	0.05	0.24	99.10	465.0	78.9	110	0.71	8.01	4.69	2.38	20.9	8.89	5.9
DP1-5	0.18	0.12	0.02	0.01	1.14	99.87	133.0	25.8	140	1.30	3.51	2.01	1.19	17.2	3.26	1.8
MK7-8	0.20	0.11	0.02	0.02	0.42	100.89	148.0	25.7	170	1.47	3.28	2.07	0.98	15.5	3.33	2.2
DP2-2	0.22	0.17	0.02	0.02	1.06	99.83	163.0	39.0	70	1.29	4.95	2.91	1.53	17.7	4.84	3.4
<i>Wyoming</i>																
95SP-2	0.23	0.24	0.03	0.03	0.31	101.45	240.0	40.4	80	1.60	4.47	2.73	1.84	22.0	5.38	3.8
95SP-2*	0.22	0.24	0.03	0.03	0.39	100.51	242.0	242.0	42.0	80.0	1.46	4.67	2.38	1.88	20.8	5.15
BM1-4	0.24	0.29	0.02	0.02	1.27	101.07	202.0	42.5	90	0.63	6.70	4.29	2.25	24.0	7.27	4.7
BM3	0.22	0.22	0.03	0.02	0.38	98.02	226.0	50.2	90	1.08	5.94	3.30	2.04	21.4	6.86	4.8
SM3	0.18	0.14	0.02	0.02	1.75	99.28	180.5	25.9	110	0.55	4.13	2.49	1.23	19.9	4.77	3.1
SM4	0.21	0.16	0.02	0.02	1.11	100.58	159.5	26.3	100	0.99	5.13	2.89	1.41	22.1	4.70	2.9
SM6-3	0.15	0.26	0.02	0.03	1.40	98.76	249.0	54.6	60	2.58	6.39	4.11	1.81	20.2	6.91	5.3
SM7-3	0.19	0.12	0.01	0.02	1.73	99.31	143.0	20.0	150	0.99	3.89	2.50	1.15	19.1	4.43	2.4
GM5-3	0.17	0.08	0.02	0.01	1.49	98.36	88.5	19.8	430	0.30	3.25	1.72	0.92	16.1	3.60	2.1
GM7-1	0.30	0.15	0.02	0.05	0.97	101.37	430.0	20.4	180	0.29	4.98	3.34	1.23	19.9	5.07	3.0
GM8	0.17	0.10	0.02	0.01	0.68	98.68	131.5	18.1	160	0.53	4.57	2.67	0.99	17.8	4.24	2.5
GM11	0.19	0.12	0.02	0.10	1.93	100.26	896.0	25.2	110	0.75	4.07	2.43	1.20	19.6	4.19	2.8
LR2	0.18	0.06	0.02	0.02	2.09	99.47	157.5	11.1	190	0.16	2.44	1.66	0.71	16.1	2.49	1.5
LR5-6	0.18	0.20	0.02	0.02	1.66	98.63	204.0	42.9	120	0.34	5.53	3.26	1.64	19.9	6.31	4.0
LR6-1	0.19	0.38	0.02	0.03	0.97	99.29	245.0	69.2	70	0.34	6.69	4.10	2.23	21.8	7.80	5.3
LR8-7	0.28	0.17	0.02	0.08	1.93	99.46	672.0	25.4	110	1.58	4.22	2.32	1.36	17.6	4.55	2.4
LR16-2	0.21	0.13	0.03	0.03	1.91	99.53	269.0	28.3	120	0.50	4.06	2.10	1.22	19.6	4.25	2.7
<i>Dykes subparallel to 2.1 Ga dykes</i>																
LR11-6	0.23	0.28	0.02	0.03	1.52	98.23	314.0	59.6	80	1.31	7.65	4.63	2.37	21.2	8.41	5.4
GM15-3	0.13	0.11	0.02	0.01	2.47	99.86	47.0	30.5	40	1.72	4.16	2.17	1.07	18.5	4.46	3.2
GM16	0.19	0.09	0.02	0.02	1.77	99.32	220.0	28.2	50	1.26	3.46	2.19	1.14	19.5	4.25	3.0
LR1-8	0.19	0.07	0.02	0.09	2.44	99.72	810.0	9.7	250	0.09	2.48	1.60	0.70	15.8	2.64	1.3
<i>Karelia</i>																
15SD_082	0.25	0.10	0.01	0.01	-0.04	100.21	53.5	12.5	50	0.03	4.05	2.74	1.16	18.3	3.80	2.2
15SD_092	0.18	0.16	0.04	0.02	2.02	98.90	185.5	28.7	600	0.25	4.04	2.17	1.50	22.0	4.94	3.6
16SD_019A	0.25	0.09	0.01	0.01	0.97	100.05	54.3	8.5	210	0.09	5.13	3.42	0.93	17.3	3.89	2.0
10-LT-05 Std – A	0.14	0.38	0.11	0.06	0.64	101.18	547	49.7	190	0.3	3.7	1.97	1.66	20.5	4.61	3.5
10-LT-05 Std – B	0.14	0.37	0.11	0.06	0.66	100.36	546	47.3	180	0.2	3.68	2.01	1.47	19.1	4.54	3.7
10-LT-05 Std – C	0.14	0.37	0.11	0.06	0.59	99.63	539	45.8	180	0.2	3.7	2.03	1.54	19.5	4.5	3.5
10-LT-05 Std – Avg	0.14	0.38	0.11	0.06	0.75	100.27	538.3	48.0	184	0.23	3.63	1.97	1.59	19.9	4.45	3.6
2σ	0.00	0.02	0.00	0.00	0.34	1.69	36.6	3.9	23	0.06	0.29	0.15	0.16	1.4	0.44	0.3

Table 5-3 continues on the next page

Notes: Major elements in wt. % and trace elements in ppm. 10-LT-05 is an internal basalt standard used at Carleton University. Analytical results for three batches A, B and C are provided. Superscripts A, B or C following sample numbers identify corresponding batches. The average composition and 2σ is presented for 25 analyses spanning 6 years. 1. The “Granary Lake dyke”; 2. Site 183 from Halls *et al.* (2008).

Table 5-3. Continued.

Sample Detection limit	Ho 0.01	La 0.5	Lu 0.01	Nb 0.2	Nd 0.1	Pr 0.03	Rb 0.2	Sm 0.03	Sn 1	Sr 0.1	Ta 0.1	Tb 0.01	Th 0.05	Tm 0	U 0.05
<i>Superior</i>															
BNB-12-021A ^{1A}	0.73	6.9	0.33	4.2	9.9	2.14	24.1	3.04	1	160.5	0.3	0.57	1.36	0.32	0.47
BNB-12-021B ^{1A}	0.76	8.5	0.29	4.6	11.2	2.62	36.0	2.80	1	237.0	0.3	0.57	1.87	0.30	0.56
BNB-12-022 ^A	1.41	34.3	0.60	17.8	36.6	9.16	46.2	7.34	2	227.0	1.1	1.11	5.92	0.55	0.92
BNB-12-023A ^A	0.96	7.6	0.39	5.9	12.9	2.66	65.7	3.91	1	226.0	0.4	0.76	0.73	0.41	0.21
BNB-12-023B ^A	0.94	7.2	0.37	5.7	12.5	2.59	68.0	3.90	1	203.0	0.4	0.80	0.76	0.40	0.19
BNB-12-024 ^A	0.88	6.6	0.33	5.7	11.5	2.46	66.5	3.26	1	198.0	0.3	0.70	0.69	0.34	0.18
BNB-12-027 ^A	0.66	7.5	0.29	5.9	9.7	2.27	27.0	2.27	1	175.0	0.4	0.45	0.93	0.29	0.25
BNB-12-028A ^A	1.26	33.8	0.50	16.5	36.8	8.88	107.0	6.86	2	242.0	0.9	1.00	5.65	0.54	0.85
BNB-12-028B ^A	1.13	26.6	0.47	13.8	30.4	7.39	32.5	5.71	1	216.0	0.8	0.88	4.68	0.48	0.75
BNB-14-012A ^A	1.10	15.2	0.47	13.5	20.2	4.56	15.8	4.48	1	191.0	0.8	0.81	1.77	0.45	0.40
BNB-14-019C ^{2A}	1.12	11.6	0.50	9.6	17.6	3.84	11.9	4.05	1	163.0	0.6	0.81	1.47	0.51	0.37
BNB-14-019D ^{2A}	1.08	11.1	0.46	8.8	16.1	3.63	11.4	3.86	1	159.0	0.5	0.77	1.40	0.44	0.35
BNB-14-019E ^{2A}	1.06	12.3	0.44	9.0	16.6	3.84	20.3	3.98	2	185.5	0.5	0.77	2.34	0.44	0.55
LL20-3	1.04	21.3	0.47	22.6	24.7	5.68	38.9	4.94	2	233.0	1.3	0.81	3.27	0.42	0.65
TK48-19	1.54	36.4	0.57	26.3	43.0	9.96	30.3	9.40	2	197.5	1.7	1.30	4.68	0.66	0.83
DP1-5	0.68	11.5	0.29	7.6	14.5	3.21	21.6	3.24	1	215.0	0.4	0.52	1.27	0.31	0.19
MK7-8	0.62	11.8	0.29	7.5	14.0	3.26	11.6	3.25	1	178.0	0.4	0.57	1.51	0.27	0.28
DP2-2	0.97	18.3	0.40	11.9	20.2	4.85	26.5	4.85	1	166.5	0.6	0.83	3.13	0.45	0.54
<i>Wyoming</i>															
95SP-2	0.91	17.6	0.32	15.9	23.2	5.42	22.9	5.50	2	254.0	0.9	0.72	1.85	0.3	0.39
95SP-2*	0.87	18.7	0.32	16.0	25.1	5.48	22.6	5.48	2	254.0	0.9	0.84	1.93	0.3	0.41
BM1-4	1.46	17.8	0.64	20.9	26.6	5.67	20.0	6.62	2	205.0	1.0	1.15	1.69	0.6	0.56
BM3	1.14	22.5	0.44	27.3	28.4	6.51	20.1	6.22	2	236.0	1.5	1.00	3.10	0.5	0.69
SM3	0.89	11.5	0.36	8.4	16.2	3.43	58.2	3.85	1	227.0	0.5	0.74	1.76	0.4	0.53
SM4	1.03	12.1	0.43	7.4	16.2	3.61	30.0	4.22	1	162.0	0.4	0.75	1.37	0.4	0.46
SM6-3	1.36	25.0	0.56	13.5	29.4	6.86	76.7	6.24	2	171.0	0.9	1.07	4.13	0.6	1.21
SM7-3	0.83	8.7	0.36	5.1	11.8	2.43	41.5	3.03	3	142.5	0.2	0.66	1.37	0.4	0.37
GM5-3	0.60	8.6	0.19	5.7	11.9	2.68	25.4	3.41	1	152.5	0.3	0.54	1.35	0.3	0.75
GM7-1	1.05	7.8	0.45	5.5	13.8	2.85	61.2	4.11	1	175.0	0.3	0.83	1.30	0.5	0.37
GM8	0.89	7.5	0.41	6.8	11.8	2.55	17.8	3.27	1	142.0	0.4	0.68	1.37	0.4	0.39
GM11	0.82	11.2	0.30	8.0	15.8	3.40	70.8	4.00	1	199.0	0.4	0.65	1.90	0.4	0.45
LR2	0.57	4.8	0.23	2.7	7.5	1.55	40.5	2.00	1	110.5	0.1	0.39	1.06	0.3	0.23
LR5-6	1.16	20.3	0.47	15.6	24.2	5.44	54.3	5.39	1	128.5	0.9	0.94	2.96	0.5	0.84
LR6-1	1.40	33.0	0.51	24.9	37.1	8.70	37.3	8.30	1	152.0	1.4	1.15	3.24	0.6	0.72
LR8-7	0.79	11.4	0.36	12.5	17.1	3.36	58.6	3.56	1	161.5	0.8	0.64	1.49	0.3	0.45
LR16-2	0.72	13.5	0.29	14.0	16.4	3.52	33.8	3.62	1	222.0	0.8	0.69	2.05	0.3	0.53
LR11-6	1.50	28.9	0.64	26.4	33.9	8.05	74.5	6.88	2	207.0	1.5	1.24	3.88	0.7	1.18
GM15-3	0.84	13.5	0.33	6.6	18.0	4.01	85.6	4.16	2	181.5	0.4	0.71	3.20	0.3	2.3
GM16	0.74	12.4	0.27	6.2	16.1	3.67	86.7	3.57	1	195.5	0.4	0.66	2.82	0.3	1.16
LR1-8	0.62	4.1	0.27	2.5	6.5	1.45	35.0	1.86	1	129.0	0.1	0.42	0.67	0.3	0.12
<i>Karelia</i>															
15SD_082 ^B	0.98	5.2	0.37	4.4	9.6	1.95	2.3	2.83	1	125.5	0.3	0.68	0.36	0.4	0.09
15SD_092 ^B	0.77	12.5	0.28	10.4	18.5	4.15	24.3	4.57	1	283.0	0.7	0.72	2.33	0.3	0.54
16SD_019A ^A	1.16	3.1	0.51	2.9	7.6	1.31	5.1	2.77	1	95.1	0.3	0.79	0.24	0.5	0.09
10-LT-05 Std – A	0.75	23.0	0.25	11.5	26.7	6.76	13.0	4.82	2	948	0.6	0.68	2	0.29	0.57
10-LT-05 Std – B	0.75	21.7	0.24	10.5	25.6	6.15	12.1	5.35	1	894	0.6	0.66	2	0.27	0.69
10-LT-05 Std – C	0.67	22.0	0.27	10.8	24.7	6.08	12.0	5.20	1	904	0.6	0.66	2	0.27	0.56
10-LT-05 Std – Avg	0.73	22.4	0.25	10.9	25.6	6.21	12.5	5.15	1.2	938.2	0.6	0.66	1.99	0.3	0.59
2σ	0.07	2.0	0.02	0.9	2.0	0.66	0.8	0.51	0.8	38.5	0.1	0.06	0.22	0	0.1

Table 5-3 continues on the next page

Notes: Major elements in wt. % and trace elements in ppm. 10-LT-05 is an internal basalt standard used at Carleton University. Analytical results for three batches A, B and C are provided. Superscripts A, B or C following sample numbers identify corresponding batches. The average composition and 2σ is presented for 25 analyses spanning 6 years.

1. The “Granary Lake dyke”; 2. Site 183 from Halls *et al.* (2008).

Table 5-3. Continued.

Sample Detection limit	V 5	W 1	Y 0.5	Yb 0.03	Zr 2	Ag 0.5	As 5	Cd 0.5	Co 1	Cu 1	Li 10	Mo 1	Ni 1	Pb 2	Sc 1	Tl 10	Zn 2
<i>Superior</i>																	
BNB-12-021A ¹	332	<1	19.8	2.09	73	<0.5	<5	0.5	49	77	10	<1	66	<2	40	<10	84
BNB-12-021B ¹	298	<1	19.2	1.95	73	<0.5	<5	<0.5	47	71	20	1	62	<2	36	<10	77
BNB-12-022	400	<1	36.9	3.74	197	<0.5	<5	<0.5	57	40	10	1	37	<2	42	10	115
BNB-12-023A	400	<1	25.6	2.65	93	<0.5	<5	0.6	54	95	20	1	85	<2	42	10	98
BNB-12-023B	391	<1	25	2.57	93	<0.5	6	<0.5	52	90	30	<1	88	<2	43	<10	99
BNB-12-024	386	<1	22.8	2.24	84	<0.5	<5	0.5	53	89	20	1	91	<2	43	<10	102
BNB-12-027	295	<1	17.3	1.75	50	<0.5	7	0.6	42	67	20	<1	91	<2	41	10	69
BNB-12-028A	358	4	35.2	3.28	173	<0.5	<5	0.6	55	51	20	1	38	4	39	<10	126
BNB-12-028B	372	<1	31.1	2.93	148	<0.5	6	<0.5	53	18	20	1	38	<2	39	<10	109
BNB-14-012A	527	<1	30.3	2.98	113	<0.5	<5	0.5	66	135	10	<1	73	<2	45	<10	136
BNB-14-019C ²	412	<1	31.5	3.20	94	<0.5	<5	0.6	55	178	10	1	80	<2	50	<10	120
BNB-14-019D ²	404	<1	29.8	3.05	87	<0.5	<5	0.6	56	169	10	<1	85	<2	51	<10	115
BNB-14-019E ²	380	1	28.7	2.98	85	<0.5	<5	0.6	49	155	10	<1	73	<2	46	<10	103
LL20-3	329	<1	27.9	3.10	164	<0.5	<5	<0.5	38	140	10	1	39	<2	31	10	114
TK48-19	425	1	41.9	4.05	241	<0.5	<5	0.8	55	64	10	1	54	5	39	<10	174
DP1-5	294	<1	18.1	1.87	72	<0.5	5	<0.5	46	86	30	1	63	2	39	<10	85
MK7-8	263	<1	17.3	1.81	81	<0.5	<5	0.5	52	80	10	1	74	<2	37	<10	99
DP2-2	354	1	26.5	3.00	120	<0.5	<5	<0.5	53	84	10	1	53	2	41	<10	99
<i>Wyoming</i>																	
95SP-2	402	1	22.9	2.29	147	<0.5	<5	0.5	54	145	20	1	83	<2	34	<10	166
95SP-2*	371	1	23.2	2.1	147	<0.5	8	0.5	55	143	20	1	81	<2	34	<10	161
BM1-4	466	1	38.6	3.89	189	<0.5	5	0.5	49	74	10	<1	62	2	41	<10	147
BM3	378	1	31.9	3.25	187	<0.5	<5	0.6	47	177	10	1	55	<2	40	<10	128
SM3	320	<1	23.4	2.61	113	<0.5	8	<0.5	42	115	10	1	61	3	37	<10	71
SM4	350	1	26.9	2.80	112	<0.5	9	0.5	46	132	20	1	64	<2	36	<10	147
SM6-3	397	1	37.8	4.17	207	<0.5	<5	<0.5	30	49	10	1	46	<2	38	<10	49
SM7-3	320	1	22.9	2.28	88	<0.5	5	<0.5	52	157	10	<1	102	<2	37	<10	100
GM5-3	251	<1	16.8	1.62	85	<0.5	7	<0.5	41	37	10	<1	74	<2	34	<10	72
GM7-1	395	1	29.2	3.26	124	<0.5	<5	0.5	49	75	10	<1	88	<2	35	<10	126
GM8	317	<1	24.5	2.51	93	<0.5	7	<0.5	47	90	10	<1	77	2	48	<10	87
GM11	318	<1	21.5	2.11	109	<0.5	6	<0.5	49	120	20	<1	76	<2	39	<10	84
LR2	264	1	15.3	1.54	61	<0.5	<5	<0.5	40	99	10	<1	96	<2	39	<10	92
LR5-6	410	1	32.1	3.69	157	<0.5	<5	<0.5	42	32	10	<1	51	3	43	<10	64
LR6-1	331	1	35.9	3.98	217	<0.5	<5	<0.5	39	41	10	1	34	3	39	<10	51
LR8-7	437	<1	21.2	2.44	95	<0.5	5	<0.5	60	77	10	1	61	31	53	<10	169
LR16-2	377	1	20.4	2.21	100	<0.5	6	<0.5	53	165	10	<1	87	<2	41	<10	106
<i>Dykes subparallel to 2.11 Ga dykes, Wyoming</i>																	
LR11-6	417	2	42.8	4.65	217	<0.5	<5	0.5	45	116	10	1	44	3	40	10	144
GM15-3	279	1	22.7	2.25	120	<0.5	6	<0.5	29	25	30	<1	20	10	33	<10	45
GM16	270	1	20.6	2.12	111	<0.5	9	<0.5	41	48	20	<1	28	3	34	<10	47
LR1-8	287	<1	15.6	1.63	50	<0.5	5	0.5	43	132	10	<1	105	<2	43	<10	90
<i>Karelia</i>																	
15SD_082	280	<1	23.4	2.72	77	<0.5	<5	0.5	58	180	10	<1	58	2	42	<10	123
15SD_092	362	<1	19.5	1.85	117	<0.5	<5	<0.5	51	60	20	<1	276	2	26	<10	98
16SD_019A	364	1	31.5	3.48	67	<0.5	<5	0.8	58	137	20	<1	85	5	52	<10	109
10-LT-05 Std – A	173	<1	19.5	1.72	155	<0.5	<5	<0.5	42	43	10	1	124	4	18	<10	100
10-LT-05 Std – B	159	1	18.6	1.56	151	<0.5	<5	0.6	38	42	10	2	114	6	18	<10	94
10-LT-05 Std – C	154	1	17.4	1.64	137	<0.5	<5	<0.5	39	41	10	<1	116	<2	17	<10	91
10-LT-05 Std – Avg	163	1	18.64	1.66	150.4	NA	7	1	39	42	10	1	115	4	17	10	92
2σ	21.7	0	1.48	0.21	14.05	NA	1	0	4	4	0	1	9	3	2	0	7

Notes: Major elements in wt. % and trace elements in ppm. 10-LT-05 is an internal basalt standard used at Carleton University. Average composition and 2σ is presented for 25 analyses spanning 6 years. 1. The “Granary Lake dyke”; 2. Site 183 from Halls *et al.* (2008).

Table 5-4. Sm-Nd isotopic data for 2.11 Ga dykes from the Superior, Wyoming and Karelia-Kola cratons.

Craton, Sample	Age (Ma)	Geochemical group	Nd (ppm)	Sm (ppm)	$^{143}\text{Nd}/^{144}\text{Nd}_m$	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}_i$	ϵNd_i
<i>Superior</i>								
TK48-19	2104	3	44.38	9.62	0.511646	0.1310	0.509831	-1.6
MK7-8	2106	3	15.31	3.53	0.511780	0.1395	0.509845	-1.3
DP2-2	2109	3	22.57	5.24	0.511797	0.1403	0.509848	-1.1
DP1-5	2112	3	17.74	4.19	0.511846	0.1428	0.509861	-0.8
BNB-14-019C (183)	(Site 2121)	3	15.98	4.23	0.512119	0.1599	0.509885	-0.1
LL20	2125	2	25.07	5.70	0.511875	0.1374	0.509952	+1.3
MA9-1-1	*	3	21.17	5.05	0.511807	0.1443	0.509797	-2.0
MA22-2-1	*	3	29.90	6.34	0.511465	0.1282	0.509679	-4.3
MA4-3-1	*	3	47.78	10.14	0.511590	0.1283	0.509803	-1.9
MD31-2	*	3	36.21	7.67	0.511647	0.1281	0.509852	-0.9
MA19-1-1	*	3	10.48	2.80	0.512293	0.1617	0.510041	+2.8
MA18-1-1	*	3	16.77	4.26	0.512003	0.1536	0.509864	-0.7
MD1-1-1A	*	2	19.40	4.44	0.511899	0.1383	0.509972	+1.4
<i>Wyoming</i>								
95SP-2	2106	3	25.54	5.88	0.511887	0.1391	0.509958	+0.9
GM11	2111	3	15.28	3.89	0.511912	0.1538	0.509773	-2.6
GM7-1	*	3	14.30	4.12	0.512291	0.1741	0.509866	-0.6
LR1-8	*	3	10.11	3.00	0.512444	0.1797	0.509941	+0.8
GM8	*	3	11.88	3.37	0.512275	0.1713	0.509888	-0.2
SM3	*	3	17.33	4.46	0.512030	0.1556	0.509863	-0.7
BM1-4	2113	1B	24.21	6.26	0.512100	0.1562	0.509927	+0.5
LR6-1	*	3	36.79	7.91	0.511579	0.1299	0.509769	-2.5
BM3	*	2	27.22	6.32	0.511934	0.1405	0.509978	+1.5
GM5-3	*	3	15.01	3.81	0.511976	0.1534	0.509839	-1.2
LR11-6	*	3	32.66	7.48	0.511823	0.1384	0.509896	0.0
<i>Karelia</i>								
15SD_082	2114	1A	11.05	3.36	0.512604	0.1837	0.510045	+2.9
15SD_092	2128	3	23.76	5.97	0.512023	0.1518	0.509895	+0.3
16SD_019A	2135	1A	7.74	2.80	0.513065	0.2189	0.509987	+2.2

$^{143}\text{Nd}/^{144}\text{Nd}_m$ is the measured present-day value and $^{143}\text{Nd}/^{144}\text{Nd}_i$ is the initial ratio.

ϵNd_i values are calculated using known crystallization ages when available (as shown in Age Ma column). When a crystallization age is not available, ϵNd_i is calculated based on an estimated at 2115 Ma, as indicated by * in column Age (Ma).

Isotope ratios are normalized to $^{146}\text{Nd}/^{144}\text{Nd}=0.72190$.

2 σ uncertainties of $^{143}\text{Nd}/^{144}\text{Nd}_m$ are 0.000011 based on reproducibility of the in-house Nd standard.

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

5.5.2.2 Nd isotopes

Dykes from Karelia, Superior and Wyoming have a large range in $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values from +2.9 to -4.3 (**Figure 5-9d**) but their ranges are generally consistent within each of the three defined multi-element patterns. The highest $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values are from +2.2 to +2.9 and are found in the MORB-like dykes, unique to the Karelia-Kola craton. Pattern 2 geochemical signatures have a range of $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values from +1.4 to +0.5. The largest range in $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values is found in dykes with geochemical signatures part of pattern 3, from +2.8 to -4.3.

5.6 Discussion

The Superia supercraton reconstruction of Bleeker and Ernst (2006), and more recent iterations place the Superior, Wyoming, Karelia-Kola and Hearne cratons adjacent to one another (Ernst and Bleeker, 2010; Söderlund *et al.*, 2010; Pisarevsky *et al.*, 2015; see also Roscoe and Card, 1993; Heaman, 1997). Their placement and “nearest neighbour” relationship is largely based on matching magmatic barcodes (Ernst and Bleeker, 2010). A natural test for the Superia reconstruction, therefore, is a comparison of geochemical signatures of contemporaneous dyke swarms and sill provinces from each craton. Such a test is presented below for dykes dated at 2.11 Ga and, with the establishment of a shared magmatic history, provides an opportunity to assess plumbing system models and sources (e.g., Kastek *et al.*, 2018; Rogers *et al.*, 2018; Ernst *et al.*, 2019; Mackinder *et al.*, 2019). The major, trace element and Nd isotope geochemistry from this study are combined with additional analyses from the Tohmajärvi-Pirtguba dykes, Misi gabbro sills and Jouttiaapa

and Hirsimaa mafic volcanic rocks from the Karelia-Kola craton (Niiranen et al., 2003, 2005; Stepanova and Stepanov, 2010; Stepanova et al., 2014; GTK, 2018), Marathon dykes from the Superior craton (Halls et al., 2008), and Kazan dykes and Griffin sills from the Hearne craton (Aspler et al., 2002; Ernst and Buchan, 2010). The Chipman dykes from the Hearne craton have been omitted from this analysis due to their amphibolite facies metamorphic grade and low Nd concentrations (Williams et al., 1995; Flowers et al., 2006). Low Nd concentrations paired with high degrees of metamorphism and migmatitic textures are an indication that the trace element concentrations and Nd isotope values may have been changed during metamorphism and do not necessarily represent magmatic compositions.

5.6.1 Geochemical comparison of 2.11 Ga mafic rocks across cratons

The geochemical comparison of 2.11 Ga mafic rocks from the Superior, Karelia-Kola, Wyoming and Hearne cratons reveal multiple geochemical groups within swarms on individual cratons that match either to all or some subgroups identified in dykes, sills and/or mafic volcanic rocks in other cratons. The geochemical groups discussed here include the patterns recognized in samples analyzed as part of this study and integrated with previous geochemical studies (Nykänen *et al.*, 1994; Aspler *et al.*, 2002; Stepanova *et al.*, 2014). The model presented identifies two end-member magmas in the Tohmajärvi-Pirtguba dykes and Jouttiaapa mafic volcanic rocks in the Karelia-Kola craton, and the Griffin sills, Hearne craton and one dyke from the Wyoming craton, that mixed together and then assimilated a crustal component and fractionated prior to final emplacement. The rare mixed magmas are found in the Superior, Wyoming and Karelia-Kola craton whereas the

most abundant magmas, those that underwent AFC, are recognized in units within all cratons that are part of this study.

5.6.1.1 Group 1: Source magmas

The major elements characteristics of the samples indicate that the primary magmas are tholeiitic basalts (**Figure 5-6 to Figure 5-9**). Correlations between Mg number and FeO, NaO, CaO/Al₂O₃ show that the primary magma was crystallizing three phases including olivine, plagioclase and clinopyroxene. Magmas most representative of the source are further identified based on their apparent lack of crustal interaction. Crustal contamination is typically indicated by a negative Nb anomaly on multi-element plots (normalized to primitive mantle; Sun and McDonough, 1989) and this can be quantified by with Nb/Nb* values ($Nb/Nb^* = Nb_{PM} / [(Th_{PM} \times La_{PM})^{0.5}]$; following Stepanova et al. 2014) or by ΔNb values ($\Delta Nb = 1.74 + \log(Nb/Y) - 1.92 \log(Zr/Y)$; following Fitton *et al.*, 1997) that are less than one. Conversely, source magmas have Nb values similar to La and Th on multi-element plots (normalized to primitive mantle) and Nb/Nb* values greater than one. The identified source magmas include a subset of the Tohmajärvi-Pirtguba dykes and Jouttiaapa mafic volcanic rocks, Karelia-Kola craton, one dyke (BM1-4) from the Bear Mountain swarm, Wyoming craton, and a subset of the Griffin sills, Hearne craton (**Figure 5-11a**).

The Tohmajärvi-Pirtguba dykes categorized as Group 1A have flat patterns comparable to primitive mantle on a multi-element diagram with $\epsilon Nd_{T=2.1Ga}$ values ranging from +2.2 to +2.9 and include the dated Nieminen dyke at 2114.4 ± 3 Ma (**Figure 5-11a**). The Jouttiaapa mafic volcanic rocks, located in the Peräpohja schist belt, have almost identical trace element signatures but include a La peak and Ce trough, and have a larger

$\epsilon\text{Nd}_{T=2.1\text{Ga}}$ range from +3.1 to +4.8 (Huhma *et al.*, 1990). The petrogenetic studies by Nykänen *et al.* (1994) and Stepanova *et al.* (2014) on the MORB-type dykes concluded that these magmas were generated in an extensional tectonic environment at shallow levels within the spinel to spinel-garnet stability fields which caused 10-15% partial melting of the upper mantle. Variations in the depth of melting explain the variation in Tb/Yb ratios in **Figure 5-11b**. The undepleted multi-element signature (shown by the similarity of Group 1A to primitive mantle) and $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values ranging from +2.2 to +2.9, indicate that the mantle source of these magmas was not generated from depleted mantle (depleted mantle would have an $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ of +5.0). Instead, the mantle source may either have been derived from a less depleted (more chondritic) source or may have been enriched by a recycled ocean (pyroxenite) component. Enrichment from a pyroxenite component would draw the $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values down to those that are lower than +5 without drastically modifying relative abundances of Th_{PM} , Nb_{PM} and La_{PM} .

With distinctly lower $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values ranging from +0.5 to +1.0, Group 1B source magmas include one dyke (BM1-4) from the Wyoming craton and many of the Griffin sills from the Hearne craton (**Figure 5-11a**). Dyke BM1-4 from the Wyoming craton has an E-MORB signature on a multi-element diagram with a flat HREE slope, shallow negative LREE slope, a Nb peak and with $\text{Th} < \text{La}$. The OIB-like Griffin sills, in contrast, have steep negative HREE and LREE slopes which are attributed to enriched asthenosphere akin to modern ocean island basalts, possibly with a mantle plume origin (Aspler *et al.*, 2002). With similar $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values, derivation of the OIB-like and E-MORB patterns associated with Group 1B can be explained by varying degrees of partial melting and melting at varying depths. For example, the flat Tb/Yb slope and shallow La/Sm slope characteristic

of the E-MORB-like BM1-4 dyke, may be explained by having been derived from the same source as the OIB-like basalts of the Griffin sills but having undergone a higher degree of partial melting at a shallower depth above the garnet-spinel boundary.

5.6.1.2 Group 2: Mixed magmas

Representatives with Group 2 geochemistry are found in dykes and the Hirsimaa mafic volcanic rocks of Karelia-Kola, as well as dykes from the Wyoming and Superior cratons (**Figure 5-11c**). Group 2 is characterized by shallow negative HREE and LREE slopes, Th abundances $\leq$ La and $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values between +1.3 and +1.5. Similar to Group 1, these magmas lack a Nb anomaly, but have intermediate multi-element patterns and isotope characteristics between Groups 1A and 1B. A bulk mixing line is shown on a Nb/Yb versus Tb/Yb plot (**Figure 5-11b**) between the MORB and OIB end-members identified in Group 1. Group 2 magmas plot along the bulk mixing line between the MORB and OIB end-members suggesting that they may be a hybrid that formed during melt transport within the broader magmatic system. Their identification as a mixture of Group 1A and 1B source magmas is further supported by bulk mixing curves comparing $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ against Tb/Yb, La/Yb and Zr/Y ratios (**Figure 5-12a**). Note that two curves are calculated for each plot to represent the highest and lowest $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values in the Tohmajärvi-Pirtguba dykes in order to approximate the range of possible isotope values. In each case, Group 2 samples sit along the bulk mixing curves in support of having been derived by mixing of Group 1A and 1B magmas.

5.6.1.3 Group 3: Contaminated magmas

As the most abundant signature, Group 3 geochemistry is identified in representatives from all four cratons studied. In the Karelia-Kola craton, dykes and Jouttiaapa mafic volcanic rocks with this geochemistry have been previously classified as Fe-tholeiitic basalts (Vuollo *et al.*, 1992; Nykänen *et al.*, 1994; Perttunen and Hanski, 2003). The multi-element patterns (normalized to primitive mantle) for Group 3 include steep to shallow negative HREE slopes, steep negative LREE slopes, negative Nb anomalies of variable magnitude and Th abundances that are typically greater than La (**Figure 5-11e**). **Figure 5-11d** shows a Th/Yb versus Nb/Yb plot (after Pearce, 2008) in which dykes and sills with Group 3 geochemistry plot above the MORB-OIB array.

The geochemical characteristics of Group 3 are hallmark signs of magmas that have interacted with crust during their petrogenesis (Nykänen *et al.*, 1994). Assimilation-fractional crystallization (AFC) curves are calculated starting from Group 1 MORB and OIB magmas each trending towards crustal end-members represented by the 3.5 Ga Siurua gneiss and a trondhjemite from Karelia-Kola. The AFC calculations follow DePaolo (1981) where r (the ratio of the rate of assimilation to rate of fractional crystallization) is constant and equal to 0.3, partition coefficients used for these calculations are those from Pearce (2008) and crustal end-members are from Mutanen and Huhma (2003). Additional AFC curves have been calculated for $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ as well as trace element ratios Th/Yb, Nb/Nb* and La/Sm (**Figure 5-12d, e, and f**). Together, **Figure 5-11d** and **Figure 5-12d-f**, show that Group 3 dykes and sills follow AFC trends closely up until $F=0.7$ and support an overall petrogenetic history that involved crustal assimilation incorporating variable contaminants. The most negative samples part of Group 3 can be explained by assimilation

of very old components (i.e., *ca.* 3.5 Ga Siurua gneiss; Mutanen and Huhma, 2003) whereas those with less negative $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ values may have incorporated younger, *ca.* 2.7 – 2.9 Ga TTG granitoids, common amongst Archean cratons.

The negative correlation of $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ against SiO_2 wt. % provides additional evidence that Group 1 and 2 magmas assimilated a crustal component. Generation of Group 3 magmas are explained by the source magmas and their hybrids, Groups 1 and 2 respectively, having been given time to assimilate a crustal component within staging chambers at or just above the sub-continental lithospheric mantle. A thermal heat source from a mantle plume would help drive assimilation of crust.

Figure 5-11. *On the next page.* Flow chart showing the geochemical evolution of the *ca.* 2.11 Ga magmatic system. For panels a, c, and e, dykes and sills are represented by solid lines and mafic volcanic rocks are represented by dashed lines. For panels b and d, dykes and sills are represented by darker shades and mafic volcanic rocks by lighter shades. The geochemical data included are a compilation of Huhma *et al.* (1990); Aspler *et al.* (2002); Niiranen *et al.* (2003, 2005); Halls *et al.* (2008); Ernst and Buchan (2010); Stepanova and Stepanov (2010); Stepanova *et al.* (2014, 2015); GTK (2018) and this study. See also Toivola (1988); Toivola *et al.* (1991); Nykänen *et al.* (1994); Vuollo and Huhma (2005). A. Multi-element patterns for end-member magmas. Group 1A is a subset from the Tohmajärvi-Pirtguba dykes. Group 1B includes one dyke from the Bear Mountain dykes, Wyoming and a subset of sills from the Griffin Gabbros, Hearne craton. B. Nb/Yb versus Tb/Yb plot with a bulk mixing curve between end-member magmas Group 1A and 1B. Intervals marked every 10%. C. Multi-element patterns for Group 2, the mixture of Group 1A (sample 15SD_082) and B (sample 98-22-2 from Aspler *et al.*, 2002). D. Th/Yb versus Nb/Yb plot. Diagram and partition coefficients from (Pearce, 2008). VAA: Volcanic arc array. AFC calculations following DePaolo (1981) in which $r=0.3$. Siurua gneiss sample A1602 from Mutanen and Huhma (2003), Trondhjemite sample A1909 from Mikkola *et al.* (2011). E. Multi-element patterns of Group 3 magmas which were produced by AFC. All multi-element patterns are normalized to primitive mantle (Sun and McDonough, 1989).

5.6.2 Tectonic setting

A tectonic setting in which melting is driven by the thermal pulse and subsequent impingement of a mantle plume can be used to explain the variation in geochemistry observed in the contemporaneous 2.11 Ga dykes (**Figure 5-13**). In the scenario presented in **Figure 5-13**, melts are generated from the plume as well as above, in the overlying lithospheric mantle. Melting of the upper mantle, driven by the thermal influence of the mantle plume, produced the Group 1A MORB-type basalts found as a sub-set of the Tohmajärvi-Pirtguba dykes and Jouttiaapa mafic volcanic rocks. Group 1B OIB-like magmas identified in the Griffin sills are representative of plume-derived melts. Batches of Group 1B also ponded at the base of the crust which resulted in mixing of Group 1A and 1B together in magmatic underplates and ultimately lead to the generation of the rare Group 2 dykes and mafic volcanic rocks. When erupted into the crust, magma with Group 2 geochemistry stalled at various levels and accumulated in staging chambers, where AFC resulted in the generation of Group 3 geochemistry.

Figure 5-13. An illustration of the tectonic setting and plumbing system components for the *ca.* 2.11 Ga magmatism as part of a combined large igneous province event including dykes, sills and mafic volcanic rocks hosted by intracratonic basins after (Ernst *et al.*, 2019). The schematic is an expansion of the model by Nykänen *et al.* (1994) and follows the style of Ernst *et al.* (2019). The occurrence of dykes, sills and mafic volcanic rocks is distinguished by source and petrogenesis related to the geochemical Groups 1A, 1B (end-members), 2 (mixtures of 1A and 1B) and 3 (AFC). Inset shows the configuration of the craton fragments in their ancestral context of supercraton Superia, in which 2.11 Ga magmatism was emplaced. Superior (S), Hearne (H), Wyoming (W) and Karelia-Kola (K) cratons. The approximate profile line A-A' for the block model is shown in red and the blue star in the middle represents the magmatic centre.

5.6.3 Secular geochemical evolution of *ca.* 2.11 Ga dykes

Secular geochemical evolution – geochemical change through time – of the Marathon LIP was first discussed by Halls *et al.* (2008). These authors used their extensive paleomagnetic record to group the dykes belonging to the Marathon, Fort Frances and Kapuskasing swarms into an older sub-group with normal polarity and a younger sub-group with reverse polarity. Based on these parameters, Halls *et al.* (2008) found that the geochemistry of the dykes changed from LREE enriched and highly fractionated in the older dykes (Marathon and Kapuskasing dykes) to uncontaminated MORB-like basalts in the younger Fort Frances dykes (which are not included in this analysis). Older dykes with normal polarity were characterized as Fe-Ti tholeiitic basalts, whereas the younger dykes with reverse polarity were classified as Mg-tholeiitic basalt. Because Mg-tholeiitic basalts cannot be derived by fractionation of Fe-tholeiitic basalts, Halls *et al.* (2008) proposed that these magmas were products of melting and magma fractionation from distinct chambers. To explain the overall evolution of the magmatic system, the diminishing degree of LREE enrichment was attributed to the reduced influence of crustal assimilation and/or progressively thinning lithosphere with higher degrees of partial melting as the magmatic system developed through time.

Studies on the geochemistry of LIPs with well-constrained field and age relationships have found that a LIP may involve more than one geochemical signature. The Siberian and Deccan Traps are examples where magmatism constrained within about a million years (Burgess and Bowring, 2015; Schoene *et al.*, 2019) includes multiple geochemical groups derived from distinct sources (Lightfoot *et al.*, 1990; Lightfoot and Hawkesworth, 1997). The variation in the geochemistry within these LIPs is considered to

represent complexities within their sources (i.e., derived from heterogeneous mantle) and variation within the plumbing system (Ernst *et al.*, 2019) which may result in contemporaneous generation of multiple magma chambers in which magmas undergo variable degrees of assimilation and fractional crystallization (AFC).

Figure 5-14. Geochemistry of samples with U-Pb ages (from **Table 5-1**). Geochemistry data sources as in **Figure 5-11**. Top panel: Nb/Nb^* versus age (Ma). Lower panel: $\epsilon Nd_{T=2.1Ga}$ versus age (Ma).

Expansion of the 2.11 Ga geochemical and geochronological database into the Wyoming, Karelia-Kola and Hearne cratons provide further insight into the complexity of this long-lived (2135-2100 Ma), geochemically diverse magmatic system. The variation

and changes in geochemistry can be explained by an ascending mantle plume and related melts that interacted with a pre-existing extensional tectonic environment. The presence of a pre-existing extensional setting (in place prior to the onset of major 2.11 Ga magmatism) is proposed because the oldest dykes are the Group 1A Tohmajärvi-Pirtguba dykes in the Karelia-Kola craton, which have a total emplacement time spanning 21 Ma, from 2135 Ma to 2114 Ma, are spatially limited and have MORB-like (Group 1A) geochemical signatures (**Figure 5-13**). The petrogenetic studies by Nykänen *et al.* (1994) and Stepanova *et al.* (2014) found that these were generated at shallow depths below thin lithosphere due to their MORB-type, flat REE patterns and near absence of crustal contamination indicators such as negative Nb anomalies accompanied by $\epsilon\text{Nd}_{T=2.1\text{ Ga}}$ values between +2.9 and +2.2. Almost simultaneously, as discussed by Nykänen *et al.* (1994), Group 3 magmas were intruding into thicker lithosphere as dykes. Although plume-derived magmas were not emplaced until *ca.* 2.11 Ga as the Griffin sills (Group 1B; Heaman and LeCheminant, 1993), indicators of a rising plume and related magmas are evidenced by older dykes that are either mixtures of plume-derived melts with MORB-type magmas in which Groups 1A and 1B mixed together to form Group 2, or additionally assimilated crust to form Group 3. The first occurrences of Group 2 at 2125.7 ± 1.3 Ma (LL20, Superior) and Group 3 at 2121 ± 8 Ma (site 183 in the Superior craton) and 2128 ± 1.2 Ma (Sipojuntti dyke in the Karelia-Kola craton) are nearly contemporaneous given that their ages are within analytical error. As this magmatic system progressed and the thermal influence from the plume intensified, the generation of Group 3 from 2128 Ma to 2104 Ma magmas increased to become the most abundant geochemical signature found in dykes across each craton (**Figure 5-14**). A comparison of Nb/Nb* and $\epsilon\text{Nd}_{T=2.1\text{ Ga}}$ versus U-Pb age, while

preliminary, shows a negative correlation in samples with Group 3 geochemistry. This negative correlation suggests that crustal assimilation increased as the magmatic system developed. The assimilation of crust continued until what Halls *et al.* (2008) attributed to a progressive thinning of the lithosphere; culminating in the younger 2.08 Ga Fort Frances dykes, which have flat MORB-type uncontaminated REE patterns.

5.7 Conclusions

Prolific Paleoproterozoic mafic magmatism at 2.11 Ga has been studied using geochronology and geochemistry to test the Superia supercraton paleogeographic reconstruction. The Nieminen dyke, from Western Karelia, is dated at 2114.4 ± 3 Ma and provides further confirmation of an exact age match to dykes and sills in the Superior, Hearne, and Wyoming cratons. A geochemical analysis using whole-rock major and trace element, and Nd isotopes shows that the dykes and sills are sourced from two end-members; Group 1A and 1B. Group 1A involves dykes with MORB-type multi-element patterns with $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ between +2.2 and +2.9 that were generated from shallow melting of an enriched upper mantle likely below thin lithosphere (such that very little crust was assimilated during ascent and emplacement). Group 1B is represented by the Griffin sills, which have OIB-type multi-element patterns and $\epsilon\text{Nd}_{T=2.1\text{Ga}}$ from +0.5 to +1.0 and are characteristic of enriched asthenosphere melts from the garnet stability field; possibly from a plume. Bulk mixing of Groups 1A and 1B created the rare Group 2 dykes found in the Superior, Wyoming and Karelia-Kola cratons. Group 2 geochemical patterns have moderate HREE and LREE slopes, lack a negative Nb anomaly and include $\epsilon\text{Nd}_{T=2.1\text{Ga}}$

values from +1.3 to +1.5. Magmas from the geochemical Groups 1A, 1B and 2 ponded below the lithosphere and accumulated within staging chambers where they underwent assimilation and fractional crystallization, leading to the generation of Group 3. Group 3 magmas have characteristic arc-like signatures with enriched LREEs, negative Nb anomalies and higher Th abundances relative to La. The multiple geochemical matches related by bulk mixing and AFC provide evidence for magmatic evolution of a shared plume-generated LIP now located in the fragmented pieces of the former Superia supercraton including the Superior, Hearne, Wyoming and Karelia-Kola Archean cratons.

Chapter 6

Concluding remarks

6.1 Summary

Information obtained from field observations such as the trend and dip of dykes, U-Pb geochronology, and whole-rock major and trace element and Sm-Nd isotope geochemistry are used to test whether the Karelia-Kola craton was adjacent to the Superior craton within the context of the Superia supercraton paleogeographic reconstruction.

U-Pb ID-TIMS geochronology forms the backbone of this thesis and provides the constraints necessary to interpret field observations and whole-rock major, trace element and Sm-Nd isotope geochemistry. The integration and interpretation of these data support the paleogeographic reconstruction of Superia by Bleeker and Ernst (2006), that was later updated in Ernst and Bleeker (2010). The data collected for this thesis provides further evidence that Karelia-Kola craton was indeed a fragment of the Superia supercraton but modifies its position within the reconstruction. Instead of the Karelia-Kola craton being positioned such that it is adjoined to the southeastern margin of the Superior craton, our data yield a ‘best-fit’ configuration that places the Karelia-Kola craton farther south of the Superior craton. In this reconfiguration, the Karelia-Kola craton is rotated such that the western margin of the Karelia-Kola craton is aligned with and placed adjacent to the eastern margin of the Wyoming craton. This reconfiguration relaxes the tight fit between the Karelia-Kola and Superior cratons based on our findings in Chapter 4 that concluded that

the Nipissing sills did not directly feed the Karjalitic sills. While maintaining and adding (e.g. 2400 Ma du Chef and Uolevinlehto correlations) dyke swarm correlations between the Superior and Karelia-Kola cratons, the repositioning of Karelia-Kola also is in better agreement with the D' VGP of Mertanen et al. (1999).

U-Pb ID-TIMS geochronology is reported for eleven dykes from the Karelia-Kola craton and two sills/intrusions from the Superior craton. The ages obtained in this study have further enhanced correlations to known events in the Superior craton including ages related to a recently recognized match at *ca.* 2.40 Ga between the Karelia-Kola and Superior cratons (first noted by Stepanova et al., 2017). The *ca.* 2.40 Ga Uolevinlehto swarm has been defined by two ages in Western Karelia: the Uolevinlehto dyke at 2386.9 ± 2.4 Ma, and the Viianki dyke at 2400.7 ± 1.8 Ma. These *ca.* 2.40 Ga dykes are contemporaneous with the du Chef dykes in the Superior craton. The *ca.* 2.32 Ga Taivalkoski-Kuito swarm is dated at three localities including the NE-trending Kakikangas dyke at 2329.2 ± 8.2 Ma, NW-trending Tulisaari dyke at 2320 ± 13 Ma, and the WNW-trending Tilsanvaara dyke at 2292 ± 8.8 Ma. A match to these 2.32 Ga dykes craton has not yet been found in the Superior or Wyoming cratons. The *ca.* 2.13-2.10 Ga Marathon dyke swarm in the Superior craton corresponds to the Tohmajärvi-Pirtguba dyke swarm in the Karelia-Kola craton, which has been dated at three dyke localities: the NW-trending Myllykangas dyke at $2135 + 3.6 / - 3.7$ Ma, the NW-trending Sipojuntti dyke at 2128.9 ± 1.5 Ma and the WNW-trending Nieminen dyke at 2114.4 ± 3 Ma. With a tentative age match only to the *ca.* 1.98 Ga Mine Centre dyke in the Superior craton, the Paukkajanvaara dykes (part of the Pechenga-Onega LIP) have been dated directly at two localities: the NW-trending Kivikevätti dyke at 1975.6 ± 9.7 Ma and NW-trending Eno dyke *ca.* 1980 Ma. A

third age for this swarm is obtained from zircons that were picked from the WNW-trending *ca.* 2.32 Ga Tilsanvaara dyke. The concordant zircon fractions yield a weighted mean of 1972.8 ± 11.2 Ma and this is interpreted as a tentative crystallization age for the cross-cutting NW-trending dyke at Tilsanvaara. In thin section, zircons were found in veinlets that can be traced directly to the contact with the younger cross-cutting NW-trending dyke. Two U-Pb ages are presented for mafic magmatic rocks belonging to the *ca.* 2.22 Ga Ungava LIP located in the Superior craton, complementary to the Karjalitic sills in the Karelia-Kola craton, including refined ages for the Triangle Mountain sill at 2216.5 ± 2.1 Ma and the Shakespeare intrusion at $2217.0 \pm 1.7/-1.5$ Ma.

Chapter 2 and **Chapter 3** combine U-Pb ID-TIMS geochronology with dyke dips and trends as a means of distinguishing the regional dyke swarms and to evaluate the influence of Svecofennian orogenic deformation on the various crustal blocks in Western Karelia. In **Chapter 2**, the *ca.* 2.40 Ga Uolevinlehto, *ca.* 2.13 Ga Myllykangas and *ca.* 2.12 Ga Sipojuntti dykes from the Pudasjärvi block show consistent $\sim 30^\circ$ offset from their contemporaneous matches in the adjacent Kuhmo and Taivalkoski blocks. Structural restoration of the Pudasjärvi block, by vertical axis rotation of 30° counter clockwise, provides constraints on regional swarm patterns of the *ca.* 2.45 Ga Pääjärvi, *ca.* 2.40 Ga Uolevinlehto, and 2.11 Ga Tohmajärvi-Pirtguba dyke swarms. This rotation also aligns the *ca.* 1.98 Ga Paukkajanvaara dykes (identified only from aeromagnetic maps) to their known counterpart in the Kuhmo block. In **Chapter 3**, the orientations of dykes dated at *ca.* 2.40, 2.32 and 1.98 Ga are evaluated and integrated with field observations, U-Pb geochronology and aeromagnetic data from dykes in the Taivalkoski block and compared to dykes in adjacent crustal blocks. With a limited data set of three dykes, the EW-trending

ca. 2.40 Ga dykes show no significant difference in trend across the Kuhmo and Taivalkoski blocks. In contrast, the more widely dated *ca.* 2.32 Ga Taivalkoski-Kuio and *ca.* 1.98 Ga Paukkajanvaara dyke swarms show systematic curvature within the Taivalkoski block and at the northern margin of the Kuhmo block (proximal to the Oulujärvi shear zone). This curvature is consistent with deformation associated with sinistral displacement of the Oulujärvi shear zone.

Chapter 4 and **Chapter 5** combine U-Pb geochronology with whole-rock major, trace element and Sm-Nd isotope geochemistry to test whether contemporaneous magmas are geochemically similar. With matching ages confirmed at the Triangle Mountain and Shakespeare intrusion, **Chapter 4** is focused on comparing the geochemistry of the Nipissing sills to their proposed feeder dyke swarms, Senneterre and Maguire. We find that there is a subset of Senneterre and Maguire dykes that are compatible with the main geochemical signature of the *ca.* 2.22 Ga Nipissing sills in terms of whole-rock major and trace element characteristics as well as with respect to $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values. Although the Nipissing sills show a wider range in negative ϵNd values, the $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values of the Senneterre and Maguire dykes are within the upper range of the Nipissing sills. This match indicates that negative $\epsilon\text{Nd}_{T=2.2\text{Ga}}$ values were, at least in part, a characteristic of the magma prior to emplacement, and possibly a signature of the source. A key result from this geochemical study is that, with consistently higher ϵNd values from 0 to +2 (Hanski et al., 2010), the Karjalitic sills in the Karelia-Kola craton are not the ‘downstream’ continuation of the Nipissing sills. **Chapter 5** presents the most extensive geochemical comparison of LIP fragments identified in the Superia supercraton paleogeographic reconstruction. Dykes, sills and mafic volcanic rocks determined to be contemporaneous by either U-Pb

geochronology or paleomagnetism from the Superior, Hearne, Wyoming and Karelia-Kola cratons are compared in terms of their whole-rock major and trace element, and Sm-Nd isotope geochemical characteristics. This geochemical analysis finds that the magmas are sourced from two distinct mantle end-members. The magmas produced from melting of these two sources mixed together and later underwent assimilation and fractional crystallization (AFC) within staging chambers in the lower crust. While some of the end-member magmas are restricted to separate craton fragments (one end-member in Karelia-Kola and the other in the Hearne craton), the products of mixing and AFC are more abundant and found on multiple cratons, thereby providing evidence of a shared petrogenesis relating to the same plume-generated LIP.

6.2 Future work

Following the work presented in this thesis, a number questions are uncovered that should be addressed in the future.

Immediate work is recommended to confirm the age of the zircons from the Tilsanvaara dyke (15SD_076) with in-situ U-Pb geochronology (**Chapter 3**). The younger cross-cutting dyke (15SD_077) should also be dated to confirm the age of the younger dyke and confirm that this cross-cutting dyke is the source for the zircons in the veinlets and the cause of zircon overgrowth on the primary baddeleyite grains.

A number of additional research questions also come out of this project. For example, the ages on the Kivikevätti and Eno dykes could be improved with the addition of carefully picked zircons. These are of interest because they are coeval with the *ca.* 1.98 Ga Mine Centre dyke located in the Superior craton (**Chapter 2**). With current estimations

of break up timing between the Karelia-Kola and Superior craton between 2.1 and 2.0 Ga, it would be interesting to know whether these *ca.* 1.98 Ga contemporaneous events are related and whether this is an indication of later break up than previously proposed.

U-Pb geochronology is recommended on the Palomaa dyke, presently dated with a Sm-Nd whole-rock and mineral internal isochron age of 2077 Ma, to test whether its tentative correlation to the *ca.* 2.08 Ga Fort Frances dyke swarm in the Superior craton is valid (**Chapter 2**). Based on cross-cutting relationships and degree of preservation, the EW dykes found in the Pudasjärvi block are likely younger than 2.11 Ga but without U-Pb ID-TIMS it is impossible to determine whether these dykes have a relevant match within supercraton Superia, making the EW dykes a target for future geochronology.

In the Superior craton, it is recommended that the Nipissing sills part of the high Fe-Ti (high Al) and Low REE subgroups should be dated with U-Pb geochronology (**Chapter 4**). Additionally, a thorough geochemical comparison of whole-rock major, trace, and Sm-Nd isotope geochemistry between the Ungava LIP components (Superior) and the Karjalitic sills (Karelia) is recommended to further explore the nearest neighbour relationship between the Superior and Karelia-Kola cratons (**Chapter 4**).

Lastly, while outside of my own scope of research, paleomagnetic studies should be completed on the *ca.* 1.98 Paukkajanvaara and 2.32 Ga Taivalkoski-Kuito dykes in the Kuhmo and Taivalkoski block to test whether the curvilinear trends of these swarms are due to drag by ductile deformation related to the Oulujärvi shear zone as proposed in **Chapter 3**.

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Statement of co-authorship of Chapter 4

I, Dr. Wouter Bleeker, confirm that I contributed expertise relating to the Nipissing sills, collected samples of Nipissing sills that Sarah Davey crushed for geochemical analysis, co-wrote the geochronology section with Dr. Sandra L. Kamo and provided thorough edits on the manuscript:

Davey, S., Bleeker, W., Kamo, S., Davis, D., Easton, M., and Sutcliffe, R.H., 2019. Ni-Cu-PGE potential of the Nipissing sills as part of the ca. 2.2 Ga Ungava large igneous province; *in* Targeted Geoscience Initiative: 2018 report of activities, (ed.) N. Rogers; Geological Survey of Canada, Open File 8549, p. 403–419.
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I acknowledge that this manuscript is being integrated into the PhD thesis by Sarah Davey.

Dr. Wouter Bleeker

Statement of co-authorship of Chapter 4

I, Dr. Sandra L. Kamo, confirm that I completed the U-Pb geochronology for the Triangle Mountain sill, co-wrote the geochronology section with Dr. Wouter Bleeker, and provided comments on the manuscript:

Davey, S., Bleeker, W., Kamo, S., Davis, D., Easton, M., and Sutcliffe, R.H., 2019. Ni-Cu-PGE potential of the Nipissing sills as part of the ca. 2.2 Ga Ungava large igneous province; *in* Targeted Geoscience Initiative: 2018 report of activities, (ed.) N. Rogers; Geological Survey of Canada, Open File 8549, p. 403–419.
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Dr. Sandra L. Kamo

Statement of co-authorship of Chapter 4

I, Dr. Don Davis, confirm that I completed the U-Pb geochronology on the Shakespeare sill, which is included in the manuscript:

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Dr. Don Davis

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I, Dr. Richard Sutcliffe, confirm that I contributed expertise relating to the Nipissing sills and provided thorough edits on the manuscript:

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Dr. Richard Sutcliffe

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I, Dr. Michael Easton, confirm that I contributed expertise relating to the Nipissing sills and provided thorough edits on the manuscript:

Davey, S., Bleeker, W., Kamo, S., Davis, D., Easton, M., and Sutcliffe, R.H., 2019. Ni-Cu-PGE potential of the Nipissing sills as part of the ca. 2.2 Ga Ungava large igneous province; *in* Targeted Geoscience Initiative: 2018 report of activities, (ed.) N. Rogers; Geological Survey of Canada, Open File 8549, p. 403–419.
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I acknowledge that this manuscript is being integrated in the PhD thesis by Sarah Davey.

Dr. Michael Easton

Appendix B : Methods from Chapter 4

B.1 Geochronology

Sample preparation and analyses were completed at the Jack Satterly Geochronology Laboratory at the University of Toronto. Approximately 10 kg of coarse-grained, pegmatoidal varitextured gabbro of the Triangle Mountain sill and Shakespeare intrusion were crushed and pulverized in a jaw crusher and steel ring mill. Baddeleyite (Triangle Mountain sill) and zircon (Shakespeare intrusion) crystals were collected using a Wilfley table following the method of Söderlund and Johansson (2002). Prior to dissolution, zircon grains from the Shakespeare intrusion were air abraded.

Isotope dilution thermal ionization mass spectrometry (ID-TIMS) was used to analyze U-Pb. Baddeleyite and zircon grains were rinsed with 8N HNO₃ before being loaded into a Teflon capsule spiked with a ²⁰⁵Pb-²³⁵U solution. The grains were dissolved in 0.10 ml of concentrated hydrofluoric acid and 0.02 ml of 7N HNO₃ for 3 to 5 days. Following dissolution, samples were dried down and dissolved in 0.15 ml of 3N HCl for 24 hours (Krogh, 1973). Anion exchange columns (50 µl) were used to extract U and Pb with HCl. Elutions were dried down with H₃PO₄ and loaded onto Re filaments with silica gel (Gerstenberger and Haase, 1997). U-Pb measurements were completed on a VG354 mass spectrometer using a Daly detector in pulse counting mode.

B.2 Geochemistry

Whole-rock major and trace element geochemical analyses were acquired for 25 samples from various Nipissing sill localities. Each sample was crushed with a steel jaw crusher

and pulverized for 2.5 to 3 minutes using an agate mill at Carleton University, Ottawa. The pulverized samples were sent to ALS Laboratories for analysis of 13 major elements, and 42 minor and trace elements. A lithium borate fusion method was used to prepare the samples before being dissolved in acid and analyzed by inductively coupled plasma-atomic emission spectrometry (ICP-AES) and inductively coupled plasma-mass spectrometry (ICP-MS).

B.3 Isotope geochemistry

Powdered samples weighing between 40 and 105 mg were spiked with a ^{148}Nd - ^{149}Sm mixture and then dissolved in $\sim 29\text{M}$ HF and 16M HNO_3 on a hot plate for 5 days. Following complete dissolution, the solution was dried down and digested with 8M HNO_3 and 6M HCl . The sample was then dried down again and dissolved in 2.5M HCl and then loaded into 14 mL Bio-Rad borosilicate glass chromatography columns that contain 3 mL of Dowex AG50W-X8 cation resin. A total of 23 mL of 2.5M HCl was passed through the column before rare earth elements (REEs) were collected using 9 mL of 6M HCl . The resulting REE elution was then dried down, dissolved in 0.26 M HCl and loaded into Eichrom Ln Resin chromatographic columns holding Teflon powder coated with HDEHP [di(2-ethylhexyl)] orthophosphoric acid (Richard et al., 1976). Separate elutions of Nd and Sm were collected with 0.26 M HCl and 0.5 M HCl , respectively, and $5\text{ }\mu\text{l}$ of 0.3 M H_3PO_4 was added to each Teflon vial prior to being dried down. Samples were loaded onto Re filaments onto a sample wheel with a double filament assembly. Ratios were measured between 1500 and 1800°C using a Thermo-Finnigan Triton TI thermal ionization mass spectrometer. Isotope ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$.

Over a three year period an average values for the BCR-2 standard include Nd = 28.53 ppm, Sm = 6.618 ppm, and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512643 \pm 0.000011$ (n=13, 2σ). An IGGRC in-house Nd standard measured routinely over a period of three years yield $^{143}\text{Nd}/^{144}\text{Nd} = 0.511831 \pm 0.000011$ (2σ), an equivalent value of $^{143}\text{Nd}/^{144}\text{Nd} = 0.511861 \pm 0.000009$ (2σ) for the La Jolla Nd standard. Uncertainties on Sm and Nd concentrations are estimated less than 0.2%.

Appendix C : Standard analyses used in geochronology

From April 2015 to August 2018 seven standards of natural zircon (R33 and OG1) and synthetic solution (EARTHTIME 2.0 Ga solution) were analyzed. These standards are used in the geochronology community to confirm consistency of results across laboratories and to verify that systems and materials are performing optimally.

R33 – Natural Zircon

As characterized by Black et al. (2004) by the analysis of 50 single zircon crystals, R33 has a concordant weighted mean of 419.26 ± 0.39 Ma. Three analyses of individual R33 zircon crystals analyzed at the Jack Satterly Geochronology Laboratory by ID-TIMS yield concordant $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 422.07 ± 3.78 Ma, 419.76 ± 1.22 Ma and 421.16 ± 4.57 Ma (Figure C-1a) and overlap in error with the accepted R33 ID-TIMS values published by Black et al. (2004).

OG1 – Natural Zircon

The U-Pb isotope characteristics of natural zircon OG1 have been reported by Stern et al. (2009). The OG1 zircon standard has an average weight mean age of 3465.4 ± 0.6 Ma, representing the isotopic age of the natural zircon and an age of 3467.05 ± 0.63 Ma following chemical abrasion of zircons. The results obtained after chemical abrasion are considered to better represent the igneous age of the zircon (Stern et al., 2009). One chemically abraded zircon crystal was analyzed at the Jack Satterly Geochronology Laboratory using ID-TIMS and yield a 5.5% discordant result with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3451.82 ± 1.30 Ma (Figure C-1). A reference line anchored at 0 Ma with an upper intercept

of 3467.05 Ma is shown for comparison to the accepted age of OG1 (Figure C-1b). Although this particular analysis does not confirm analytical repeatability, it does highlight that zircons may exhibit lead loss despite the aggressive application of chemical abrasion (Mattinson, 2005).

EARTHTIME 2.0 Ga – Standard solution

Three aliquots of the EARTHTIME U-Pb 2.0 Ga synthetic solution (Condon et al., 2015) were measured between April and August 2015. These three analyses are concordant and yield $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2000.88 ± 0.36 Ma, 2000.93 ± 0.38 Ma and 2000.75 ± 0.73 Ma (Figure C-1c), confirming that the Jack Satterly Geochronology Laboratory at the University of Toronto was producing reliable and consistent results compared to standards published by other laboratories (Condon et al., 2015).

Figure C-1: U-Pb concordia diagrams for natural zircon and synthetic solution standards measured from April 2015 to August 2018. A. Concordant results for the natural zircon standard R33 (Black et al., 2004). B. Discordant result for a single analysis of the chemically abraded natural zircon standard OG1 (Stern et al., 2009). C. Three concordant results of the EARTHTIME 2.0 Ga synthetic standard (Condon et al., 2015).

Table C-1. U-Pb isotope results for standards analyzed between April 2015 and August 2018.

Fraction	Weight (mg)	U (ppm)	$\frac{\text{Th}}{\text{U}}$	Pbtot (pg)	PbC (pg)	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$ Meas.	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$ Meas.	2 σ	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	2 σ	Error Corr.
<i>R33</i>											
Zrc 1 (sk32p5)	0.011	95	0.63	75	0.67	6741	0.5119	0.0020	0.06723	0.00021	0.858
Zrc 2 (sk38p166)	0.012	53	0.69	166	0.39	24745	0.5115	0.0018	0.06715	0.00020	0.885
Zrc 3 (sk38p192)	0.020	68	0.64	98	0.29	19811	0.5080	0.0009	0.06676	0.00011	0.952
<i>OG1</i>											
Zrc 1 (sk38p169)	0.0049	199	2.1	1041	2.31	18081	27.37	0.11	0.6695	0.0026	0.979
<i>EARTHTIME 2.0 Ga synthetic solution</i>											
Sol 1 (sk36p131)	0.001	1238	0.09	445	3.1	9361	6.1755	0.0063	0.36399	0.00038	0.981
Sol 2 (sk36p132)	0.001	885	3.49	590	2.2	9517	6.1736	0.0035	0.36388	0.00022	0.931
Sol 3 (sk36p170)	0.001	696	0.09	251	2.6	6142	6.1665	0.0043	0.36349	0.00027	0.838
<i>Table C-1 continued on the next page</i>											

Notes: Zrc: Zircon; Meas. – Measured

R33 standard from Black et al. (2004). OG1 standard from Stern et al (2009) and EARTHTIME 2.0 Ga solution from Condon et al. (2015).

Th/U calculated from radiogenic $^{208}\text{Pb}/^{206}\text{Pb}$ ratio and $^{207}\text{Pb}/^{206}\text{Pb}$ age assuming concordance.

Pbtot is total amount of Pb excluding blank.

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

$^{206}\text{Pb}/^{204}\text{Pb}$ corrected for fractionation and common Pb in the spike.

Correction for ^{230}Th disequilibrium in $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ 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 $^{207}\text{Pb}/^{206}\text{Pb}$ age.

Uranium decay constants are from Jaffey *et al.* (1971).

Table C-1. Continued

Fraction	²⁰⁷ Pb	2 σ	²⁰⁶ Pb	2 σ	²⁰⁷ Pb	2 σ	²⁰⁷ Pb	2 σ	Disc. (%)
	²⁰⁶ Pb		²³⁸ U		²³⁵ U		²⁰⁶ Pb		
			Age (Ma)		Age (Ma)		Age (Ma)		
<i>R33</i>									
Zrc 1 (<i>sk32p5</i>)	0.055232	0.000112	419.47	1.27	419.73	1.36	421.16	4.57	0.4
Zrc 2 (<i>sk38p166</i>)	0.055243	0.000094	418.96	1.19	419.43	1.24	422.07	3.78	0.8
Zrc 3 (<i>sk38p192</i>)	0.055186	0.000030	416.63	0.69	417.11	0.61	419.76	1.22	0.8
<i>OG1</i>									
Zrc 1 (<i>sk38p169</i>)	0.296458	0.000248	3304.00	9.89	3396.67	3.96	3451.82	1.30	5.5
<i>EARTHTIME 2.0 Ga synthetic solution</i>									
Sol 1 (<i>sk36p131</i>)	0.123048	0.000025	2001.08	1.78	2000.98	0.89	2000.88	0.36	-0.01
Sol 2 (<i>sk36p132</i>)	0.123051	0.000027	2000.52	1.02	2000.72	0.49	2000.93	0.38	0.02
Sol 3 (<i>sk36p170</i>)	0.123039	0.000051	1998.71	1.28	1999.71	0.61	2000.75	0.73	0.12

Notes: Zrc: Zircon; Meas. – Measured

R33 standard from Black et al. (2004). OG1 standard from Stern et al (2009) and EARTHTIME 2.0 Ga solution from Condon et al. (2015).

Th/U calculated from radiogenic $^{208}\text{Pb}/^{206}\text{Pb}$ ratio and $^{207}\text{Pb}/^{206}\text{Pb}$ age assuming concordance.

Pbtot is total amount of Pb excluding blank.

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

$^{206}\text{Pb}/^{204}\text{Pb}$ corrected for fractionation and common Pb in the spike.

Correction for ^{230}Th disequilibrium in $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ 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 $^{207}\text{Pb}/^{206}\text{Pb}$ age.

Uranium decay constants are from Jaffey *et al.* (1971).

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