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of a terrane translation model**

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Chronostratigraphy of the Hottah terrane and Great Bear magmatic zone of Wopmay orogen, Canada, and exploration of a terrane translation model

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Abstract

The Paleoproterozoic Hottah terrane is the western-most exposed bedrock of the Canadian Shield, and a critical component for understanding the evolution of the Wopmay orogen. Thirteen new high-precision U-Pb zircon crystallization ages are presented and support field observations of a volcano-plutonic continuum from Hottah terrane through to the end of the Great Bear magmatism; from >1950 Ma to 1850 Ma. The new crystallization ages, new geochemical data, and newly published detrital zircon U-Pb data, are used to challenge hitherto accepted models for the evolution of the Hottah terrane as an exotic arc and micro-continent that arrived over a west-dipping subduction zone and collided with the Slave craton at ca. 1.88 Ga. While the Hottah terrane does have a tectonic history that is distinct from that of the neighbouring Slave craton, it shares a temporal history with a number of domains to the south and east, domains that were tied to the Slave craton by ca. 1.97 Ga. It is interpreted herein that Hottah terrane began to the south of its current position and evolved in an active margin over an always east-dipping subduction system that began prior to ca. 2.0 Ga and continued to ca. 1.85 Ga, and underwent tectonic switching and migration.

The stratigraphy of the ca. 1913 to 1900 Ma Hottah plutonic complex and Bell Island Bay Group comprises a subaerial rifting arc sequence, followed by basinal opening that is represented by marginal marine quartz arenite and overlying ca. 1893 Ma pillowed basalt flows and lesser rhyodacites. We interpret this stratigraphy to record Hottah terrane rifting off its parental arc crust, in essence the birth to the new Hottah terrane. This model is similar to rapidly rifting arcs in active margins, for example modern Baja California. These rifts generally occur at the transition between subduction zones (e.g., Cocos-Rivera plates) and transtensional shear zones (e.g., San Andreas fault) and we suggest that extension driven transtensional shearing, or more simply terrane translation, was responsible for the evolution of Bell Island Bay Group stratigraphy and that it transported this newly born Hottah terrane laterally (northward in modern coordinates), arriving adjacent to the Slave craton at ca. 1.88 Ga. Renewed east-dipping subduction led to the Great Bear arc flare-up at ca. 1876 Ma, continuing to ca. 1869 Ma. This was followed by voluminous Great Bear plutonism until ca. 1855 Ma. The model implicates that it was the westerly Nahanni terrane and its subducting oceanic crust that collided with this active margin, shutting down the > 120 million year old, east-dipping subduction system.

1 **Keywords**

- 2 Hottah terrane , Great Bear magmatic zone , CA-ID-TIMS U-Pb zircon , terrane translation ,
3 rifting arc

Draft

1 Introduction

2 In September 2012, NASA's Curiosity rover discovered an outcrop of alluvial derived
3 conglomerate in Gale Crater, Mars and was named Hottah, after Hottah Lake, Northwest
4 Territories, Canada (Williams et al. 2013; <http://mars.jpl.nasa.gov/msl/>). Ironically, the details of
5 some of the remote bedrock exposures around Hottah Lake, here on Earth, remain enigmatic.
6 The Wopmay orogen, the westernmost exposed bedrock of the Canadian Shield, is a middle
7 Paleoproterozoic orogenic belt that was one of the first Precambrian domains to have been
8 attributed to plate tectonic processes (Fraser et al. 1972; Hoffman 1973, 1980, 1988, 2012). The
9 Hottah terrane figures prominently in proposed evolutionary models for the Wopmay orogen and
10 it was recognized early on that understanding the Hottah terrane was key to deciphering the
11 Wopmay orogen (Hoffman 1973; McGlynn 1976, 1979; Hoffman and McGlynn 1977;
12 Hildebrand 1981). Forty years later, while new evidence continues to accumulate regarding the
13 development of Wopmay orogen (e.g., Hoffman et al. 2011; Jackson et al. 2013), details
14 regarding the evolution of Hottah terrane remain sparse (Bowring 1984). Presented herein are
15 new high-precision U-Pb zircon ages from the Hottah terrane and select volcanic horizons in the
16 central Great Bear magmatic zone, complemented with trace-element and Nd-isotope data. The
17 new and previously available data allow the construction of a stratigraphic framework that can be
18 used to evaluate previous tectonic models. We also use these data to propose a transtensional
19 shear model (terrane translation; Umhoefer 2011).

21 Geological Setting

22 Framework elements of the Wopmay orogen include the Coronation margin and its
23 cratonic and reworked Archean Slave craton basement in the east, and the Great Bear magmatic
24 zone and Hottah terrane in the west (Figs. 1-2; Fraser et al. 1972; Badham 1973; Hoffman 1973,
25 1980, 1984, 1988; Hoffman and McGlynn 1977; Hoffman et al. 1976; St-Onge and King 1986,
26 1987; Villeneuve et al. 1991, 1993; Aspler et al. 2003; Cook and Erdmer 2005; Jackson 2008;
27 Hildebrand et al. 1987a, b; 2010a, b; Jackson and Ootes 2012; Jackson et al. 2013; Ootes et al.
28 2013). The Coronation margin includes Archean crust, the ca. 2.014 Ga Melville Group rift to
29 drift sequence, 1.97 Ga Epworth Group passive margin sedimentary rocks, and ca. 1.88 Ga
30 Recluse Group foredeep sedimentary rocks preserved in the east-verging Asiatic thrust-fold belt,
31 as well as correlatives preserved as supracrustal, gneissic and intrusive rocks in the metamorphic

1 internal zone (Hoffman 1973, 1980, 1984; St-Onge and King 1986, 1987; Bowring and
2 Grotzinger 1992; Hoffman et al. 2011; Jackson et al. 2013). The Wopmay fault zone is a north-
3 south striking crustal-scale fault that delimits the western extent of exposed Archean rocks in the
4 Wopmay orogen (Figs. 1-2; St-Onge et al. 1983; King 1987; Bowring and Podosek 1989; Housh
5 et al. 1989; Jackson 2008; Jackson et al. 2006, 2013; Spratt et al. 2009; Jackson and Ootes 2012;
6 also called medial zone by Hildebrand et al. 1991, 2010a). West of Wopmay fault zone are the
7 remnants of the >1.9 Ga Hottah terrane, which was extensively overprinted by ca. 1.875 to 1.85
8 Ga Great Bear magmatism (e.g., Fraser et al. 1972; Badham 1973; Bowring 1984; Gandhi et al.
9 2001; Hildebrand et al. 2010b). Northeast-striking brittle faults dextrally offset all units and are
10 transected by the ca. 1740 Ma Cleaver dykes (Irvine et al. 2004; Hayward and Corriveau 2014).

11 The northern Wopmay orogen is covered by the Paleo- to Mesoproterozoic Coppermine
12 Homocline (Fig. 2; Kerans et al. 1981). The Hottah terrane and Great Bear magmatic zone are
13 covered in the west by Cambro-Ordovician and Cretaceous-Tertiary rocks of the Phanerozoic
14 platform (Figs. 1-2). Geophysical studies have shown that the Hottah terrane and Great Bear
15 magmatic zone have properties distinct from those of the Slave craton and overlying Coronation
16 margin to the east, and the Fort Simpson magnetic high and enigmatic Nahanni terrane to the
17 west (Fig. 1; Fraser et al. 1972; Hoffman 1988; Cook et al. 1999; Aspler et al. 2003; Cook and
18 Erdmer 2005; Wu et al. 2005; Spratt et al. 2009; Oueity and Clowes 2010; Ootes et al. 2013;
19 Hayward and Corriveau 2014). Much of what has been ‘imaged’ in these experiments is covered
20 by the Phanerozoic platform and directly assessing the underlying bedrock components,
21 especially of the Hottah terrane, is not currently possible. This lack of exposure introduces
22 uncertainty in the regional interpretation, for example a 100 km wide magnetic low under the
23 Phanerozoic platform, between the Great Bear magmatic zone and Fort Simpson terrane, is
24 interpreted to be Hottah terrane (Fig. 1; Aspler et al. 2003). As there are no sub-Phanerozoic
25 samples from this area, it is difficult to confirm what rocks are causing this uniformly low
26 magnetic signature. In fact, the actual Hottah bedrock exposures that are examined in this study
27 (Figs. 1 and 2) are commonly weakly to strongly magnetic and generally have mid- to high-
28 magnetic signatures on regional surveys.

29
30 ***Hottah terrane and Great Bear magmatic zone***

On the basis of its bedrock and geochemical characteristics, the Hottah terrane can be compartmentalized into at least five components. 1) A cryptic ca. 2.6 to 1.97 Ga source domain, indicated from U-Pb zircon ages in sedimentary rocks (Bowring 1984; Davis et al. submitted) and radiogenic tracer isotope studies on younger magmatic rocks (Nd, Pb; Bowring and Podosek 1989; Housh et al. 1989). 2) The ca. 1950 Ma Holly Lake metamorphic complex comprises the oldest recognized bedrock exposures west of the Wopmay fault zone. It consists of pelites and metabasalt, and is preserved at Great Bear Lake, Leith ridge, and along the Zebulon River on the east side of Hottah Lake (Fig. 2 and 4A; Hildebrand 1984; Hildebrand et al. 1983, 1984; Bowring 1984; Jackson and Ootes 2012; Davis et al. submitted). 3) Circa 1930 to 1910 Ma plutonic rocks that generally have penetrative deformation fabrics (Fig. 4B; McGlynn 1979; Hildebrand 1984, 1985; Hildebrand et al. 1983, 1984; Bowring 1984), referred to as Hottah plutonic complex. 4) The Bell Island Bay Group volcanic and sedimentary rocks that were erupted and deposited on top of the older and deformed Holly Lake metamorphic complex and Hottah plutonic complex rocks (Fig. 4C-E; Hildebrand et al. 1983, 1984; Reichenbach 1991). These are best preserved at Bell Island, the east shore of Hottah Lake, and at Beaverlodge Lake (Fig. 2). The Bell Island Bay Group consists of the basal Beaverlodge Lake sandstone (Fig. 4C), which is overlain by the $1898.4 \pm 7.2/-6.3$ Ma Zebulon Formation subaerial volcanic rocks (Fig. 4D; Reichenbach 1991). Hildebrand (1984) assigned the marginal marine quartz arenite and siltstone of the Conjuror Bay Formation, and overlying Bloom Basalts, as the base of the Great Bear magmatic zone. Reichenbach (1991), however correlated the Bloom Basalts with the ca. 1900 Ma Grant Subgroup and Akaitcho Group of Easton (1980, 1981a,b) and assigned the Conjuror Bay Formation and Bell Island Bay Group as the upper part of the Bell Island Bay Group. 5) The <1885 and >1875 Ma Treasure Lake Group metasedimentary rocks are considered the youngest component of the Hottah terrane (Gandhi and van Breemen 2005). The Treasure Lake Group type section is in the southern Great Bear magmatic zone with another exposure documented at DeVries Lake (Fig. 2A; Bennett and Rivers 2006b; Jackson 2008).

Two models have been proposed for the evolution of Hottah terrane (Fig. 3). The most common and widely used model advocates that the Hottah terrane was a ca. 2.5 to 2.0 Ga micro-continent with an overriding ca. 1.9 Ga volcanic arc. This arc developed over west-dipping subduction of the Coronation ocean crust that underwent a rift to drift transition at ca. 2.01 Ga (Hoffman et al. 2011). Coronation ocean subduction was terminated by the arrival and collision

of this exotic micro-continent and arc with the western Archean Slave craton and the overlying Coronation margin at ca. 1.88 Ga (Fig. 3A; Hildebrand 1981; Bowring and Podosek 1989; Bowring and Grotzinger 1992; Hildebrand et al. 2010a; Hoffman et al. 2011; Hoffman 2012). The time-frame of the transition from Coronation passive to active margin sedimentation (ca. 1.97 to 1.88 Ga) is consistent with modern ocean opening and closing, and micro-continent – continent collision is the attributed driver for this transition (Hoffman 2012; Hoffman et al. 2011). The second and less favourable model suggests that Hottah terrane evolved as an arc that developed on attenuated crust of the Slave craton and these were separated by a back-arc basin (Fig. 3B; Reichenbach, 1991; Bowring and Grotzinger 1992). This model was consistent with bedrock mapping and petrological studies within the metamorphic internal zone (east of the Wopmay fault zone) and the results were interpreted to record opening and closing of an ensialic back-arc basin between 1.89 and 1.88 Ga (Easton 1980, 1981a; St-Onge and King 1986, 1987; Lalonde 1989).

The ca. 1875 to 1850 Ma Great Bear magmatic zone igneous rocks (Fig. 4F-G) were erupted onto and intruded into the older Hottah terrane stratigraphy and dominate the bedrock exposures west of the Wopmay fault zone (Figs. 1 and 2; Bowring 1984; Gandhi et al. 2001; Hildebrand et al. 2010b). Screens of volcanic rocks that range from andesite to trachyandesite to rhyolite are preserved in a sea of younger plutons that only rarely contain penetrative deformation fabrics. These plutonic rocks range in composition from gabbro to syenogranite, although they are predominantly monzonitic to monzogranitic (e.g., Fraser et al. 1972; Badham 1973; Hildebrand 1984; Hildebrand et al. 1987, 2010b; Gandhi 1994; Gandhi et al. 2001; Ootes et al. 2013).

The Great Bear magmatic zone is considered to have been an Andean-like magmatic arc resultant from oblique east-dipping subduction under the Hottah terrane and Slave margin (Fig. 3A; Badham 1973; Hildebrand 1981; Hildebrand et al. 1987a, 2010a, b). Geophysical experiments support this and show that east-dipping, subducted Paleoproterozoic oceanic crust is now frozen beneath the Great Bear magmatic zone and the western Slave craton (Fig. 1B; Cook et al. 1999; Cook and Erdmer 2005; Wu et al. 2005; Oueity and Clowes 2010). The geophysical features intimate that this oceanic crust was rooted in the enigmatic Nahanni terrane to the west (Fig. 1).

Bedrock mapping and U-Pb zircon geochronology of Archean basement in the metamorphic internal zone of the Wopmay orogen and bedrock mapping in the vicinity of Hottah Lake has raised a number of questions about the previously ascribed relationship of the Slave craton and Hottah terrane (Jackson and Ootes 2012; Jackson et al. 2013). For this study we collected and dated key volcanic and plutonic units within the central area of the Wopmay orogen, mostly between 64°N and 65°N (Fig. 2), in order to help resolve stratigraphic challenges that arose during bedrock mapping, and to assess the temporal and petrological evolution of the Hottah terrane – Great Bear magmatic zone transition. The new and previously published data (Hildebrand et al. 1983; Bowring 1984; Reichenbach 1991; Gandhi et al. 2001; Bennett and Rivers 2006a) allow the construction of a time-constrained stratigraphic column that records the transition from Hottah terrane to Great Bear magmatic zone tectonics. Such constraints are needed to further evaluate the proposed models for the evolution of Wopmay orogen.

U-Pb Zircon Geochronology & Lithogeochemistry

U-Pb zircon geochronology was completed at the Geological Survey of Canada, Ottawa, Ontario, Canada. Zircon separates were analyzed by chemical abrasion isotope dilution thermal ionization mass spectrometry (CA-ID-TIMS; Mattinson 2005). The zircon separates from one sample (09vj171) were mechanically abraded prior to analysis (Krogh 1982). Those from sample 09vj148c were of poor quality and therefore zircons from this sample were analyzed using the sensitive high resolution ion microprobe (SHRIMP II; Stern 1997). Sample locations are identified on Figure 2, analytical techniques are in the Appendix, and results are tabulated in Tables 1 and 2 and presented in Figure 5.

To complement the dating and evaluate the petrologic evolution of the Hottah terrane and Great Bear magmatic zone rocks, a number of samples were selected for major and trace-element geochemistry and Sm-Nd isotope geochemistry (Tables 3, A1, and A2). Analytical techniques are in Appendix I. More in-depth bedrock descriptions are available in the online Supplementary data.

Hottah Plutonic Complex

Porphyritic Granite (Hg)

Coarse-grained K-feldspar porphyritic granite with partially chloritized biotite was collected from Beaverlodge Lake ridge ('Hg' on Fig. 2B). This granite is similar to other 'syenogranites' throughout this area and previous investigators assigned it to the Hottah terrane (Hildebrand and Roots 1985). Locally, the granite is unconformably overlain by coarse pebble conglomerate of the Conjuror Bay Formation, or rhyolite flows of the Zebulon Formation, both of the Bell Island Bay Group. Five single grains were analyzed by CA-ID-TIMS, yielding four near concordant and one 2.2% discordant analyses (Table 1). A $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean of the four most concordant grains provides a date of 1930 ± 1.0 Ma and a best-fit discordia line through the data provides an indistinguishable upper intercept age of 1930.6 ± 1.3 Ma, with a lower intercept of 276 ± 160 Ma (MSWD = 0.71; Fig. 5A; Table 1). The 1930.6 ± 1.3 Ma upper intercept age is interpreted as the timing of magma crystallization.

Hypabyssal Porphyry (Hp)

Feldspar and quartz phyric porphyry was sampled on Beaverlodge ridge ('Hp' on Figs. 2B and 4E). This porphyry has a fine-grained groundmass and is interpreted as hypabyssal in nature. The porphyry is spatially associated with the porphyritic granite discussed above (09lo9078a); however, no contacts are exposed. The porphyry is unconformably overlain by two formations of the Bell Island Bay Group. At the lower unconformity hematized, fragmental rhyolite flows and volcanic breccia of the Zebulon Formation overlie the porphyry. The Conjuror Bay Formation quartz arenite overlies and cuts into both the Zebulon Formation volcanic rocks and the older porphyry (Fig. 4E; Shakotko et al. 2012). The porphyry is moderately to extensively altered, with abundant hematite in the groundmass, minor magnetite replacement of some mafic crystals, and hematite-chlorite veining. Hildebrand and Roots (1985) assigned this porphyry to the Great Bear magmatic zone as the unconformable nature of the overlying units (Fig. 4E) was not yet recognized.

A least altered sample of the porphyry yielded pristine igneous zircon, of which five grains were analyzed by CA-ID-TIMS (Table 1). Three grains yield concordant dates at 1931 Ma whereas two others are very slightly discordant. A weighted mean of the five analyses indicates a $^{207}\text{Pb}/^{206}\text{Pb}$ crystallization age of 1930.5 ± 1.1 Ma (MSWD = 0.95; Fig. 5B; Table 1). This crystallization age is indistinguishable from that of the porphyritic granite and these are

collectively assigned as part of the Hottah plutonic complex. Samarium-Neodymium isotopic data indicate this quartz-plagioclase porphyry has an ϵNd^{1930} value of -1.5 (Fig. 6; Table 3)

Strongly deformed granodiorite (ZR)

A strongly deformed to mylonitic granodiorite with abundant biotite-schlieren represents the oldest plutonic phase recognized south of the Zebulon River ('ZR' on Fig. 2A). This is the most pervasively deformed rock we have observed in the study area west of Archean mylonite within the Wopmay fault zone. The deformed granodiorite occupies a 100 x 100 m area, is overlain by one small outcrop of quartzite, and is intruded by hornblende-biotite granodiorite, which is also deformed, albeit much more weakly (sample #11lo4155 below). Zircons from the granodiorite are clear and colourless prismatic grains. Five of six CA-ID-TIMS analyses yield a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1927.9 ± 0.9 Ma (MSWD = 0.68, Fig. 5C; Table 1), interpreted as the crystallization age of the rock. This plutonic rock is slightly younger, albeit much more strongly deformed, than the porphyritic granite and hypabyssal porphyry at Beaverlodge Lake. The granodiorite has an ϵNd^{1930} value of -1.6 (Fig. 6a; Table 3).

Foliated Tonalite (Ht)

On an island in the east-central Hottah Lake, a tonalite ('Ht' on Fig. 2B) is intruded by diorite, quartz diorite, and equigranular granite. The rocks contain an east-southeast striking, steeply south-dipping foliation and have undergone moderate secondary alteration, including epidote replacement of hornblende and localized magnetite-bearing veins. Mafic dykes, likely Fishtrap Lake gabbros (Reichenbach 1991), cross-cut the more leucocratic plutonic phases.

Zircons from an unaltered tonalite, representing the oldest phase on the island, consist of euhedral prismatic grains or grain fragments, many of which are heavily fractured. Six single grains were analyzed by CA-ID-TIMS, yielding a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1912.9 ± 0.7 Ma (MSWD = 0.35; Fig. 5D; Table 1), interpreted as the crystallization age of the rock. The tonalite has an ϵNd^{1913} value of +1.8 (Fig. 6; Table 3).

Bell Island Bay Group

Zebulon Formation Spherulitic Rhyolite (Zr)

On the southwest end of Beaverlodge ridge, the Zebulon and Conjuror Bay formations unconformably overlie the Hottah plutonic complex (Fig. 2B). The Zebulon Formation includes rhyolitic and basaltic volcanic rocks and a distinctive flow of variably hematized spherulitic rhyolite, with spherules up to 5 cm in diameter and quartz-filled lithophysae, was collected for U-Pb zircon dating ('Zr' on Fig. 2B; Fig. 4C). This flow appears identical to a similar flow on the east side of Hottah Lake, about 30 km to the north.

The sample yielded euhedral prismatic zircon. Four single zircon analyses by CA-ID-TIMS have overlapping $^{207}\text{Pb}/^{206}\text{Pb}$ ages with a weighted mean age of 1905.6 ± 1.4 Ma (MSWD = 0.88; Fig. 5E; Table 1). A sample of Zebulon basalt from the east side of Hottah Lake has $\epsilon\text{Nd}^{1906} = +2.1$, and the Zebulon Formation rhyolite dated above has an ϵNd^{1906} value of +0.8 (Fig. 6; Table 3).

Bloom Rhyodacite (Br)

One of the islands in central Hottah Lake consists entirely of rhyodacite flows, interpreted to overlie the mafic volcanic dominated Bloom Basalt pile of the Bell Island Bay Group (Fig. 2B; Reichenbach, 1991). A sample of a fragmental rhyodacite flow ('Br' on Fig. 2) was carefully selected in order to test the timing of this mafic volcanism. The rhyodacite sample yielded clear, colourless euhedral prismatic zircon. Four single grain CA-ID-TIMS analyses yielded concordant to reversely discordant data with overlapping $^{207}\text{Pb}/^{206}\text{Pb}$ ages giving a weighted mean of 1895 ± 2.3 Ma (MSWD = 1.3; Fig. 5F; Table 1). This is interpreted as the age of the rhyodacite and is a minimum age for the Bloom Basalts. The Bloom rhyodacite sample has an ϵNd^{1895} value of -1.8 and a Bloom Basalt sample has an ϵNd^{1895} value of +2.3 (Fig. 6; Table 3).

Grant Subgroup

Grant Rhyodacite (GL)

Reichenbach (1991) correlated the Bloom Basalts to the Grant Subgroup (Easton, 1981b) that occur east of Wopmay fault zone. A sample of rhyodacite from within the Grant pillowed basalts, north of Grant Lake ('GL' on Fig. 2) was collected for U-Pb zircon dating. The rhyodacite yielded clear, colourless equidimensional to stubby zircons, devoid of fractures or mineral inclusions. Five mechanically-abraded single zircon grains were analyzed by ID-TIMS

and yielded concordant to 0.7% discordant data points that are aligned with upper and lower intercept ages of 1895 Ma and 549 Ma, respectively. A weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1892.9 ± 1.5 Ma (MSWD = 0.29; Fig. 5G; Table 1) is interpreted as the age of crystallization of the rhyodacite and the interbedded pillowed basalts. Two samples from the Grant Subgroup basalts, one from the rhyolite sample location north of Grant Lake and the other from the north end of Rebesca Lake ~40 km south to the south (Fig. 2), have ϵNd^{1893} values of +1.7 and +1.7 (Fig. 6; Table 3).

LaBine Group

Rhyolitic Ash tuff (DL)

At DeVries Lake ('DL' on Fig. 2A) strongly deformed and altered Treasure Lake Group metasedimentary rocks are overlain by volcanic rocks of the Great Bear magmatic zone (Gandhi 1994; Jackson 2008; Ootes et al. 2010). The volcanic rocks, which are preserved in a few small islands in the southern part of the lake, are variably altered; lithic fragments are altered to either magnetite, actinolite, or biotite. These volcanic rocks contain a weak foliation, defined by the flattened fragments. Rare, fine-grained felsic ash-tuff beds were observed on the east side of one of the islands and sampled to determine the timing of volcanism. Zircon separates form clear, colourless, well-terminated euhedral prisms 150 to 200 μm in length. Three single-grain analyses were undertaken by CA-ID-TIMS and yield concordant results with a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1876.4 ± 2.4 Ma (MSWD = 1.5; Fig. 5H; Table 1), interpreted as the time of crystallization.

Fragmental Andesite (BLa)

On the east side of Beaverlodge Lake ('BLa' on Fig. 2B), stratigraphically on top of quartz arenite is a small isolated outcrop of altered vesicular basalt (Bloom Basalt) and more extensive outcrops of unaltered fragmental rhyodacite and andesite; however, the contacts between these various units are obscured by thick flora. A sample of grey-white feldspar and quartz phyric volcanoclastic andesite with resorbed quartz grains and sparse angular lithic clasts was collected. The sample yielded colourless, euhedral prismatic zircon. Four single-grain analyses by CA-ID-TIMS yield <1% discordant data with a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of

1872.6 ± 0.9 Ma (MSWD = 0.95; Fig. 5I; Table 1) interpreted as the crystallization age of the andesite. The andesite has an εNd¹⁸⁷³ value of -0.6 (Fig. 6; Table 3).

Faber Group

Fiamme-bearing Rhyolite Ignimbrite (SL)

In the vicinity of Swan Lake, a sample of rhyolitic to rhyodacitic, fiamme-bearing ignimbrite was collected for U-Pb zircon dating (‘SL’ on Fig. 2A; Fig. 4F). The sample yielded clear, colourless, euhedral prismatic zircon, some with brownish glass inclusions typical of volcanic zircon. A sub-population, which contained possible cores, was excluded from analysis. Five of six single-grain CA-ID-TIMS analyses are concordant and yield a weighted mean ²⁰⁷Pb/²⁰⁶Pb crystallization age of 1868.8 ± 1.0 Ma (MSWD = 0.97; Fig. 5J; Table 1). A single analysis yields a slightly older age of 1876 ± 1.7 Ma (Fig. 5J; Table 1), and is interpreted to be a xenocryst, or includes an inherited component from a slightly older magmatic event. The rhyolite has an εNd¹⁸⁶⁹ value of -1.0 (Fig. 6; Table 3).

Deformed Pluton

Weakly Deformed Hornblende-Biotite Granodiorite (ZR)

South of Zebulon River (‘ZR’ on Fig. 2B), hornblende-biotite granodiorite intruded strongly deformed to mylonitic granodiorite (sample 11lo4156 above). The younger granodiorite contains a moderately developed hornblende- and biotite-defined foliation and, based on this, it was hypothesized to be a component of the Hottah plutonic complex. A sample of this granodiorite yielded zircons that were clear and colourless, with euhedral prismatic habit, although most were fractured multi-faceted prisms. Four single zircon grains were analyzed by CA-ID-TIMS and yielded <1% discordant ages with a weighted mean ²⁰⁷Pb/²⁰⁶Pb age of 1869.3 ± 1.1 Ma (MSWD = 0.68; Fig. 5K; Table 1), interpreted as the crystallization age of the sample. The granodiorite has an εNd¹⁸⁶⁹ value of -2.8 (Fig. 6; Table 3).

Sloan Group

Rhyolitic Crystal Tuff (SBr)

North of Stairs Bay of Hottah Lake, there are isolated outcrops of rhyolite that are cut by mafic dykes feeding overlying vesicular basalts. The rhyolites are unaltered and well-preserved,

they do not contain the hematite-alteration prevalent in the older Zebulon Formation volcanic rocks. A crystal-rich tuff, with phenocrysts of partially broken alkali feldspar up to 1 cm across, along with slightly smaller partially resorbed quartz in a groundmass of recrystallized spherules, was collected for U-Pb zircon analysis ('SBr' on Figs. 2; Fig. 4G). The sample yielded colourless, euhedral prismatic zircon. Ten single grain analyses by CA-ID-TIMS yield variably discordant ages between 1858 and 1865 Ma (Table 1). Nine of the ten analyses yield a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1861.7 ± 1.1 Ma (MSWD = 1.5; Fig. 5L; Table 1). With one younger analyses at 1858 Ma rejected, a regression of the data yields an age of 1862.8 ± 1.8 Ma and a lower intercept within error of zero at 257 ± 330 Ma (MSWD = 1.4; Fig. 5L), interpreted as the crystallization age of the rhyolite. The vesicular basalt has $\epsilon\text{Nd}^{1862} = 0.4$, whereas the dated rhyolite has an ϵNd^{1862} value of -0.5 (Fig. 6; Table 3).

Dumas Group

Rhyodacite Flow (HL)

Along the north-striking Wopmay fault zone, mixed basalt, rhyolite, and red arkosic rocks have been assigned to the Dumas Group (Hoffman 1978; Easton 1981; Hildebrand et al. 1987b; Jackson and Ootes 2012), and are considered as the eastern equivalent of the LaBine Group (Hildebrand et al. 1987b). Near Ham Lake ('HL' on Fig. 2) a sample of siliceous, purple-weathered plagioclase-phyric rhyodacite with local amygdules, lithic fragments, and preserved flow banding was selected for U-Pb zircon dating. Zircons from the rhyodacite sample were stubby to equidimensional prisms. The grains were of various morphology and poor quality, containing numerous cracks and evidence of alteration. These zircons were analyzed using the sensitive high-resolution ion microprobe (SHRIMP II; analytical techniques in Appendix I). Nineteen of twenty analyses yielded <5% discordant results with a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean age of 1869.5 ± 2.3 Ma (MSWD = 1.3; Fig. 5M; Table 2), the interpreted crystallization age of the unit.

Discussion

Stratigraphy and Petrological Evolution

The stratigraphic sections in Figure 7 are based on new data and mapping for this study and previously published mapping, nomenclature, and data (Fraser et al. 1972; Hoffman and

McGlynn 1977; Hoffman 1978, 1984; Hoffman et al. 1976; McGlynn 1976, 1979; Wilson 1979; Easton 1980, 1981a,b; Hildebrand 1981, 1984, 1985; Hildebrand et al. 1983, 1984, 1987a, b, 2010a, b; St-Onge et al. 1983, 1991; Bowring 1984; Reichenbach 1991; Gandhi 1994; Goad et al. 2000; Gandhi et al. 2001; Gandhi and van Breemen 2005; Bennett and Rivers 2006a,b; Jackson 2008; Shakotko et al. 2012; Jackson and Ootes 2012; Jackson et al. 2013). Figure 7 illustrates a minimum 80 million year magmatic evolution from ~1930 Ma, the time of intrusion of the Hottah plutonic complex, to ca. 1850 Ma, the youngest Great Bear plutonic phases in the Great Bear magmatic zone (Bowring 1984; Gandhi et al. 2001). The detrital zircon record of the Holly Lake metamorphic complex indicates the presence of a >2.0 to 1.95 Ga magmatic arc in its source environment, potentially suggesting a >150 million year magmatic evolution of the Hottah terrane through Great Bear magmatism. The U-Pb crystallization ages of the younger ca. 1913 Ma plutons, e.g., the dated tonalite and associated gabbros, are in general complexly intermingled with a full-range of intrusive phases and have trace-element attributes similar to arc-derived magmas (Figs. 4B, 8, 9A, 10A-B; Table A1). Some of the major elements and mobile trace elements show various signs of secondary alteration (Fig. 8A-B, 9A); however, the relatively immobile trace elements and rare earth elements (REE) appear unaffected (Fig. 8C, 9A, 10A-B; Pearce 1996).

The Bell Island Bay Group erupted in a subaerial environment between ca. 1906 and 1895 Ma (Figs. 5 and 7) and this is a key time-frame for deciphering Hottah terrane evolution. Our field mapping and geochronological results, and those in Reichenbach (1991), indicate the 1905.6 ± 1.4 Ma Zebulon Formation (Fig. 5E) is dominated by subaerial rhyolite, with lesser vesicular basalt. Using SiO₂ contents, Reichenbach (1991) argued that the Zebulon Formation plot as a continuum from basalt to rhyolite, but is dominantly andesitic in composition (Reichenbach 1991). However, these rocks show visible evidence of alteration, such as bleaching, reddening, or brecciation, cemented by hematite, also discussed by Reichenbach (1991). The rhyolites plot on a line between Al₂O₃ and K₂O, on the chemical index of alteration plot (K:Ca+Na = close to 100:0) and plot on a tie line on Shand's Index diagram (Fig. 8A-B; Table A1; Reichenbach 1991). These major elements are not primary rhyolite compositions and are best interpreted as being influenced by post-volcanic alteration, suggesting that the major elements (e.g., SiO₂) are not rigorous discriminators of rock type (Fig. 8A-B). The more robust HFSE ratios (Pearce 1996) of the Zebulon Formation volcanic rocks demonstrate a bimodal

distribution of basalt and dacite/rhyolite with few rocks of intermediate composition (Fig. 8C; Table A1), consistent with our field observations and those documented on the map and in the text of Reichenbach (1991). A few samples from Reichenbach (1991) contain anomalously high zirconium (Fig. 8D) and plot in the alkaline/sub-alkaline trachyte field (Fig. 8E). Cross-checking the sample locations on the map in Reichenbach (1991) indicate that most of these high zirconium samples were collected along strike in a single mapped flow (unit PZ-Ld). Regardless, these are not typical arc andesites (Kelemen et al. 2004). Like some of the major elements, some of the trace-elements may be biased by alteration, particularly the mobile trace elements like Sr and possibly Eu (due to their relationship with plagioclase and the obvious loss of Ca and Na in these rocks; Fig. 8A-B). The remaining trace-elements and REE appear to have escaped extensive alteration and normalized diagrams indicate the Zebulon rhyolites are comparable to high-temperature rhyolites derived from partial melts from a basaltic source (Fig. 9; e.g., Nicholson et al. 1991; Cousens 1996). These Zebulon Formation felsic rocks are similar to high-silica rhyolites from suprasubduction zone environments, such as the Taupo volcanic zone in New Zealand, the Long Valley caldera of the western USA (Fig. 9B), as well as non-arc settings such as Krafla volcano in Iceland (Cole 1990; Jónasson 1994; Cousens 1996; Moulton et al. 2011). The Zebulon Formation basalts are geochemically comparable to low-Ti basalts derived from a modified mantle wedge in a suprasubduction zone setting (Figs. 9C and 10D-E; Pearce and Norry 1979; Shervais 2001; Pilet et al. 2004). This suprasubduction zone signature in the Zebulon Formation volcanic rocks (Figs. 8-10), and the fact they erupted subaerially on a plutonic basement that was unroofed within 7 Ma of crystallization, is most compatible with a suprasubduction zone environment undergoing rapid extension (Umhoefer 2011). Whether or not these plutonic and volcanic rocks were derived from an active subduction zone, or from a previously subduction-modified mantle, is not resolvable based on geochemistry (e.g., Bryan et al. 2015).

The ca. 1900 Ma Conjuror Bay Formation marginal marine quartz arenite sits unconformably on the older Hottah stratigraphy, including the Zebulon Formation (Fig. 4E; (Hildebrand 1984, 1985; Reichenbach 1991; Jackson and Ootes 2012), and marks drowning of the rifting arc (Fig. 7). A well-developed regolith occurs at this unconformity within the underlying Hottah porphyry (Fig. 4E) and it is quite possible that the processes altering this horizon were also responsible for the significant alteration in the underlying units of the Zebulon

Formation (Shakotko et al. 2012). Hildebrand (1984, 1985) assigned the Conjuror Bay Formation as the base of the ca. 1875 Ma LaBine Group in the Camsell River district, whereas at Hottah Lake the Conjuror Bay Formation is overlain by the Bloom sequence, now dated at 1895 ± 2.3 Ma (Fig. 5F and 7). Consequently, we argue that the Conjuror Bay Formation and Bloom Basalts are part of the Bell Island Bay Group rather than the LaBine Group (as in Table 1 in Reichenbach 1991; Fig. 7). It is likely that the Moose Bay tuff (Hildebrand 1984) or the 1876.4 ± 2.4 Ma volcanic rocks at DeVries Lake (Fig. 5H), represent the lowest stratigraphy in the Great Bear volcanic arc.

The Bloom Basalts and lesser rhyodacites are temporally and geochemically correlative with the 1892.9 ± 1.5 Ma Grant Subgroup east of the Wopmay fault zone (Fig. 6; Easton 1981; Reichenbach 1991). The basalts have geochemical attributes, such as HFSE concentrations (e.g., Th, Nb, Ta concentrations) that are best explained as a result of melting from a suprasubduction zone, or subduction-modified mantle (Figs. 9D and 10B-E; Table A1; Pearce and Norry 1979; Pilet et al. 2004) and the stratigraphic position of these basalts is consistent with their eruption in a highly attenuated arc, or marginal basin setting (Figs. 7 and 10; Wilson 1979; Easton 1980; Reichenbach 1991), which was previously termed the Akaitcho seaway (Fig. 3B; Reichenbach 1991). In the west, the basalts were erupted on the older Hottah terrane stratigraphy, but to the east we have been unable to demonstrate conclusively if the Grant Subgroup basalts were erupted on Archean basement, or if they were thrust over this basement (Fig. 7; Jackson and Ootes 2012; also see Easton 1980, 1981b; St-Onge et al. 1983; Hildebrand et al. 2010a). Both the Bloom Basalt and Grant Subgroup basalt samples have modestly positive $\epsilon_{\text{Nd}}^{1895}$. This indicates their parental melts did not extensively interact with earliest Proterozoic or Archean basement (Fig. 6; Table 2) and further supports their eruption in a highly attenuated suprasubduction zone environment, be it rifted arc or marginal basin.

Due to erosion and overprinting by younger magmatism, the stratigraphy of the Hottah terrane (Fig. 7) is preserved in a disconnected fashion over about 1800 km² (Fig. 2). The limited extent of preservation means it is not possible to constrain with certainty the larger-scale processes driving the ca. 1930 to ca. 1895 Ma petrological evolution. However, the evolving magmatic system is consistent with a rapidly rifting arc setting. Such an environment may result from a variety of subduction zone adjustments including slab roll-back, delamination, or flattening, or possibly the development of a pseudo-slab window from outboard ridge subduction

(e.g., Dickinson 2002; Thorkelson et al. 2011; Giba et al. 2013). However, the suprasubduction zone geochemical signature of the rocks can also be attained from a prior, rather than active, subduction-modified mantle (e.g., Bryan et al. 2014). What we have not found is support of a lateral migration in magmatic activity from west to east across the Hottah terrane during this time-interval as proposed by Hildebrand et al. (2010a). As discussed below, this finding has significant implications on the tectonic interpretations for the Hottah terrane.

The Treasure Lake Group is preserved in the southern Great Bear magmatic zone (Fig. 2) and represents an enigmatic sedimentary unit. It has a maximum deposition age of ca. 1885 Ma, it was deformed and metamorphosed prior to or during the onset of Great Bear magmatism at ca. 1875 Ma (Fig. 7; Gandhi and van Breemen 2005; Bennett and Rivers 2006b), and may represent a syn-collisional basin similar to the ca. 1882 Ma Recluse Group in the Coronation margin (Hoffman et al. 2011). It is not known if there is any place where the Treasure Lake Group is in contact with the older Bell Island Bay Group, Hottah plutonic complex, or Holly Lake metamorphic complex. The Treasure Lake Group contains older Archean detrital zircons with ages typical of the Slave craton, in addition to 2.33 to 2.26 Ga detrital zircons that were attributed to a Hottah terrane source (Gandhi and van Breemen 2005); however, such ages are also characteristic of Paleoproterozoic tectonic events recorded in the western Rae craton (e.g., Davis et al. submitted).

The Great Bear magmatic zone volcanic rocks unconformably overly Hottah terrane rocks in the central and northern areas and the Treasure Lake Group in the south (Fig. 2; Hildebrand 1984, 1985; Hildebrand et al. 1983; Goad et al. 2000; Gandhi et al. 2001; Jackson 2008; Jackson and Ootes 2012). The newly obtained U-Pb crystallization ages for Great Bear volcanic rocks support previous data that indicate the LaBine Group volcanic arc erupted between ca. 1878 and 1870 Ma (Figs. 5 and 7; Table 3; Bowring 1984; Bennett and Rivers 2006a). The 1868.8 ± 1.0 Ma rhyolite in this study is similar in both age and geochemistry to the dominantly rhyolitic Faber Group in the southern Great Bear magmatic zone (Figs. 7-10; Gandhi et al. 2001; Azar 2007). In addition, the 1869.5 ± 2.3 Ma Dumas Group that spatially coincides with Wopmay fault zone (Figs. 2 and 7) is indistinguishable in age from the Faber Group and the deformed 1869.3 ± 1.1 Ma Great Bear hornblende-biotite granodiorite to the west (Fig. 5). Volumetrically dominant ca. 1865 Ma monzonitic to granitic plutonic phases (Ghandi et al. 2001; Bennett and Rivers 2006a) are not ductilely deformed. We, therefore, estimate the tectonic

event that led to the deformation of the volcanic rocks and hornblende-biotite granodiorite to have occurred at approximately 1867 Ma, coincident with a significant deformation event now recognized to the east of the Wopmay fault zone (Jackson et al. 2013).

Relatively pristine 1862.8 ± 1.8 Ma felsic tuff and associated amygdaloidal basalts that occur east of Hottah Lake are likely correlative with the more northerly Sloan Group volcanic rocks and intrusive porphyries (Hoffman et al. 1976; Hoffman 1978; Bowring 1984) and a rhyodacitic ignimbrite that post-dates mineralization at the Sue-Diane Cu-Au prospect to the south (Fig. 7; Gandhi et al. 2001). The bimodal Sloan Group volcanic package, recognized in this study (Figs. 7-10; Table A1), post-dates the ca. 1865 Ma plutonic events (e.g., Gandhi et al. 2001; Bennett and Rivers 2006a; Hildebrand et al. 2010b) and pre-dates the final ca. 1855 Ma rapakivi-textured plutonism (Gandhi et al. 2001), as well as the younger ca. 1843 Ma syenogranite magmatism (Bowring 1984).

Whence came Hottah terrane?

The Slave craton and overlying Coronation margin did not share a common tectonostratigraphic, or plutonic history with the Hottah terrane prior to ca. 1876 Ma (Bowring 1984; Hildebrand et al. 2010a; Jackson et al. 2013). In addition, detrital zircons and Nd and Pb isotope data do not support that the Hottah terrane was built on attenuated Slave craton crust (Bowring and Podosek 1989; Housh et al. 1989; Hildebrand et al. 2010a; Davis et al. submitted). The simple marginal basin model entirely within Slave crust, depicted in Figure 3B, is therefore ruled as an unfavourable solution for the origin and evolution of Hottah terrane (Bowring and Grotzinger 1992).

Samarium-Neodymium isotopic values from the Hottah terrane and Great Bear magmatic zone were used to indicate a juvenile ca. 2.0-2.4 Ga crustal basement (Bowring and Podosek 1989). Our new Nd isotopic results are similar (Fig. 6; Table 3) and provide further information on the nature and provenance of the yet undiscovered >1950 Ma basement. The T_{DM} ages range from 2.2 to 2.5 Ga (Table 3). One interpretation is that these could represent an accurate measure of a crustal protolith age (e.g., Bowring and Podosek 1989). Alternatively, these values record a mixture of juvenile mantle and a crustal end member, as is typical of igneous rocks in magmatic arc settings. This would suggest a minimum age of the crustal endmember at ca. 2.5 Ga, the maximum T_{DM} age (Table 3). Assuming a continental arc setting, which is consistent with the

1 detrital zircon ϵHf and $\delta^{18}\text{O}$ results in the Hottah terrane (Davis et al. submitted), this requires
2 the Hottah basement to be made up of either juvenile Neoarchean crust, or Neoarchean crust that
3 was recycled in the Proterozoic, not juvenile Paleoproterozoic crust.

4 The independent evolution and the transition from passive to active Coronation margin
5 sedimentation are cited to support the micro-continent – continent collision model (Fig. 3A), in
6 which the Hottah terrane evolved as an <1940 Ma exotic arc built on a ca. 2.0-2.4 Ga basement.
7 In this model, arc magmatism began after the opening of the Coronation ocean and the micro-
8 continent and overriding arc accreted to the Slave craton at ca. 1882 Ma, the onset of the
9 Calderian orogeny (Fig. 3A; Hildebrand 1981; Bowring and Grotzinger 1992; Cook and Erdmer
10 2005; Hildebrand et al. 2010a; Hoffman et al. 2011; Hoffman 2012). The rift to drift transition in
11 the Coronation margin is recorded at ca. 2014 Ma (Hoffman et al. 2011), followed by passive
12 margin sedimentation at ca. 1970 Ma (Bowring and Grotzinger 1992). The detrital zircon record
13 in the ca. 1950 Ma Holly Lake metamorphic complex metasedimentary rocks, the oldest exposed
14 bedrock of the Hottah terrane, indicates that arc magmatism in Hottah terrane began prior to ca.
15 2.0 Ga (Davis et al. submitted). Thus, it is difficult to reconcile active arc magmatism above a
16 long lived >2.0 to 1.89 Ga, west-dipping arc system (Fig. 3A), as arc magmatism initiated within
17 the Hottah terrane prior to, or simultaneously with the opening of the <2.01 Ga Coronation
18 oceanic basin (Hoffman et al. 2011). The model in Figure 3A is therefore ruled as an
19 unfavourable solution for the origin and evolution of Hottah terrane (Davis et al. submitted).

20 The Holly Lake metamorphic complex metasedimentary rocks have a detrital zircon
21 provenance that strikingly matches the igneous history of the western Rae craton (Davis et al.
22 submitted), including a record of the Queen Maud, Arrowsmith, Taltson-Thelon and Ksituan
23 magmatic events (Bostock and Loveridge 1988; van Breemen and Henderson 1988; Villeneuve
24 et al. 1993; Ross 2002; Berman et al. 2013). These domains were potentially tied to Slave craton
25 prior to the deposition of the Holly Lake metamorphic complex (Hoffman 1987, 1988a). If that
26 crust was the source of the Holly Lake metamorphic complex detritus, this would indicate that
27 the Hottah terrane was on the North American margin prior to ca. 1950 Ma. The Holly Lake
28 metamorphic complex was intruded by the Hottah plutonic complex and overlain by the Bell
29 Island Bay Group, which coincides with recorded extension-related metamorphism in the
30 Coronation margin (St-Onge and King 1986, 1987), deposition of the Akaitcho Group (Easton
31 1980, 1981a, b; Hoffman et al., 2011), and possibly the early stages of Hepburn Intrusive Suite

(Bowring 1984; Lalonde 1989). While isotopic arguments have been used to suggest that the Hepburn intrusive suite is exotic to the Slave craton (Bowring and Podosek 1989; Housh et al. 1989; Hildebrand et al. 2010a), the alternative interpretation that they intruded an attenuated basin is equally permissible (Lalonde 1989). The results and coincidences, in addition to others detailed below, allow the exploration of an alternative hypothesis for the evolution and arrival of the Hottah terrane to its present location.

An Alternative Model: Terrane Translation

The ca. 20 million year evolution from 1912.9 ± 0.8 Ma Hottah plutonic complex intrusions, through to 1895.0 ± 2.3 Ma Bloom Basalt and 1892.9 ± 1.5 Ma Grant Subgroup volcanism, is best interpreted within a framework of rapid extension, and directly comparable in its nature and duration to rifts in active continental margins. An analogue may be Baja California (Fig. 11) where the peninsula began as part of an arc terrane on the North American plate, subsequently rifted off and was transferred via transtensional shear to the Pacific plate, and was accommodated by opening of the Gulf of California marginal basin. In the last ca. 12 million years, Baja has been dextrally transported at least 300 km north and west of its origin (Fig. 11A; Umhoefer 2011; Sutherland et al. 2012; Bennett et al. 2013; also see Bryan et al. 2014), while the Gulf of California evolved on top of the previous and deceased arc system (Fig. 11A; Umhoefer 2011; Bennett et al. 2013). Basin opening and strike-slip translation have also been related to the evolution of thick slab subduction that transitions to strike-slip faulting in Alaska, also over a short time-period (Trop et al. 2012). The Taupo volcanic zone analogy is another example of upper plate extension, driven by slab roll back, within a complex arc – back-arc environment adjacent to the transition of the subduction zone to a large-scale transtensional shear zone, the Alpine fault (e.g., Wallace et al. 2012; Giba et al. 2013).

Based on the active margin translation model (extension driven strike-slip movement), the evolution of the ca. 1913 to 1893 Ma Bell Island Bay Group could record the birth of the Hottah terrane, when it was rifted off of its parent plate to form a new micro-plate (Fig. 11B). To account for the detrital zircon record in Hottah terrane (Davis et al. submitted), and the lack of Slave craton basement, the absolute starting position would have been to the south, in an active margin, possibly related to the development of the Taltson or Ksituan magmatic zones, but potentially as close as just off the southwest flank of the Slave craton (Fig. 11B; Davis et al.

submitted). In this model, rather than Hottah terrane arriving adjacent to the Slave craton above a west-dipping subduction system, it arrived by extension-driven dextral transport during basin opening. This can be envisaged in terms of the future transport of the Baja micro-plate adjacent to California (Fig. 11). If we were able to look at that margin in ca. 20 million years, it is possible Baja could be juxtaposed to central California, and look exotic to it, even though they began from the same plate. The current passive margin off western California will be equally displaced northwestward. In the Hottah terrane model, the stratigraphy of the Bell Island Bay Group (Fig. 7) is recording this extension-driven translation. The timing of the Hottah terrane arrival was sometime around ca. 1882 Ma, the time of Recluse Group foredeep sedimentation and onset of the Calderian orogeny (Hildebrand et al. 2010a; Hoffman et al. 2011). Significantly, in the southwest Slave craton, the Great Bear magmatic zone is in direct contact with the Slave craton (*sensu stricto*) along the Wopmay fault zone, and there is no Coronation margin in this region (Fig. 1 and 2A; Jackson et al. 2006). We suggest that the southwest Slave craton and overlying Coronation margin, if it existed in the southern Wopmay orogen, may have been transported northward along this fault system.

If this rift model and suggested regional correlations are correct, then the Hottah terrane represents the southwestern margin of the rifted basin, whereas ca. 1.9 Ga components of the western and southern Slave craton, along with overlying Coronation margin and potentially Wilson Island or Union Island groups in the East Arm basin (Bowring et al. 1984; Johnson 1990), represent the northeastern continental margin of the rifted basin (Fig. 11B). Circa 1.9 Ga rift-related rocks have not been discovered in the Taltson magmatic zone, but the evolution of many volcano-sedimentary units there remain understudied (K. Bethune pers. comm. 2015). Further complicating this is the fact that the western margin of the Taltson and entire Ksituan magmatic zones are hidden below the Phanerozoic sedimentary cover (Fig. 1A). The Hottah terrane translation model presented here post-dates the movement on the 1.97-1.92 Ga Great Slave Lake shear zone that bounds Slave craton and Taltson magmatic zone (Hofmann 1987; Hanmer et al. 1992) and pre-dates the ca. 1.84-1.75 Ga MacDonald Fault movement (Fig. 1; Ross 2002). Further studies are required to demonstrate if it is geometrically possible to accommodate the movement of Hottah terrane as suggested. Deciphering the destination of the removed western Slave craton and Coronation margin, and evaluating putative correlations to the East Arm basin, are also key factors for testing this model. A currently untestable implication is

what effects such a model has on the correlation and assignments of the terranes across the MacDonald Fault in the subsurface of Alberta (Fig. 1A; Villeneuve et al. 1993; Ross 2002).

Oblique east-dipping subduction has been suggested to account for Great Bear magmatism at 1.875 Ga (Hildebrand et al. 1987) and is further supported in this study. However, if this subduction geometry had been established by ca. 2.0 Ga or earlier, rather than at 1.88 Ga, then Hottah terrane rocks could have evolved marginal to terranes to the south, such as Taltson or Ksituan magmatic zones (Figs. 1 and 11), accounting for the detrital zircon record in the Hottah terrane (Davis et al. submitted). While a ca. 1.88 Ga micro-continent – continent collision model (Fig. 3A) is not entirely ruled out in this study, a one-sided, long-lived subduction system would undoubtedly undergo tectonic switching and migration, accounting for the petrological progression in magmatism and the evolution of the Hottah terrane and Great Bear magmatic zone, and potentially a migration from south to north (in present coordinates), finally impinging on the Slave craton margin and resulting in the Great Bear arc flare-up. An always east-dipping subduction system also helps explain the east-dipping stacked lithospheric mantle under the western Slave craton and Great Bear magmatic zone (Cook et al. 1999; Cook and Erdmer 2005; Wu et al. 2005; Oueity and Clowes 2011), including the frozen oceanic plate that was likely related to the late arrival of the Nahanni terrane to the west (Fig. 1B). Late stage, NE-striking, brittle-dextral faulting, such as the dextral McDonald Fault, continued to the Cordilleran thrust front (Fig. 1A; Pilkington et al. 2000; Aspler et al. 2003; Hayward and Corriveau 2014) and therefore brittle deformation post-dated arrival of the Nahanni terrane (cf. Hayward and Corriveau 2014). It is speculated here that these faults were related to some tectonic adjustments even further west, or perhaps even related to the Trans-Hudson orogen to the southeast.

Conclusions

This study provides new geochronological and geochemical results for the exposed Hottah terrane and Great Bear magmatic zone of the northwest Canadian Shield that further highlight an evolving tectonostratigraphy from ca. 1950 to 1850 Ma (Hildebrand et al. 2010a, b; Davis et al. submitted). The oldest bedrock is the ca. 1950 Ma Holly Lake metamorphic complex, which was intruded by the Hottah plutonic complex at ca. 1930 Ma and 1913 Ma. The youngest of these plutons were unroofed and unconformably overlain by Beaverlodge Lake sandstone and the Bell Island Bay Group, beginning with the ca. 1906 Ma Zebulon Formation

1 rhyolite-dominated subaerial volcanic rocks. The Conjuror Bay Formation marginal marine
2 quartz arenite was deposited on top and capped by the ca. 1895 Ma Bloom Basalts and lesser
3 rhyodacite. These basalts can be chronologically and geochemically correlated with the ca. 1893
4 Ma Grant Subgroup, which overly mylonitized Archean granites east of the Wopmay fault. The
5 stratigraphic and petrological evolution of the Hottah terrane from 1913 to 1893 Ma is
6 compatible with an evolving arc that rapidly evolved from subduction to a transtensional
7 environment. The results of this study and detrital zircon data in Davis et al (submitted) suggest
8 the Hottah terrane began on the margins of the Taltson (or Ksituan) magmatic zones, south of its
9 current resting place, and was dextrally transported by transtensional shear movement during
10 basin opening as recorded by the Bell Island Bay Group. Transtensional shear movement, or
11 terrane translation (Umhoefer, 2011) may prove a valuable alternative to subduction-collision
12 models invoked for the tectonic evolution of relict terranes globally.

13 The Great Bear arc flare-up began at ca. 1876 Ma and continued through ca. 1869 Ma
14 and was driven by oblique east-dipping subduction (Hildebrand 1981; Hildebrand et al. 1987a,
15 2010a). Voluminous felsic-intermediate plutonism followed and the bimodal Sloan Group
16 erupted at 1862 Ma. Post-orogenic, ca. 1855 Ma Rapakivi-textured granitoids and ca. 1845
17 syenogranites were the last magmatic pulses (Bowring 1984; Gandhi et al. 2001). The results and
18 models presented for the Hottah terrane and Great Bear magmatic zone are fully compliant with
19 the geophysical models that indicate an east-dipping stacked lithospheric mantle (e.g., Cook et
20 al. 1999). On a broader scale the age continuum of the Hottah terrane and Great Bear magmatic
21 zone matches the complex interplay of continental-scale subduction processes from ca. 2.0 Ga to
22 the amalgamation of the supercontinent Nuna.

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Figure Captions

Figure 1. A) Tectonic subdivisions of the western Canadian Shield (Modified after Hoffman 1988; Villeneuve et al. 1993; Ross 2002; Aspler et al. 2003). Solid thick black line between Great Bear Lake and western Saskatchewan demarcates the eastern edge of the Phanerozoic platformal cover. Grey solid lines represent major faults. Blue line is the seismic line. B) Composite block diagram comprising the interpreted seismic section (vertical section; modified from Cook and Erdmer 2005) and interpreted domain subdivisions (horizontal plane; modified from Aspler et al. 2003). Blue line (also shown in Fig. 1A) represents seismic line.

Figure 2. A) Geological map of the western Wopmay orogen. Locations are DeVries Lake (DL), Grant Lake (GL), Ham Lake (HL), Leith Ridge (LR), Rebesca Lake (RL), Swan Lake (SL), Treasure Lake Group (TLG; this is the type section of Gandhi and van Breemen, 2005), and Zebulon River (ZR). B) Geological map of exposed rocks assigned to Hottah terrane, including units of the Bell Island Bay Group, and supracrustal rocks of the younger Great Bear magmatic zone (GBMZ). Arrows point to sample locations discussed in text; Beaverlodge Lake andesite (BLa), Hottah plutonic complex plagioclase-quartz porphyry (Hp), Hottah plutonic complex K-feldspar phenocrystic granite (Hg), Hottah plutonic complex tonalite (Ht), Stairs Bay rhyolite (SBr), and Zebulon Formation rhyolite (Zr). Modified after Henderson (1949), McGlynn (1979), Hildebrand and Roots (1985), Reichenbach (1991), Hoffman and Hall (1992), and Jackson and Ootes (2012).

Figure 3. Previously proposed tectonic models for the evolution of the Hottah terrane and Wopmay orogen. A) Hottah terrane as an exotic arc and micro-continent that collided with the western Slave craton at ca. 1880 Ma (modified after Bowring and Grotzinger 1992). B) Hottah terrane developed as a ca. 1940-1885 Ma arc that was separated from Slave craton by a back-arc basin (modified after Reichenbach 1991; Bowring and Grotzinger 1992). HIS – Hepburn Intrusive Suite; MIZ – Metamorphic internal zone; TK – Turmoil klippe; WFZ – Wopmay fault zone.

Figure 4. Outcrop photographs of selected units discussed in this paper. A) Holly Lake metamorphic complex migmatitic semipelites (Ms) cross-cut by granodiorite (Gr) and younger pegmatite (P) dykes at Leith ridge (Fig. 2). Inset photograph shows compositional layering (dashed white line), likely primary, in migmatized semi-pelite that is cut by quartz vein that was later boudinaged. B) Complex intrusive breccia of Hottah plutonic complex, including early gabbro (Gab), diorite (Drt), and a youngest phase of granodiorite (Grdt), near sample location Ht (Fig. 2B). C) Unconformity (dashed white line) between Beaverlodge Lake sandstone and underlying Hottah plutonic complex, east side of Bell Island (Fig. 2B). Symbols: flat lying, upright-facing bedding in the coarse conglomerate on top (white barbed symbol) are at a high angle to moderately developed foliation, defined by feldspar and micas, in the underlying granite (black foliation symbol). D) Coarsely spherulitic (S) rhyolite with quartz-filled lithophysae (L) of the Zebulon Formation at sample location Zr (Fig. 2B, 5E; Table 1). E) Unconformities at Beaverlodge Lake where Zebulon Formation (ZF) overlies Hottah plutonic complex plagioclase-quartz porphyry and quartz arenite of the Conjuror Bay Formation (CBF) overlies both units. Dashed white lines depict unconformities. White diamond is U-Pb sample location of Hottah plagioclase-quartz porphyry (Hp; Fig. 2B, 5B; Table 1); black diamond is detrital zircon sample location of Conjuror Bay Formation quartz arenite (Davis et al. submitted). View is to the north, person for scale is circled. Inset shows unconformity of quartz arenite on heavily altered Hottah porphyry with blocks of altered porphyry at the base of quartz arenite. Hammer for scale (circled). F) Moderately flattened fiamme-bearing rhyolitic ignimbrite from Swan Lake at sample location SL (Fig. 2A, 5J; Table 1). E) Quartz and feldspar crystal-rich rhyolite from Stairs Bay, southeast Hottah Lake at sample location SBr (Fig. 2B, 5L; Table 1).

Figure 5. Concordia diagrams of the U-Pb zircon results using TIMS (A-L) and SHRIMP (M). A) Hottah plutonic complex K-feldspar porphyritic granite, B) quartz-plagioclase porphyry, C) strongly deformed granodiorite, D) tonalite, E) Zebulon Formation spherulitic rhyolite, F) Bloom rhyodacite, G) Grant Subgroup rhyodacite, H) Great Bear magmatic zone rhyolitic ash tuff, I) Beaverlodge Lake fragmental andesite, J) Swan Lake rhyolitic ignimbrite, K) deformed hornblende-biotite granodiorite, L) Stairs Bay rhyolitic quartz-feldspar crystal tuff. M) SHRIMP U-Pb zircon results from Dumas Group rhyodacite. Data-point error ellipses are 2σ . Results are in Tables 1 and 2 and analytical techniques are in Appendix I.

Figure 6. ϵNd^T versus Age (Ma) plot. The data from Bowring and Podosek (1989) only includes samples from west of Wopmay fault zone. Data from the Fort Simpson terrane granites are from Villeneuve et al. (1991). The Slave array is projected from 2.6 to 2.58 Ga granites in the southwestern Slave craton (Yamashita et al. 1999) and any older or more evolved Slave granitoids would project below this array. Upper boundary on the Depleted mantle is from Goldstein et al. (1984) and the lower boundary from DePaolo (1981).

Figure 7. Stratigraphy of the Hottah terrane – Great Bear volcano-plutonic continuum (no vertical scale implied). Age determinations are in Ma ($\pm 2\sigma$ uncertainty); Great Bear plutonic ages are not shown. Previously reported ages are italicized and from Bowring (1984), Reichenbach (1991), Gandhi et al. (2001), and Bennett and Rivers (2006a). The detrital zircon (DZ) maximum depositional ages for the Holly Lake metamorphic complex are from Davis et al. (submitted) and the Treasure Lake Group are from Gandhi and van Breemen (2005) and Bennett and Rivers (2006b). For Archean basement ages see Jackson et al. (2013) and Bowring (1984).

Figure 8. Geochemical plots. A) $\text{CaO} + \text{Na}_2\text{O} - \text{Al}_2\text{O}_3 - \text{K}_2\text{O}$, or chemical index of alteration (CIA) plot of Hottah plutonic complex and Zebulon Formation demonstrating the aluminium and potassium enrichment of the units relative to ‘typical’ unaltered end-member examples (open circles). B) Maniar and Piccoli (1989) plot to illustrate that Zebulon Formation rhyolites and some Hottah plutonic complex samples fall on a ‘mixing line’, interpreted as a result of alteration of the mobile major elements. C) Immobile trace element ratio diagram with all units in this study plotted. This diagram is preferred over the total alkali – silica diagram because the alkali elements in the Zebulon Formation and some Hottah plutonic complex samples have been compromised due to alteration (see Fig. 8A-B). Diagram from Pearce (1996) and modified after Winchester and Floyd (1977). D) Zr – TiO_2 plot. Lines are 20% increment end-member mixing trends (denoted by points on line). Dashed line is Bloom-Grant basalts. Two end-member mixing lines are shown for Zebulon Formation basalts and basaltic andesites, and one for Zebulon Formation rhyolite samples. Grey shaded area denotes distinct high-Zr dacite-andesite-trachyte samples. E) Immobile trace element ratio versus SiO_2 (Winchester and Floyd, 1977) for all Zebulon Formation samples. Some of the samples plot as andesites but are better termed

andesitic basalt as they may be enriched in SiO_2 either from alteration, or incorporation of amygdales (see Figure 4C). Samples plotted include those from this study and from Reichenbach (1991). Analytical techniques are in Appendix I and Table A1. Relevant geochemical features are summarized in Table 3. See text for further discussion.

Figure 9. Chondrite and Primitive Mantle-normalized plots (Sun and McDonough 1989). Symbols as in Figure 8. A) Hottah plutonic complex. Vertical grey bars in Primitive Mantle-normalized diagram indicate those elements suspected to be mobile during alteration and are removed from the subsequent plots (Rb, Ba, U, K, Sr). Circa 1913 Ma tonalite (Ht) and ca. 1930 Ma granite and feldspar-quartz porphyry (Hg-Hp). B) Zebulon Formation rhyolite. Pleistocene Long Valley Caldera data are from Moulton et al. (2011). C) Zebulon Formation basalt. D) Bloom Basalt (squares) and Grant Subgroup basalt (circles). Closed circles are samples from north of Grant Lake (GL) and open circles are from samples north of Rebesca Lake (RL; Fig. 2A). Heavy dashed grey line on Primitive Mantle-normalized plot is Zebulon Formation basalt for comparison. E) Bloom rhyodacite and Grant Subgroup rhyodacite. Heavy solid black line on Primitive Mantle normalized plot is Bloom Basalt for comparison. F) Great Bear magmatic zone volcanic rocks, including ca. 1873 Ma LaBine Group andesite (BLa) from the Beaverlodge Lake area, ca. 1862 Ma Sloan Group basalt (SBr-b) and rhyolite (SBr) from the Stairs Bay area, ca. 1869 Ma Faber Group ignimbrite (SL) from the Swan Lake area (Fig. 2). Grey shaded band is andesitic to rhyolitic volcanic rocks and intrusive porphyries from the Fab Lake area (n=26; Azar 2007). All other samples plotted are from this study.

Figure 10. Tectonic discrimination diagrams. A) Nb-Y plot (Pearce et al. 1984) for Hottah plutonic complex samples and Great Bear (GB) magmatic zone volcanic rocks (includes data from Azar 2007); fields are ocean ridge granite (ORG), volcanic arc granite (VAG), syn-collisional granite (syn-COLG), and within plate granite (WPG). B) V-Ti discrimination diagram for basalts (Shervais 1982); fields are back-arc basin (BAB), mid-ocean ridge basalt (MORB), boninite (BON), island arc tholeiite (IAT), and alkaline basalt. Plot includes Bloom and Zebulon Formation basalt data from Reichenbach (1991). C) Zr-(Ti/100)-(Y*3) discrimination diagram for basalts with CaO+MgO 12-20% (Pearce and Cann 1973). Fishtrap Lake gabbro data (Reichenbach 1991) and Grant Subgroup basalt data (Type 2 in Easton 1980) are overlapped by

the Bloom-Grant basalt and not visible. Figure includes Bloom and Zebulon Formation basalt data from Reichenbach (1991). D) Th-(Hf/3)-(Nb/16) diagram (after Wood, 1980). Only data from this study are plotted as the data in Reichenbach (1991) does not contain Th values. E) Th/Yb-Ta/Yb discrimination diagram (after Pearce 1982). Western North America arc (WNArc) and back-arc basin basalt (BABB) fields from Cole and Stewart (2009), fore-arc basalt (FAB) field from Regan et al. (2010), and ocean island basalt (OIB), enriched-mid-ocean ridge basalt (E-MORB), and normal mid-ocean ridge basalt (N-MORB) from Sun and McDonough (1989).

Figure 11. A) Transtensional shear evolution of Baja California and Gulf of California. Baja California began as an arc on the North American plate and was transferred to the Pacific plate during rapid rifting and basin opening in an active margin, over the last 20 Ma (adapted from Bennett et al. 2013). Baja California could move another 500 to 1000 km northward during the next 20 Ma. B) Application of the Baja California model to the 1913 to 1893 Ma evolution of the Hottah terrane rocks, where the Hottah plutonic complex represents the root of a volcanic arc that began on the North American plate (potentially the Taltson or Ksituan magmatic zones to the south) and was dextrally translated off this parental crust during arc rifting and basinal opening. The Hottah plutonic complex was overlain by the Zebulon Formation (ZF) subaerial, rhyolite-dominated volcanism in a rifting arc between ca. 1906 and 1900 Ma. Continued rifting led to drowning of the subaerial volcanic rocks and deposition of the Conjuror Bay Formation (CBF) quartz arenite at ca. 1900 Ma. Further attenuation led to the eruption of the subaqueous Bloom Basalt (BB) in the west and Grant Subgroup (GS) in the east between ca. 1900 and 1893 Ma in a either rifted arc or marginal basin setting. In this model transtensional shear related to basinal opening would eventually transfer the Hottah terrane northward along a proto-Wopmay fault, arriving adjacent to the Slave craton at ca. 1880 Ma (not shown).

Appendix I – Analytical Methods

U-Pb Zircon Geochronology

Zircon fractions from 11 samples were prepared using the annealing and chemical leaching technique modified from that described by Mattinson (2005) and isotopic and U-Pb compositional data were determined by isotope dilution thermal ionization mass spectrometry (CA-ID-TIMS) at the Geochronology laboratory, Geological Survey of Canada. Results are tabulated in Table 1. The samples were annealed at 1000°C for 48 hours and then leached for 12–16 hours, in HF-HNO₃ at 180°C in Savillex 3 mL PFA capsules within a Parr digestion vessel. Zircon fractions from one sample (09vj171) were prepared by mechanical abrasion method of Krogh (1982; Table 1). Sample dissolution and chemical methods are described in Parrish et al. (1987). Individual crystals were selected under binocular microscope to avoid inclusions and other imperfections, spiked with a mixed ²⁰⁵Pb-²³³U-²³⁵U tracer solution calibrated to ±0.1% against a gravimetric solution, and dissolved in high-pressure bombs in HF-HNO₃. Data reduction and error propagation follow methods outlined in Roddick (1987). Uranium and Pb isotopic ratios were measured using a Triton mass spectrometer operated in either static multi-collection mode or using a secondary electron multiplier and ion counting system. The ²⁰⁵Pb, ²⁰⁶Pb, ²⁰⁷Pb, and ²⁰⁸Pb isotopes were measured simultaneously in Faraday collectors, with ²⁰⁴Pb in an axial secondary electron multiplier. Faraday-multiplier gain was monitored and corrected by peak jumping ²⁰⁵Pb into the axial cup. A Pb mass fractionation correction of 0.1 ± 0.04%/amu was applied as determined by replicate analyses of the NBS981 standard (Table 1). U fractionation was corrected using the ²³³U-²³⁵U double spike and was typically in the range of 0.12%/amu. Dead time for the ion counting system was determined by replicate analyses of the NBS982 solution. Accuracy and reproducibility of the ion counting measurements were monitored by analyses of the GSC standard 6266 with a ²⁰⁶Pb/²³⁸U ratio of 0.0963.

Zircon fractions from one sample (09vj148c) were prepared for sensitive high resolution ion microprobe analysis (SHRIMP II) at the Geological Survey of Canada. Results are tabulated in Table 2. SHRIMP analytical procedures followed those described by Stern (1997). Briefly, zircons were cast in 2.5 cm diameter epoxy mounts along with fragments of the GSC laboratory standard zircon (z6266, with ²⁰⁶Pb/²³⁸U age = 559 Ma, Stern and Amelin 2003). The mid-sections of the zircons were exposed using 9, 6, and 1 μm diamond compound, and the internal features of the zircons (such as zoning, structures, alteration, etc.) were characterized in back-

1 scattered electron mode (BSE) utilizing a Zeiss Evo 50 scanning electron microscope. Mount
2 surfaces were evaporatively coated with 10 nm of high purity Au. Analyses were conducted
3 using an $^{16}\text{O}^-$ primary beam, projected onto the zircons at 10 kV. The count rates at eleven
4 masses including background were sequentially measured over 5 scans with a single electron
5 multiplier and a pulse counting system with deadtime of 23 ns. Off-line data processing was
6 accomplished using SQUID version 2.22. The 1σ external errors of $^{206}\text{Pb}/^{238}\text{U}$ ratios reported in
7 the data table incorporate a minimum $\pm 1.0\%$ error in calibrating the standard zircon (see Stern
8 and Amelin 2003). Pb isotopic values were monitored by analyses of an in house standard with a
9 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2679.6 ± 0.3 Ma); a small mass bias correction (1.0029) was applied for
10 samples 10232 and 10320. Common Pb correction utilized the Pb composition of the surface
11 blank (Stern 1997). Isoplot v. 3.00 (Ludwig 2003) was used to generate concordia plots and
12 calculate weighted means. The error ellipses on the concordia diagrams (Fig. 5) and the weighted
13 mean errors are reported at 2σ .

14
15 ***Major and Trace-element Geochemistry***

16 Major and trace-element geochemistry (Tables 3 and A1) was completed on fresh and clean
17 samples at Acme Analytical labs between January and March 2012. Samples were crushed and
18 pulverized using mild steel. Major and HFSE were analyzed by pressed pellet XRF (Code 4X).
19 For LILE, HFSE, and REE, powdered samples were mixed with Lithium metaborate/tetaborate
20 and fused in a furnace, followed by cooling and dissolving of the bead in nitric acid, and then
21 analyzed by ICP-MS (Code 4B). Gallium values were determined by subjecting the prepared
22 sample powders to a HF-aqua regia leach in a hot water bath, cooling, then analyzing by ICP-ES.
23 For the base and precious metals (including Pb) a separated 0.5g split was digested in Aqua
24 Regia and analysed by ICP-MS (Code 1DX). International standards were run once for every ten
25 unknowns. Assessment of the data using the pulp and preparation duplicates and in-house and
26 international standards indicates the major element oxides have less than 2% error, except for
27 oxides with lower concentrations which may have up to 5% error. High-field strength elements
28 and REE are quoted at better than 10%, although in general are better than 5%, the higher error
29 in relation to lower concentrations of these elements in some samples. Sample location
30 coordinates and minimum detection limits are in Table A1.

Samarium-Neodymium(Sm-Nd) Isotope Geochemistry

Samarium-Neodymium isotopic analyses were completed at Carleton University, Ottawa, ON, Canada between January and March, 2012. Between 100 and 300 mg of powder from each sample was weighed into a screw-cap Teflon vial, to which a mixed ^{148}Nd - ^{149}Sm spike was added. The powder-spike mixture was dissolved in HNO_3 -HF, which was further attacked with HNO_3 and HCl until no residue remained visible. The bulk REE was separated using cation chromatography (Dowex 50-X8). The REE-bearing residue was then dissolved in 0.26N HCl and loaded into an Eichrom chromatographic column containing Teflon powder coated with HDEHP [di(2-ethylhexyl)] orthophosphoric acid (Richard et al. 1976). Neodymium was eluted using 0.26N HCl , followed by Sm in 0.5N HCl .

Total procedural blanks for Nd are <200 picograms. Concentrations are precise to $\pm 1\%$, but $^{147}\text{Sm}/^{144}\text{Nd}$ ratios are reproducible to $\pm 0.5\%$. Samples were loaded with 0.3N H_3PO_4 on one side of a Re double filament assembly, and run at temperatures of 1750-1800°C in a 9-cup ThermoFinnigan TRITON T1 multicollector mass spectrometer. Isotope ratios are normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$. Analyses of the USGS standard BCR-1 yield Nd = 29.02 ppm, Sm = 6.68 ppm, and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512668 \pm 0.000020$ (n=4). Over the last two years the La Jolla standard average is $^{143}\text{Nd}/^{144}\text{Nd} = 0.511826 \pm 0.000013$. Epsilon values at time T are calculated using the following relation:

$$\epsilon\text{Nd}^T = [(^{143}\text{Nd}/^{144}\text{Nd}_{\text{sample}}^T / ^{143}\text{Nd}/^{144}\text{Nd}_{\text{CHUR}}^T) - 1] * 10000$$

where CHUR is the Chondrite Uniform Reservoir and T is the time the rock was formed.

Depleted mantle model ages T_{DM} ages were calculated for felsic-intermediate rocks with $^{147}\text{Sm}/^{144}\text{Nd} < 0.1500$ using a modified method of Goldstein et al. (1984) assuming a modern upper mantle with $^{147}\text{Sm}/^{144}\text{Nd} = 0.214$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.513115$ (Tables 3 and A2).

Figure 1. Ootes et al., Hottah Terrane

Figure 2: Ootes et al., Hottah Terrane

Figure 3: Ootes et al., Hottah Terrane

Figure 4: Ootes et al., Hottah Terrane

Figure 5. Ootes et al., manuscriptcentral.com/cjes-pubs

Figure 5 cont'd. Ootes et al., Hotian Terrane

Figure 5 cont'd. Ootes et al., Hottah Terrane

Figure 6. Ootes et al., Hottah Terrane

Figure 7. Ootes et al., Hottah Terrane

Figure 8: Ootes et al. Hottah Terrane

<https://mc.manuscriptcentral.com/cjes-pubs>

Figure 9: Ootes et al., Hottah Terrane

E **F**

Figure 9 cont'd: Ootes et al., Hottah Terrane

Figure 10: Ootes et al., Hottah terrane
<https://mc06.manuscriptcentral.com/cjes-pubs>

Figure 11: Ootes et al., Hottah terrane

Table 1. TIMS U-Pb zircon results

		Radiogenic Ratios (± 1sigma error absolute) ³												Age (Ma, ± 2σ error) absolute					
Fraction	Description ¹	# gr	Size	Wt.	U	Pbr ²	²⁰⁶ Pb/ ²⁰⁴ Pb	Pbc ²	²⁰⁸ Pb/ ²⁰⁶ Pb	²⁰⁷ Pb/ ²³⁵ U	²⁰⁶ Pb/ ±	Correlation Coefficient	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁷ Pb/ ±	Discord (%)				
Porphyritic granite (Hg) - 09lo9078A (Z10084)																			
442095m E, 7177812m N																			
A3-1 (Z)	Z,Co,Clr,fIn,Eu,Pr,Lch,NM1°	1	100	2.3	229	86	6135	1.8	0.16	5.546	0.006	0.3407	0.0003	0.9359	0.11806	0.00005	1927.1	1.6	2.22
A6-1 (Z)	Z,Co,Clr,fIn,Eu,Pr,Lch,NM1°	1	140	2.5	261	100	5716	2.5	0.16	5.665	0.007	0.3475	0.0003	0.9326	0.11824	0.00005	1929.8	1.6	0.43
B10-1 (Z)	Z,Co,Alt,fFr,fIn,Eu,Lch,M1°	1	160	6.0	86	32	3938	2.8	0.14	5.661	0.007	0.3474	0.0003	0.9060	0.11820	0.00006	1929.2	1.9	0.43
B15-1 (Z)	Z,Co,Clr,Eu,Pr,Lch,M1°	1	170	2.4	182	70	1128	8.5	0.16	5.680	0.009	0.3485	0.0004	0.8185	0.11821	0.00012	1929.4	3.5	0.11
B8-1 (Z)	Z,Co,Alt,nFr,Eu,Pr,Lch,M1°	1	110	1.9	292	112	11059	0.4	0.16	5.662	0.007	0.3471	0.0003	0.9068	0.11832	0.00006	1931.0	1.8	0.63
Hypabyssal porphyry (Hp) - 09lo9039A (Z10083)																			
*443456m E, 7179171m N																			
A10-1 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM10°	2	100	1.4	154	58	2446	1.9	0.14	5.681	0.007	0.3483	0.0003	0.8901	0.11831	0.00007	1930.9	2.2	0.27
A12-1 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM10°	1	100	2.1	81	30	895	4.2	0.14	5.718	0.015	0.3489	0.0009	0.6439	0.11889	0.00026	1939.5	7.9	0.62
A12-2 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM5°	1	170	1.0	115	44	1604	1.6	0.15	5.698	0.008	0.3489	0.0004	0.8239	0.11844	0.00010	1932.8	3.0	0.21
A14-1 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM5°	1	120	1.0	121	45	1604	1.6	0.13	5.666	0.008	0.3475	0.0004	0.8741	0.11828	0.00008	1930.4	2.5	0.48
A16-1 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM5°	1	110	2.1	147	57	1280	5.3	0.18	5.685	0.009	0.3489	0.0004	0.8174	0.11819	0.00011	1929.1	3.3	0.00
A6-1 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM5°	1	180	2.1	105	40	3524	1.4	0.14	5.681	0.007	0.3485	0.0003	0.8753	0.11823	0.00007	1929.7	2.1	0.13
Strongly deformed hornblende-biotite granodiorite (ZR) - 11lo4156a (Z10555)																			
*452888m E, 7208251m N																			
A15-1 (Z)	Z,Co,Clr,nIn,Eu,Pr,Lch,M1°	1	150	2.9	66	25	1245	3.4	0.13	5.650	0.009	0.3474	0.0004	0.8379	0.11797	0.00011	1925.7	3.3	0.23
A16-1 (Z)	Z,Co,Clr,cln,Eu,Pr,Abr,NM1°	1	200	3.9	88	33	1800	4.2	0.14	5.661	0.008	0.3476	0.0003	0.8631	0.11810	0.00008	1927.7	2.5	0.26
A20-1 (Z)	Z,Co,Clr,fIn,Eu,Pr,Lch,NM1°	1	150	2.5	68	25	1959	1.9	0.13	5.640	0.008	0.3471	0.0004	0.8602	0.11783	0.00008	1923.5	2.5	0.16
B15-1 (Z)	Z,Co,Clr,cln,Eu,Pr,Lch,M1°	1	150	5.5	75	28	4769	1.9	0.12	5.659	0.007	0.3475	0.0003	0.9262	0.11812	0.00006	1928.0	1.7	0.33
B16-1 (Z)	Z,Co,Clr,cln,Eu,Pr,Lch,M1°	1	200	12.6	121	47	10053	3.3	0.18	5.646	0.006	0.3467	0.0003	0.9366	0.11811	0.00005	1927.9	1.6	0.54
B20-1 (Z)	Z,Co,Clr,cln,Pr,Sub,Lch,NM1°	1	125	3.4	119	46	3502	2.5	0.18	5.675	0.007	0.3483	0.0004	0.9215	0.11817	0.00006	1928.8	1.9	0.14
Foliated tonalite (Ht) - 11lo4157a (Z10556)																			
*435738m E, 7206954m N																			
A10-1 (Z)	Z,Co,Clr,nFr,Eu,Pr,Lch,M0°	1	200	6.4	98	36	6342	2.1	0.12	5.564	0.006	0.3444	0.0003	0.9316	0.11718	0.00005	1913.7	1.6	0.37
B10-1 (Z)	Z,Co,Clr,fFr,Eu,Pr,Lch,M0°	1	200	10.3	103	38	3304	6.9	0.11	5.561	0.007	0.3444	0.0003	0.9228	0.11711	0.00006	1912.6	1.9	0.28
B10-2 (Z)	Z,Co,Clr,fFr,Eu,Pr,Lch,M0°	1	225	20.6	108	39	7660	6.2	0.11	5.549	0.006	0.3437	0.0003	0.9399	0.11710	0.00005	1912.4	1.6	0.49
B5-1 (Z)	Z,Co,Clr,nFr,Fg,Lch,M0°	1	150	7.3	94	34	5578	2.7	0.12	5.555	0.007	0.3440	0.0003	0.9300	0.11712	0.00006	1912.7	1.7	0.41
B5-2 (Z)	Z,Co,Clr,nFr,Eu,Pr,Lch,M0°	1	200	21.6	109	40	11447	4.4	0.13	5.548	0.006	0.3435	0.0003	0.9405	0.11713	0.00005	1912.9	1.5	0.56
C5-1 (Z)	Z,pBr,Clr,fFr,Eu,Pr,Lch,M0°	1	150	5.4	89	32	1456	7.1	0.12	5.540	0.008	0.3429	0.0003	0.8560	0.11717	0.00009	1913.5	2.8	0.78
Spherulitic rhyolite (Zr) - 10lo1116a (Z10321)																			
*437934m E, 7174560m N																			
AB16-1 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM3°	1	90	1.2	109	40	2279	1.2	0.14	5.508	0.007	0.3426	0.0004	0.8830	0.11658	0.00008	1904.5	2.3	0.32
B16-2 (Z)	Z,Co,Clr,Eu,St,Lch,M3°	1	100	1.0	119	45	233	11.9	0.16	5.582	0.028	0.3461	0.0007	0.5964	0.11697	0.00048	1910.4	14.8	-0.35
B16-4 (Z)	Z,Co,Clr,Eu,St,Lch,M3°	1	80	0.9	98	36	3189	0.4	0.14	5.489	0.007	0.3413	0.0004	0.8912	0.11665	0.00007	1905.6	2.2	0.77
B16-5 (Z)	Z,Co,Clr,Eu,St,Lch,M3°	1	82	0.8	93	34	1185	1.4	0.12	5.529	0.010	0.3434	0.0005	0.8250	0.11679	0.00012	1907.6	3.5	0.29
Rhyodacite (Br) - 11Lo4153a (Z10553)																			
*435076m E, 7222942m N																			
A15-2 (Z)	Z,Co,Clr,fFr,Eu,Mf,Tip,Lch,NM10°	1	100	1.1	132	47	1746	1.8	0.10	5.431	0.008	0.3399	0.0004	0.8911	0.11586	0.00008	1893.3	2.4	0.42
A16-1 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM10°	1	230	4.7	90	37	3277	2.8	0.26	5.525	0.007	0.3453	0.0003	0.8783	0.11603	0.00007	1896.0	2.1	-0.99
A16-2 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM10°	1	215	2.9	47	23	1539	1.9	0.51	5.461	0.008	0.3413	0.0004	0.8419	0.11605	0.00010	1896.3	3.0	0.21
A20-2 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM10°	1	175	2.0	123	53	1484	3.6	0.34	5.457	0.009	0.3414	0.0004	0.8500	0.11594	0.00010	1894.5	3.0	0.07
Rhyodacite (GL) - 09vj171A (Z9951)																			
*520671m E, 7207791m N																			
A1A (Z)	Z,Co,Clr,Eq,Eu,Abr,NM1°	1	110	1.3	177	67	2884.5	1.7	0.17	5.419	0.008	0.3394	0.0004	0.9212	0.11581	0.00007	1892.6	2.1	0.55
A3A (Z)	Z,Co,Clr,Eu,Pr,Abr,NM1°	1	105	0.9	153	57	1270.6	2.3	0.17	5.407	0.012	0.3387	0.0006	0.9237	0.11579	0.00010	1892.3	3.0	0.72
A4A (Z)	Z,Co,Clr,Eq,Eu,Abr,NM1°	1	80	0.6	128	48	346.2	4.9	0.16	5.477	0.018	0.3422	0.0005	0.6747	0.11608	0.00030	1896.7	9.3	-0.04
B1 (Z)	Z,Co,Clr,fFr,Eu,Pr,Abr,M3°	1	90	2.0	88	33	953.2	4.0	0.17	5.440	0.009	0.3405	0.0004	0.7930	0.11587	0.00012	1893.4	3.8	0.26
B2 (Z)	Z,Co,Clr,Eu,Pr,Abr,M3°	1	100	1.9	125	47	1089.9	4.7	0.17	5.423	0.009	0.3394	0.0003	0.7793	0.11589	0.00012	1893.7	3.7	0.62
Rhyolite ash tuff (DL) - 08vj19a (Z10813)																			

***512484m E, 7130288m N**

A20-1 (Z)	Z,Co,Clr,fFr,El,Mf,Pr,Lch,M1°	1	180	1.1	92	35	1320.4	1.6	0.19	5.331	0.008	0.3371	0.0004	0.8505	0.11470	0.00010	1875.2	3	0.16
A20-2 (Z)	Z,Co,Clr,El,Mf,Pr,Lch,M1°	1	110	0.6	76	34	735.9	1.3	0.41	5.328	0.012	0.3367	0.0006	0.7706	0.11478	0.00016	1876.5	5.1	0.36
A20-3 (Z)	Z,Co,Clr,Mf,Pr,Lch,M1°	1	120	1.1	83	29	600.9	3.3	0.10	5.365	0.013	0.3381	0.0004	0.7144	0.11507	0.00020	1881	6.1	0.20

Fragmental andesite (BLa) - 10vj186a (Z10322)***447400m E, 7177920m N**

B12-2 (Z)	Z,pBr,Clr,Eu,Pr,Lch,M1°	1	186	5.3	178	65	6953	2.9	0.14	5.312	0.006	0.3361	0.0003	0.9353	0.11461	0.00005	1873.8	1.6	0.35
B12-3 (Z)	Z,pBr,Clr,Eu,Pr,Lch,M1°	1	128	2.3	85	30	3343	1.2	0.11	5.297	0.007	0.3355	0.0003	0.8985	0.11449	0.00007	1871.9	2.1	0.42
C16-1 (Z)	Z,Co,Clr,Eu,Pr,Lch,M3°	1	140	1.0	125	45	2154	1.1	0.14	5.292	0.008	0.3354	0.0004	0.8643	0.11446	0.00009	1871.4	2.8	0.44
C16-2 (Z)	Z,Co,Clr,Eu,Pr,Lch,M3°	1	138	2.4	116	41	5077	0.8	0.12	5.304	0.007	0.3359	0.0004	0.9078	0.11450	0.00007	1872.0	2.1	0.30
C16-3 (Z)	Z,Co,Clr,Eu,Pr,Lch,M3°	1	120	1.9	203	73	4284	1.9	0.14	5.292	0.006	0.3352	0.0003	0.8694	0.11452	0.00007	1872.4	2.2	0.55

Rhyolite ignimbrite (SL) - 10lo1131b (Z10324)***462754m E, 7201067m N**

A20-2 (Z)	Z,Co,Clr,Eu,Pr,Lch,M1°	1	111	1.4	148	52	1905	1.5	0.11	5.291	0.007	0.3355	0.0004	0.8417	0.11440	0.00009	1870.4	2.7	0.35
A25-1 (Z)	Z,Co,Clr,Eq,Eu,Pr,Lch,M1°	1	118	2.5	154	56	14427	0.6	0.13	5.287	0.006	0.3357	0.0003	0.9301	0.11424	0.00005	1867.9	1.6	0.13
A25-3 (Z)	Z,Co,Clr,Eu,Pr,Lch,M1°	1	167	3.1	156	57	17252	0.4	0.15	5.325	0.006	0.3366	0.0003	0.9169	0.11474	0.00006	1875.8	1.7	0.34
A25-4 (Z)	Z,Co,Clr,Eu,Pr,Lch,M1°	1	119	1.5	100	36	5041	0.6	0.13	5.288	0.007	0.3355	0.0004	0.8550	0.11430	0.00008	1869.0	2.4	0.24
A30-1 (Z)	Z,Co,Clr,Eu,St,Lch,M1°	1	113	2.0	174	63	5738	1.3	0.13	5.283	0.006	0.3352	0.0003	0.9071	0.11430	0.00006	1868.9	1.8	0.32
A30-2 (Z)	Z,Co,Clr,Eu,Lch,M1°	1	86	1.1	81	29	705	2.7	0.11	5.305	0.011	0.3362	0.0004	0.7445	0.11446	0.00016	1871.5	5.0	0.21

Deformed hornblende-biotite granodiorite (ZR) - 11lo4155a (Z10554)***452952m E, 7208216m N**

A16-1 (Z)	Z,Co,Clr,Eq,Pr,Lch,NM5°	1	100	3.4	210	77	10620	1.4	0.16	5.283	0.006	0.3353	0.0003	0.9371	0.11430	0.00005	1868.8	1.6	0.31
A16-2 (Z)	Z,Co,Clr,nFr,Mf,Pr,Lch,NM5°	1	100	1.6	154	56	2585.2	2.0	0.14	5.292	0.010	0.3360	0.0006	0.8382	0.11423	0.00012	1867.8	3.9	0.02
A16-3 (Z)	Z,Co,Clr,nFr,Mf,Pr,Lch,NM5°	1	90	1.1	154	56	2499.7	1.4	0.15	5.309	0.007	0.3367	0.0004	0.8666	0.11437	0.00008	1869.9	2.5	-0.05
A16-4 (Z)	Z,Co,Clr,nFr,St,Lch,NM5°	1	110	1.6	165	62	1930.6	2.9	0.18	5.294	0.007	0.3357	0.0003	0.8793	0.11440	0.00008	1870.4	2.4	0.29

Rhyolite crystal tuff (SBr) - 10lo1144c (Z10323)***446893m E, 7189313m N**

A20-1 (Z)	Z,Co,Clr,Eu,St,Lch,NM3°	1	162	3.6	250	94	8198	2.3	0.19	5.252	0.007	0.3339	0.0004	0.8296	0.11406	0.00008	1865.1	2.5	0.48
A20-2 (Z)	Z,Co,Clr,Eu,Pr,Lch,NM3°	1	106	1.4	285	106	5024	1.6	0.19	5.183	0.007	0.3303	0.0003	0.9293	0.11381	0.00006	1861.1	1.8	1.31
A30-1 (Z)	Z,Co,Clr,Eq,Eu,St,Lch,NM3°	1	167	6.4	62	23	8429	0.9	0.20	5.227	0.006	0.3332	0.0003	0.8970	0.11378	0.00006	1860.6	1.9	0.42
A30-2 (Z)	Z,Co,Clr,Eu,Pr,Lch,M3°	1	154	5.5	61	23	7354	1.0	0.17	5.224	0.006	0.3335	0.0003	0.9164	0.11359	0.00006	1857.7	1.8	0.14
B16-1 (Z)	Z,Co,Clr,Eu,St,Lch,NM3°	1	100	2.3	100	38	503	9.9	0.20	5.265	0.013	0.3351	0.0005	0.6368	0.11396	0.00021	1863.6	6.8	0.04
B20-1 (Z)	Z,Co,Clr,Eu,St,Lch,M3°	1	100	3.8	118	44	3787	2.5	0.19	5.229	0.007	0.3333	0.0003	0.9093	0.11377	0.00006	1860.6	2.0	0.37
B30-1 (Z)	Z,Co,Clr,Fg,Pr,Lch,M3°	1	100	1.5	114	42	907	4.0	0.17	5.264	0.010	0.3349	0.0004	0.7759	0.11398	0.00013	1863.9	4.2	0.11
B30-2 (Z)	Z,Co,Clr,Eu,Pr,Lch,M3°	1	145	1.5	111	42	5523	0.6	0.22	5.217	0.006	0.3324	0.0003	0.8781	0.11385	0.00007	1861.8	2.1	0.74
B30-3 (Z)	Z,Co,Clr,Eu,Pr,Lch,M3°	1	173	2.7	116	43	2502	2.6	0.18	5.214	0.007	0.3321	0.0004	0.9102	0.11385	0.00007	1861.8	2.2	0.82
B30-4 (Z)	Z,Co,Clr,Eu,Pr,Lch,M3°	1	124	1.2	126	47	4263	0.7	0.18	5.223	0.007	0.3327	0.0004	0.8991	0.11387	0.00007	1862.1	2.2	0.66

*sample coordinates in UTM Zone 11, NAD 83

Abbreviations in brackets refer to sample locations depicted in Figure 2 and in the text.

Description Legend: Z=Zircon, Co=Colourless, pBr=Pale Brown, Alt=Altered, Clr=Clear, fFr=Few Fractures, nFr=Numerous Fractures, cln=Clear Inclusions, fln=Few Inclusions, nln=Numerous Inclusions, Eq=Equant, Eu=Euhedral, Fg=Fragment, Pr=Prismatic, St=Stubby Prism, Sub=Subhedral, Tab=Tabular, Tip=Tip, Abr=Abraded, Lch=Leached

(1) Weights estimated Concentration uncertainty up to 50%

(2) Pbr = radiogenic Pb. Pbc = total common Pb in analysis corrected for spike and fractionation.

(3) Atomic ratios corrected for spike, fractionation, blank and initial common Pb, except 206Pb/204Pb ratio corrected for spike and fractionation only. Errors are one sigma absolute. Pb blank: 1-2 pg for zircon ; blank composition in atomic proportions = 51.966:21.356:25.288:1.3895 (208:207:206:204)

Table 2. SHRIMP II U-Pb zircon results
Rhyodacite (HL) - 09vj148c (Z9952)
*514583m E, 7189497m N

Spot	U (ppm)	Th (ppm)	Th U	206Pb* (ppm)	²⁰⁴ Pb ²⁰⁶ Pb	± ²⁰⁴ Pb ²⁰⁶ Pb	f(206) ²⁰⁴	²⁰⁸ Pb ²⁰⁶ Pb	± ²⁰⁸ Pb ²⁰⁶ Pb	²⁰⁷ Pb ²³⁵ U	± ²⁰⁷ Pb ²³⁵ U	²⁰⁶ Pb ²³⁸ U	± ²⁰⁶ Pb ²³⁸ U	Corr Coeff	²⁰⁷ Pb ²⁰⁶ Pb	± ²⁰⁷ Pb ²⁰⁶ Pb	²⁰⁷ Pb ²⁰⁶ Pb	± ²⁰⁷ Pb ²⁰⁶ Pb	Disc. (%)
9952-002.1	384	225	0.584	110	2.132E-06	2.132E-06	0.044	0.1813	0.0021	5.212	0.047	0.3319	0.0029	0.9635	0.1139	0.00028	1863	4.4	1.0
9952-030.1	252	143	0.568	73	2.640E-05	9.334E-06	0.556	0.1751	0.0025	5.333	0.051	0.3359	0.0031	0.9440	0.1151	0.00037	1882	5.7	0.9
9952-007.1	840	746	0.888	242	-1.006E-06	-1.006E-06	-0.014	0.2716	0.0031	5.272	0.045	0.3348	0.0028	0.9807	0.1142	0.00019	1868	3.0	0.4
9952-001.1	817	1024	1.252	233	3.937E-06	1.968E-06	0.037	0.3915	0.0021	5.215	0.045	0.3317	0.0028	0.9808	0.1140	0.00019	1865	3.0	1.1
9952-004.1	227	177	0.782	65	7.273E-05	1.626E-05	1.106	0.2414	0.0031	5.237	0.052	0.3320	0.0031	0.9271	0.1144	0.00043	1871	6.7	1.4
9952-017.1	315	109	0.345	91	1.415E-05	5.776E-06	0.480	0.1088	0.0017	5.311	0.049	0.3352	0.0030	0.9588	0.1149	0.00030	1878	4.7	0.9
9952-015.1	468	244	0.522	131	4.285E-05	8.569E-06	0.982	0.1603	0.0018	5.116	0.048	0.3255	0.0028	0.9228	0.1140	0.00041	1864	6.5	2.9
9952-031.1	715	681	0.952	206	2.208E-05	4.818E-06	0.274	0.2977	0.0046	5.267	0.045	0.3349	0.0028	0.9773	0.1141	0.00021	1865	3.3	0.2
9952-063.1	429	374	0.872	121	3.974E-05	8.473E-06	0.541	0.2712	0.0044	5.185	0.047	0.3280	0.0028	0.9624	0.1146	0.00028	1874	4.4	2.8
9952-056.1	699	385	0.551	199	1.184E-04	1.205E-05	2.509	0.1709	0.0028	5.217	0.045	0.3311	0.0028	0.9666	0.1143	0.00026	1868	4.0	1.5
9952-048.1	624	649	1.040	178	2.869E-05	5.983E-06	0.333	0.3184	0.0021	5.214	0.045	0.3310	0.0028	0.9736	0.1142	0.00023	1868	3.6	1.5
9952-051.1	398	142	0.357	114	2.794E-05	7.467E-06	0.965	0.1063	0.0017	5.280	0.048	0.3336	0.0029	0.9611	0.1148	0.00029	1877	4.5	1.3
9952-053.1	646	515	0.797	177	5.183E-05	9.176E-06	0.775	0.2463	0.0019	5.028	0.044	0.3194	0.0027	0.9690	0.1142	0.00025	1867	3.9	4.9
9952-054.1	838	712	0.850	233	1.195E-04	1.395E-05	1.695	0.2573	0.0020	5.072	0.053	0.3241	0.0033	0.9697	0.1135	0.00029	1856	4.6	2.9
9952-043.1	460	198	0.431	132	4.149E-05	8.298E-06	1.175	0.1294	0.0016	5.271	0.047	0.3339	0.0029	0.9643	0.1145	0.00027	1872	4.3	0.9
9952-044.1	335	213	0.637	98	2.127E-05	6.727E-06	0.401	0.1960	0.0053	5.374	0.053	0.3407	0.0030	0.8890	0.1144	0.00052	1871	8.1	-1.2
9952-040.1	374	277	0.741	107	4.156E-05	9.292E-06	0.674	0.2271	0.0023	5.253	0.048	0.3339	0.0029	0.9569	0.1141	0.00030	1866	4.8	0.5
9952-037.1	446	425	0.952	128	3.214E-05	7.374E-06	0.408	0.2910	0.0023	5.283	0.047	0.3343	0.0029	0.9654	0.1146	0.00027	1874	4.2	0.9
9952-035.1	669	682	1.020	179	1.871E-05	4.678E-06	0.219	0.3155	0.0020	4.915	0.042	0.3114	0.0026	0.9760	0.1145	0.00021	1872	3.4	7.6
9952-034.1	810	693	0.855	235	-2.590E-06	-1.495E-06	-0.037	0.2608	0.0016	5.340	0.047	0.3378	0.0028	0.9363	0.1146	0.00036	1874	5.6	-0.1

*sample coordinates in UTM Zone 11, NAD 83
Abbreviations in brackets refers to sample location depicted in Figure 2 and in the text.
Spot name follows the convention x-y.z; where x = sample number, y = grain number and z = spot number. Multiple analyses in an individual spot are labelled as x-y.z.z
Uncertainties reported at 1s (absolute) and are calculated by numerical propagation of all known sources of error
f206²⁰⁴ refers to mole fraction of total 206Pb that is due to common Pb, calculated using the 204Pb-method; common Pb composition used is the surface blank (4/6: 0.05770; 7/6: 0.89500; 8/6: 2.13840)
* refers to radiogenic Pb (corrected for common Pb)
Discordance relative to origin = 100 * (1-(206Pb/238U age)/(207Pb/206Pb age))
Calibration standard 6266; U = 910 ppm; Age = 559 Ma; 206Pb/238U = 0.09059
Error in 206Pb/238U calibration 0.8%; mount calibration error = 0.16%
Th/U calibration: F = 0.03900*UO + 0.85600

Table 3. Summary of key geochemical characteristics of Hottah and Great Bear igneous rocks

Unit	¹ Age (Ma)	La/Ybcn	La/Smcn	Dy/Ybcn	Eu/Eu*	Nb/Y	Zr/Y	εNd ^T	T _{DM} (Ga)
Sloan vesicular basalt (SBr; n=4)	1863	4.8 - 8	2.8 - 3.7	1.0 - 1.2	0.7 - 0.9	0.3 - 0.5	5.3 - 6.8	-	-
Sloan rhyolitic crystal-rich flow (SBr; n=1)	1863	17.9	5.6	1.2	0.2	0.5	4.8	-0.5	2.24
Faber group rhyolitic ignimbrite (SL; n=1)	1869	14.4	5.4	1.2	0.5	0.4	7.8	-1.0	2.31
LaBine Group andesite (BLa; n=1)	1873	20.6	7.2	1.1	0.7	0.5	9.1	-0.6	2.25
Grant (RL) and Rebesca (GL) basalt (n=5)	1893	1.8 - 2.7	1.4 - 2.0	1.1 - 1.4	0.7 - 1.0	0.1	2.9 - 3.2	+1.7, +1.7	-
¹ Bloom basalt (n=2)	1895	2.3	1.7	1.1 - 1.2	1.0, 1.5	0.1	3.1	+2.2	-
Grant rhyodacite (RL; n=2)	1893	3.5	2.5	1.0 - 1.1	0.6	0.2	3.8	+0.5	2.41
¹ Bloom rhyodacite (Br; n=1)	1895	5.3	3.6	1.6	0.4	0.3	3.0	-1.9	2.50
¹ Zebulon Formation rhyolite (n=6)	1906	1 - 6.7	1.4 - 3.6, 13.3-15.7	0.4 - 1.0	0.1 - 0.3	0.4 - 0.7, 1.1	3.1 - 8.4, 11.1	+0.8	2.24
¹ Zebulon Formation basalt (n=4)	1906	4.7 - 5.1, 13.2	2.2 - 2.4, 4.2	1.1 - 1.5	0.6 - 0.9	0.2	3.8 - 5.3	+2.1	-
Hottah tonalite (Ht; n=1)	1913	5.3	2.7	1.1	1.0	0.3	3.5	+2.0	2.21
Hottah deformed granodiorite (Ht; n=1)	1928	-	-	-	-	-	-	-1.6	2.44
Hottah quartz-plagioclase porphyry (Hp; n=2)	1931	9.3	4.2	1.0 - 1.2	0.45	0.40	7.4, 8.3	-1.5	2.44
Hottah K-feldspar phenocrystic granite (Hg; n=2)	1931	10 - 12.7	4.8 - 5.6	1.1	0.2 - 0.6	0.4 - 0.6	3.8 - 8.2	-	-

¹ - does not include age or geochemical data from Reichenbach (1991)

Abbreviations in brackets (Ht) correspond to sample locations depicted in Figure 2 and in the text. Without abbreviations indicates that samples were derived from a variety of locations in the map units on Figure 2, including geochronological sample location (where applicable).

n = number of samples

cn = chondrite normalized

Full data results are presented in Tables A1 and A2; Analytical techniques are in the Appendix

Table A1. Geochemical results

Description				Sloan rhyolite	Sloan basalt	Sloan basalt	Sloan basalt	Sloan basalt	Swan Lake rhyolite	Beaverlodge andesite	Bloom basalt
Sample				10LO1144C	10LO1144F	10LO1158A	10LO1158C	10LO1158D	10LO1131B	10VJ186A	11LO4150A
Age (Ma)				1863	1863	1863	1863	1863	1869	1873	1895
*UTM X				446893	446864	446779	446674	446661	462754	447400	434784
*UTM Y				7189313	7189461	7190986	7190991	7191017	7201067	7177920	7215381
Analyte	Method	Unit	MDL								
SiO ₂	XRF	%	0.1	74.4	56.8	57.3	59.7	57.4	70.5	60.2	47.9
TiO ₂	XRF	%	0.01	0.19	1.04	0.7	1.07	1.13	0.25	0.66	0.95
Al ₂ O ₃	XRF	%	0.01	12.32	16.38	15.52	13.22	13.69	14.65	18.13	14.94
Fe ₂ O ₃	XRF	%	0.01	2.51	9.41	9.45	11.02	11.73	2.86	5.52	11.75
MnO	XRF	%	0.01	0.03	0.19	0.12	0.24	0.22	0.07	0.1	0.21
MgO	XRF	%	0.01	0.1	3.87	3.83	4.72	3.95	0.51	4.27	7.83
CaO	XRF	%	0.01	0.97	5.16	6.69	4.1	6.61	1.7	1.93	10.48
Na ₂ O	XRF	%	0.01	2.98	2.78	3.04	2.34	2.86	2.49	1.63	2.09
K ₂ O	XRF	%	0.01	5.6	2.72	1.27	1.55	1.76	5.87	5.1	1.21
P ₂ O ₅	XRF	%	0.01	0.03	0.15	0.08	0.13	0.13	0.05	0.16	0.1
Cr ₂ O ₃	XRF	%	0.001	0.003	0.013	0.005	0.006	0.005	0.005	0.009	0.032
LOI	XRF	%		0.91	1.72	1.99	2.86	1.29	1.68	3.14	2.44
SUM		%		100.04	100.23	100	100.96	100.78	100.73	100.92	99.88
Be	¹ ICP-MS	PPM	1	114	3	4	6	<1	4	<1	1
V	¹ ICP-MS	PPM	8	<8	123	142	165	167	<8	75	237
Co	¹ ICP-MS	PPM	0.2	0.9	28.2	28.6	29.2	31.3	2.2	14.3	41.8
Ni	² ICP-MS	PPM	0.1	16.3	2.3	0.7	2.2	2.3	2.4	18.3	54
Cu	² ICP-MS	PPM	0.1	0.8	1.1	0.7	16.8	1.4	1.4	0.5	94.8
Zn	² ICP-MS	PPM	1	70	6	4	188	3	5	18	49
Ga	ICP-ES	PPM	0.5	18.1	21.7	17	16.7	17.7	16.9	22.2	14.3
As	² ICP-MS	PPM	0.5	3.5	0.6	<0.5	329	1.1	0.6	1.3	<0.5
Se	² ICP-MS	PPM	0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
Rb	¹ ICP-MS	PPM	0.1	345	111.5	49.9	58.1	59	221.9	275.7	33.2
Sr	¹ ICP-MS	PPM	0.5	33.7	215.5	254.4	96.9	141.5	174.8	230.9	153.3
Y	¹ ICP-MS	PPM	0.1	49.7	21.9	15.8	21.9	27.6	31.3	21	17.1
Zr	¹ ICP-MS	PPM	0.1	236.8	149.4	85.1	136.7	146.5	244.5	190.6	51
Nb	¹ ICP-MS	PPM	0.1	22.8	10.2	5.4	8.8	7.6	14	10.4	1.9
Mo	² ICP-MS	PPM	0.1	<0.1	0.4	0.3	1.3	0.5	0.3	0.2	0.1
Cd	² ICP-MS	PPM	0.1	<0.1	<0.1	<0.1	0.3	<0.1	<0.1	<0.1	<0.1
Ag	² ICP-MS	PPM	0.1	<0.1	<0.1	<0.1	0.2	<0.1	<0.1	<0.1	<0.1
Sn	¹ ICP-MS	PPM	1	11	3	2	1	1	5	4	<1
Sb	² ICP-MS	PPM	0.1	0.2	0.1	0.1	0.9	0.4	0.1	0.9	<0.1
Cs	¹ ICP-MS	PPM	0.1	12.8	5.4	1.5	1	0.7	5.8	5.9	0.6
Ba	XRF	%	0.01	0.04	0.05	0.02	0.04	0.04	0.11	0.11	0.01
Ba	¹ ICP-MS	PPM	1	251	483	229	322	302	995	1059	139
La	¹ ICP-MS	PPM	0.1	127.5	23.4	13.8	26.9	19.8	63.1	63.2	6
Ce	¹ ICP-MS	PPM	0.1	254.9	44.7	26.4	57.2	44	119.6	116.6	12.8
Pr	¹ ICP-MS	PPM	0.02	27.06	4.91	2.95	6.45	5.14	12.17	11.34	1.85
Nd	¹ ICP-MS	PPM	0.3	88	19	11.4	22	19.7	42.1	36.2	7.6
Sm	¹ ICP-MS	PPM	0.05	14.57	4.12	2.55	5	4.51	7.5	5.66	2.34
Eu	¹ ICP-MS	PPM	0.02	0.65	1.09	0.82	1.07	1.06	1.11	1.11	0.79
Gd	¹ ICP-MS	PPM	0.05	11.69	4.14	2.86	4.54	4.88	6.15	4.5	2.67
Tb	¹ ICP-MS	PPM	0.01	1.77	0.7	0.48	0.73	0.83	0.99	0.77	0.49
Dy	¹ ICP-MS	PPM	0.05	9.18	3.76	2.66	3.65	4.88	5.43	3.73	3.03
Ho	¹ ICP-MS	PPM	0.02	1.85	0.87	0.65	0.78	1.08	1.05	0.85	0.72
Er	¹ ICP-MS	PPM	0.03	5.27	2.5	1.8	2.29	3.26	3.22	2.54	2.11
Tm	¹ ICP-MS	PPM	0.01	0.73	0.33	0.23	0.34	0.44	0.51	0.34	0.29
Yb	¹ ICP-MS	PPM	0.05	5.1	2.1	1.83	2.17	2.94	3.15	2.2	1.87
Lu	¹ ICP-MS	PPM	0.01	0.79	0.33	0.24	0.4	0.47	0.55	0.38	0.31
Hf	¹ ICP-MS	PPM	0.1	7.5	4	2.5	3.8	4.1	7.3	5.3	1.5
Ta	¹ ICP-MS	PPM	0.1	1.6	1	0.6	0.6	0.6	1.3	1.1	0.2
W	¹ ICP-MS	PPM	0.5	3.6	<0.5	1	1.3	1.6	1.1	2.3	<0.5
Au	² ICP-MS	PPB	0.5	1	1.3	1.2	3.9	1.3	<0.5	<0.5	<0.5
Hg	² ICP-MS	PPM	0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
Tl	² ICP-MS	PPM	0.1	0.1	<0.1	<0.1	0.5	<0.1	<0.1	<0.1	<0.1
Pb	² ICP-MS	PPM	0.1	3.9	1.2	1.3	157.5	1.9	0.7	0.9	2.1
Bi	² ICP-MS	PPM	0.1	0.2	<0.1	<0.1	<0.1	<0.1	0.1	0.3	<0.1
Th	¹ ICP-MS	PPM	0.2	37	7.5	3.9	7.9	8.4	28.1	30	0.5
U	¹ ICP-MS	PPM	0.1	8	3.5	2	2	2.3	8.8	3.4	0.1

Notes:

*UTM Zone 11, NAD 83

MDL-Minimum detection limit

1 - Fusion ICP-MS

2- Aqua Regia dilution ICP-MS

Bloom basalt	Bloom rhyodacite	Rebesca basalt	Rebesca basalt	Rebesca basalt	Grant basalt	Grant basalt	Grant rhyodacite	Grant rhyodacite	Zebulon rhyolite
11LO4152A	11LO4153A	11MH8014A	11MH8015A	11LO4034A	11VJ9213A	11VJ9213B	HY.F 361.80	HY.F 362.80	10LO1114
1895	1895	1895	1895	1895	1895	1895	1895	1895	1905
436767	435077	522522	522028	522021	519038	519028	520440	520718	438720
7222049	7222942	7161224	7160913	7160853	7206068	7206070	7207824	7208224	7175363
46.8	78.8	49.5	49.7	49.2	49.6	49.4	69.5	77.1	75
1.06	0.24	1.45	1.42	1.48	1.77	1.78	0.24	0.24	0.1
16.02	10.23	13.21	13.54	13.36	12.8	12.89	10.03	10.71	11.82
12.07	2.68	14.77	14.36	15.69	14.45	14.77	2.62	3.12	4
0.2	0.03	0.22	0.24	0.21	0.22	0.23	0.12	0.04	0.01
6.22	0.89	6.48	4.87	6.26	6.06	6.14	0.87	1.4	0.66
12.33	0.36	9.99	11.75	11.24	9.32	9.89	6.58	0.98	0.01
1.6	0.71	2.17	3.2	2.29	2.59	3.27	2.64	2.97	0.07
1.56	5.01	0.94	0.53	0.27	0.46	0.25	2.35	2.54	8.06
0.11	0.15	0.12	0.12	0.12	0.16	0.16	<0.01	0.01	0.01
0.021	0.005	0.016	0.026	0.028	0.022	0.023	0.008	0.002	0.005
1.61	0.94	0.78	0.28	-0.04	1.76	2.03	5.88	1.2	0.99
99.58	100.13	99.61	100.05	100.05	99.24	100.77	100.9	100.38	100.76
<1	2	<1	2	<1	2	1	4	3	3
267	16	319	304	324	366	330	<8	<8	<8
42.5	2.8	47.8	46.2	48.8	45.5	41.3	11.6	23.1	2.3
24.2	5.5	22.5	20.2	27.2	21.9	26.3	0.3	1	1.6
88.6	14	48	25.7	83	44.6	71.4	3	1.3	1.2
17	27	41	28	19	64	69	116	173	14
16.3	11.9	17.1	16.4	17	16.9	16.4	22.7	21.7	18
<0.5	3.1	<0.5	0.7	1.1	<0.5	<0.5	<0.5	<0.5	<0.5
<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
39.5	150.9	27.1	12.1	7.8	12.4	5.1	137.8	77.1	170.5
221.7	31.7	88.4	134.4	115.7	132	119.3	164.1	151.5	12
19.8	34	28.3	30	30.7	39.5	35.6	125.6	84.9	28.1
61.3	103.3	81.3	89.5	87.9	124.6	109.2	356.8	398.6	237.2
1.9	11.6	3.5	3.7	3.2	5.7	4.5	22.4	20.7	18.9
0.2	0.7	0.2	0.2	0.2	0.3	0.2	0.4	1.6	0.7
<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	0.2	0.4	<0.1
<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1
<1	4	1	<1	<1	<1	<1	6	7	2
0.2	0.3	<0.1	<0.1	<0.1	<0.1	<0.1	0.2	<0.1	0.2
1.6	5.9	4.5	0.4	0.9	0.4	0.2	1.6	0.7	1.9
0.03	0.09	<0.01	0.01	<0.01	0.01	<0.01	0.05	0.05	0.03
321	836	97	125	49	117	64	433	491	293
6.8	42.3	10.5	8.6	8	12	12.1	56.3	51.4	35.7
16	84.2	21.2	20.3	18.5	29.6	25.8	119.4	108.3	88.4
2.31	9.92	2.76	2.84	2.82	3.72	3.58	15.15	13.83	9.56
10.8	34.8	13.5	14.1	12.2	15.8	16.1	62.6	56.5	34.9
2.78	7.54	3.46	3.51	3.69	5.04	4.53	14.95	13.09	6.36
1.45	0.95	1.12	1.23	1.23	1.36	1.58	2.82	3.11	0.14
3.1	6.9	4.11	4.6	4.78	6.16	5.31	16.63	13.66	4.69
0.58	1.11	0.78	0.82	0.85	1.08	0.96	3.2	2.53	0.88
4.03	6.51	5.21	5.44	5.37	6.75	5.91	19.74	15.98	5.49
0.79	1.1	1.05	1.07	1.16	1.03	1.3	4.4	3.36	1.13
2.6	3.02	3.21	3.44	3.61	4.73	3.92	13.52	9.86	3.27
0.34	0.5	0.49	0.5	0.52	0.66	0.58	2.03	1.47	0.57
2.21	2.67	2.95	3.13	3.18	3.19	3.65	13.67	9.32	3.81
0.3	0.37	0.48	0.5	0.49	0.59	0.55	2.06	1.43	0.64
1.6	3.9	2.7	3	2.7	3	3.4	10.3	11.7	7.3
0.1	1	0.3	0.3	0.3	0.6	0.4	1.7	1.9	1.5
<0.5	3.2	<0.5	0.9	<0.5	0.8	<0.5	101.6	202.7	0.8
<0.5	<0.5	1.3	0.9	<0.5	<0.5	<0.5	3.8	2.5	<0.5
<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
0.1	0.4	0.2	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1
2.8	7.9	1.5	0.9	0.8	2.5	1.7	15.2	26.6	2.4
<0.1	0.5	0.1	<0.1	<0.1	<0.1	<0.1	0.2	0.2	0.1
1	10.8	1.8	1.4	1.2	3.2	2.8	16.7	16.4	14.4
0.1	4.1	0.5	0.5	0.4	0.7	0.8	4.6	4.2	3.6

Zebulon rhyolite	Zebulon rhyolite	Zebulon rhyolite	Zebulon rhyolite	Zebulon rhyolite	Zebulon basalt	Zebulon basalt	Zebulon basalt	Zebulon basalt	Hottah tonalite	Hottah granodiorite
10LO1116A	10LO1116B	10PA33A	10VJ89A	11VJ9221D	09LO9040	10VJ89B	11VJ9221A	11VJ9221B	11LO4157A	11LO4156A
1905	1905	1905	1905	1905	1905	1905	1905	1905	1913	1930
437934	437911	437191	439110	431138	443234	439327	431094	431055	435738	452888
7174560	7174557	7208164	7208252	7205432	7178978	7208329	7205519	7205589	7206954	7208252
75.5	74.4	77.7	84	74.3	49.2	62.4	53.1	48.6	56.8	68.7
0.13	0.11	0.03	0.08	0.1	1.07	0.53	1.36	1.61	0.67	0.74
11.68	12.69	10.37	7.91	12.39	18.8	15.24	18.21	14.64	16.99	13.53
3.4	3.46	2.18	2.72	2.24	16.01	10.81	12.38	13.87	7.89	5.75
<0.01	0.02	<0.01	<0.01	<0.01	0.03	0.02	0.03	0.37	0.15	0.09
0.53	0.72	0.03	0.36	0.03	3.85	0.83	0.7	5.35	4.17	2.04
0.01	0.03	0.02	0.01	0.02	0.38	0.13	0.68	8.08	5.43	1.17
0.06	0.06	0.12	0.09	0.1	0.14	0.23	2.14	2.38	4.05	3.09
7.7	7.68	8.9	4.05	11.13	6.58	9.01	9.67	1.09	2.07	3.59
<0.01	<0.01	<0.01	<0.01	<0.01	0.22	0.1	0.45	0.32	0.09	0.16
0.002	0.005	<0.001	<0.001	0.003	0.029	0.035	<0.001	0.03	0.005	0.003
1.07	1.34	0.22	0.93	0.19	3.8	1.5	1.52	3.33	1.68	1.37
100.17	100.57	99.7	100.27	100.54	100.15	100.89	100.28	99.68	100.04	100.38
5	2	3	5	<1	5	2	3	1	1	2
<8	<8	<8	<8	<8	114	68	85	240	117	30
1.1	2.8	0.6	0.4	0.9	22.1	4.1	7.1	42	17.7	6.8
1.7	4.8	43.4	25.7	1.4	80.8	6	1.5	67	19.3	6.2
5.8	5.5	20.5	9	1.2	0.2	11.1	0.5	13.7	9.8	3.9
42	55	80	35	2	20	108	7	97	56	61
13.4	19.2	13.4	18.4	7.7	22	21	20.3	18.8	17.2	19.4
2.5	0.7	1.2	1.3	1.4	1	1	3.6	1.8	<0.5	0.7
0.6	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
173.4	199.4	211.5	215.1	230.5	256	357.5	340.8	32.5	47.8	118.7
6.3	8.7	15.8	15.4	15.4	7.5	19	28.8	376.8	332.3	87.8
19	20.7	33.3	58.2	31.8	23.6	23.1	35.8	32	19.7	40.6
211.1	232.5	157.4	179.9	204.5	88.7	109	190.1	154.3	69.8	337.7
22.3	23.7	23.9	23.4	18.9	4.9	5.3	8.4	6.1	5.1	19.1
0.9	1.9	0.3	0.5	0.3	<0.1	0.3	0.6	0.2	<0.1	1
<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1
<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1
4	4	3	3	2	1	2	2	1	1	6
0.3	0.3	0.3	0.8	0.3	0.1	0.3	1.4	<0.1	<0.1	<0.1
1.9	2.8	0.8	1	1.2	5.8	5.9	6.3	0.5	0.8	1.1
0.02	0.03	0.09	0.04	0.03	0.07	0.1	0.04	0.04	0.05	0.09
223	249	908	840	327	694	875	384	288	485	785
39.4	24.8	6.9	20.9	6.6	18.4	42	24.5	21.3	15.9	61.4
76.1	44	18.3	53	14.9	41.4	72.8	61.5	49.3	39.1	125.7
7.52	3.9	2.43	7.07	2.17	5.14	9.39	8.05	6.8	4.86	14.36
21.2	11	9.5	26.4	8.6	20.9	32.5	34.5	30.1	18.3	51.3
1.91	1.06	3.11	7.88	2.4	4.94	6.51	7.14	6.28	3.8	9.15
0.09	0.08	0.11	0.34	0.27	1.45	1.17	1.67	1.77	1.17	1.4
1.3	1.29	3.68	8.33	3.07	4.77	5.32	7.38	6.08	3.72	8.82
0.35	0.33	0.85	1.61	0.71	0.83	0.74	1.2	0.96	0.56	1.34
2.75	3.27	4.98	9.73	5.45	4.68	3.69	7.28	6.01	3.42	8.09
0.76	0.9	1.35	2.13	1.19	1	0.76	1.38	1.21	0.69	1.51
3.01	3.41	3.87	6.26	3.89	2.6	2.32	3.77	3.47	2.18	4.09
0.53	0.55	0.61	0.93	0.65	0.36	0.32	0.55	0.52	0.33	0.66
4.31	4.47	4.89	6.4	4.63	2.59	2.28	3.28	3.25	2.15	3.81
0.68	0.74	0.78	0.96	0.69	0.41	0.36	0.43	0.46	0.31	0.55
7.5	8.9	7.1	8.1	7.8	2.6	3.1	4.8	4.4	2.2	10.2
1.7	1.8	2	2	1.4	0.4	0.5	0.6	0.4	0.4	1.5
1.3	0.9	4	3.8	0.8	<0.5	3.5	0.8	<0.5	<0.5	1.9
<0.5	0.9	0.8	<0.5	1.5	<0.5	<0.5	0.7	<0.5	<0.5	<0.5
<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
0.1	<0.1	0.5	<0.1	<0.1	0.2	<0.1	<0.1	<0.1	<0.1	0.2
12.7	2.6	2.2	1.7	1	2.1	5	3.4	1.3	8.1	8.9
0.3	<0.1	<0.1	<0.1	<0.1	0.3	0.2	0.3	<0.1	<0.1	0.1
14.9	15.6	17.3	19.3	13	2.9	4.6	2.5	2	1.8	26.6
4.8	4.6	3.2	4.6	3.3	2.2	2.6	1.6	0.8	0.5	6.6

Hottah porphyry	Hottah porphyry	Hottah granite	Hottah granite
06LO-BLL2	09LO9039A	09LO9078A	09LO9026
1930	1930	1930	1930
443065	443456	442095	440933
7178527	7179171	7177812	7177258
68.9	70.1	71.9	77.2
0.62	0.59	0.44	0.22
13.74	13.22	12.96	11.43
5.26	4.99	4.13	2.43
0.06	0.04	0.03	0.01
1.01	1.75	1.31	0.74
1.39	0.33	0.2	0.11
3.09	2.7	2.25	0.09
4.4	4.51	5.91	6.56
0.15	0.13	0.09	0.08
0.003	0.003	0.006	0.007
1.01	1.63	1.24	1.36
99.75	100.1	100.58	100.25
3	2	5	2
36	23	32	10
6.5	5.5	4.2	1.1
4.8	4.3	5	5
12.1	3.4	2	2.6
50	56	33	17
20.9	17.8	18.6	16.5
0.8	<0.5	0.9	0.6
<0.5	<0.5	<0.5	<0.5
131.2	125.7	204.7	294.9
107.1	46.8	38.7	20.7
50.2	42.5	37.4	43.3
373.3	351.8	269.5	163.8
21.1	18.3	20.2	17.1
1.9	0.2	0.4	0.4
<0.1	<0.1	<0.1	0.1
<0.1	<0.1	<0.1	<0.1
6	7	8	10
0.2	0.8	<0.1	<0.1
2.4	1.3	2.4	18
0.1	0.09	0.08	0.04
973	770	717	371
64.5	64.3	54.2	62.3
132.3	126.9	111.3	126.3
15.31	13.67	12.4	13.3
57.3	47.4	41	43
10.25	9.58	7.25	8.26
1.57	1.28	0.9	0.57
9.62	8.48	6.39	7.35
1.49	1.41	1.07	1.21
8.73	7.71	6.47	6.83
1.81	1.61	1.29	1.5
5.25	4.56	4.03	4.48
0.79	0.66	0.58	0.66
4.91	5.01	3.89	4.29
0.72	0.69	0.58	0.67
11	10.6	8.6	5.8
1.7	1.7	1.7	2
2	1.2	2.1	1.8
4.3	1.5	2.7	<0.5
<0.01	<0.01	<0.01	<0.01
0.1	<0.1	<0.1	<0.1
7.5	4.9	4.9	7.6
0.6	0.2	0.2	0.1
27.5	27.9	37.2	47.9
7.7	8.2	8.1	11

Table A2: Sm-Nd isotope data

Description	Sample Name	Age	Nd (ppm)	Sm (ppm)	¹⁴³ Nd/ ¹⁴⁴ Nd (current)	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd (initial)	+/- (2 σ)	εNd _(CHUR) ^T	T _{DM} (0.214, 0.513115)
Stairs Bay rhyolite (SBr)	10lo1144c	1862	98.86	16.00	0.511403	0.0978	0.510205	0.000016	-0.5	2236
Stairs Bay basalt (SBr)	10lo1144f	1862	21.16	4.55	0.511840	0.1299	0.510249	0.000019	0.4	-
Great Bear hornblende granodiorite (ZR)	11lo4155a	1869	47.55	8.96	0.511480	0.1139	0.510079	0.000034	-2.7	2478
Swan Lake ignimbrite (SL)	10lo1131b	1869	43.66	7.67	0.511474	0.1062	0.510168	0.000016	-1.0	2310
Beaverlodge andesite (BLa)	10vj186	1873	29.71	4.71	0.511366	0.0959	0.510184	0.000015	-0.6	2248
Bloom rhyodacite (Br)	11lo4153a	1895	39.14	8.13	0.511656	0.1255	0.510091	0.000038	-1.8	2500
Bloom basalt	11lo4150a	1895	9.53	2.61	0.512362	0.1654	0.510300	0.000023	2.3	-
Grant basalt (GL)	11vj9213a	1893	17.50	4.95	0.512406	0.1711	0.510275	0.000024	1.7	-
Grant rhyodacite (GL)	362.8	1893	59.08	14.24	0.512028	0.1457	0.510213	0.000021	0.5	2413
Rebesca basalt (RL)	11lo4034a	1893	12.79	3.82	0.512524	0.1804	0.510276	0.000025	1.7	-
Zebulon rhyolite (Zr)	10lo1114	1906	36.50	6.77	0.511617	0.1121	0.510211	0.000017	0.8	2232
Zebulon basalt	09lo9040	1906	23.68	5.26	0.511962	0.1342	0.510278	0.000020	2.1	-
Hottah tonalite (Ht)	11lo4157a	1910	18.94	4.07	0.511891	0.1298	0.510259	0.000039	1.8	2208
Hottah granodiorite (Hg)	11lo4156a	1930	54.05	10.17	0.511503	0.1137	0.510059	0.000034	-1.6	2438
Hottah porphyry (Hp)	09lo9039a	1930	53.34	10.10	0.511517	0.1145	0.510063	0.000018	-1.5	2436

Notes: T_{DM} calculated using Goldstein et al. (1984) for samples with ¹⁴⁷Sm/¹⁴⁴Nd <0.1500

Abbreviations in brackets (Ht) correspond to sample locations depicted in Figure 2 and in the text. Without abbreviations indicates that samples were derived from a variety of locations in the map units on Figure 2, including geochronological sample location (where applicable).