Neoarchean arc magmatism and subsequent collisional orogenesis along the eastern Rae domain, western Churchill Province: implications for the early growth of Laurentia

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Abstract: The Athabasca granulite terrane, on the eastern margin of the Rae domain of the western Churchill Province, contains a well exposed record of Neoarchean magmatism, deformation, and metamorphism. Circa 2.63-2.60 Ga plutonic rocks range from 58.01 to 73.71 SiO₂, and vary from ferroan to magnesian. They form shallow linear trends on modified alkali-lime index, plotting largely in the calc-alkaline field. On tectonic discrimination diagrams, all samples plot in the volcanic arc field. Samples are LREE enriched, and contain relatively steep negative slopes, and flat, depleted, HREE profiles when normalized to chondrites. The entire Neoarchean plutonic complex, independent of lithology, contains strong enrichments in the LILEs and Pb, and depletions in HFSEs. Monazite U-Th-total Pb results from a migmatitic paragneiss suggest that partial melting occurred at ca. 2.57 Ga within the Chipman subdomain. Monazite results from the adjacent northwestern subdomain suggest that deformation occurred at ca. 2.57 Ga. IN-SIMS and ICP-MS zircon U-Pb geochronology corroborates these conclusions. Although the region is dominated by Paleoproterozoic localized shear zones, internal to each subdomain is evidence of a widespread tectonothermal event during the Neoarchean that involved arc plutonism, crustal thickening, lower crustal flow at pressures exceeding 1.0 GPa, and extensive partial melting of the lower continental crust. We interpret this sequence of magmatic, deformational, and anatectic processes to reflect the transition from a continental arc to a collisional tectonic setting. The recognized series of events presented herein is likely correlative to the 2.60 – 2.50 Ga MacQuoid Orogen, which has been recognized further to the northeast, and likely parallels the eastern margin of the Rae domain of the western Churchill Province. These data support an important component of modern style plate tectonics during continental magmatism and accretion in the Neoarchean.

INTRODUCTION

The Canadian Shield is one of the Earth's major cratons with crustal material dating back to as old as 3.9 Ga (Bowring et al., 1989). It consists of a number of Archean provinces separated by Paleoproterozoic orogenic belts (Fig. 1; Hoffman, 1988). The region offers a superb opportunity to investigate the growth and evolution of the North American continent, and to evaluate the degree with which modern tectonic processes were active during the growth and ultimate stabilization of the Archean lithosphere. Many of the stitching Paleoproterozoic orogenic belts (Hoffman, 1988), perhaps none quite as well characterized as the Trans-Hudson Orogen (Chiarenzelli et al., 1998; Andsell, 2005; Corrigan et

al., 2005; 2009; St-Onge et al., 2006; Maxeiner and Rayner., 2010), are routinely and justifiably interpreted in terms of modern plate-tectonic processes given the similar size, geometry, and composition to modern orogens (Corrigan et al., 2009). The nature of Archean boundaries in the Canadian Shield, and of Archean tectonic processes in general (Shirey et al., 2008; Van Kranendonk, 2010) Hynes, 2014), is much less well known.

The timing and character of Archean tectonics have been particularly unclear in the Churchill Province of the Canadian Shield, partly because of the vast size and remoteness of the region and partly because of the complex overprinting and interleaving of Proterozoic and Archean components (Mahan et al., 2006a; Williams et al., 2009; Dumond et al., 2010; Bethune et al., 2013). The Churchill Province has typically been subdivided into two large subprovinces, Rae and Hearne domains, separated by the geophysically defined Snowbird Tectonic Zone (STZ; Hoffman, 1988). Debate persists regarding the history of this enigmatic boundary, and has been interpreted as: 1) a Paleoproterozoic suture (Hoffman, 1988; Berman et al., 2007), 2) an intracontinental strike-slip shear zone (Hanmer et al., 1994), 3) a Neoproterozoic back arc margin (Aspler and Chiarenzelli, 1996), and 4) an anastomosing tapestry of reverse and strike slip shear zones (Mahan and Williams, 2005). Consequently, the relationship between the Rae and Hearne domains remains uncertain (Aspler et al., 1999; Berman et al., 2007).

The Athabasca granulite terrane is a large region ($>20,000 \text{ km}^2$) of high pressure granulite exposed along the eastern margin of the Rae domain. It preserves a well exposed transect across the Snowbird Tectonic zone and into the Hearne domain (Williams and Hanmer, 2006; Williams et al., 2009; Dumond et al., 2013). One important component of the Athabasca granulite terrane is a large suite of ca. 2.6 Ga meta-plutonic rocks that vary in composition from mafic to felsic. The location of this suite, along the present-day boundary between the Rae and Hearne domains, suggest that it may help to illuminate the Archean history of the boundary. This contribution presents new structural, petrologic, geochemical, and geochronologic data in order to constrain the tectonic setting and evolution of the 2.6 Ga igneous suite within the Athabasca granulite terrane. The data suggest that the rocks formed part of a large Archean arc

batholith and that soon after assembly, the margin underwent significant crustal thickening, high-grade metamorphism, and high-T lateral flow. These data are all consistent with continental collision along the boundary of the Rae domain and suggest that although the Snowbird tectonic zone is currently dominated by Proterozoic structural elements, parts of the zone may have originated as an Archean orogen.

REGIONAL BACKGROUND

Western Churchill Province

The 2800 km long, geophysically defined, Snowbird Tectonic Zone (STZ) was interpreted as the boundary between the Rae and Hearne domains of the western Churchill Province (Gilboy, 1980; Hoffman, 1988; Ross, 1994). The lineament is defined by a positive anomaly in the horizontal gravity gradient map and also consists of an anastomosing tapestry of aeromagnetic lineaments (Sharpton et al., 1987; Hoffman, 1988). This dramatic geophysical feature has been recognized from the subsurface of Alberta to northwestern Hudson Bay (Hoffman, 1988; Ross et al., 1994). The origin and geological significance of the lineament has been, and continues to be, enigmatic with some workers suggesting that portions of the STZ may not correspond to a Rae-Hearne boundary (Berman et al., 2007). The focus herein will be on the eastern Rae domain, including rocks and structures that correspond to the STZ.

The Rae domain, at least near the STZ, is characterized by high grade metamorphic assemblages (amphibolite and granulite facies), which contain evidence for multiple phases of deformation in both Neoarchean basement rocks, and infolded Paleoproterozoic cover sequences (Martel et al., 2008; Ashton et al., 2013; Berman et al., 2013; Bethune et al., 2013). Basement rocks include 2.67-2.30 Ga granitoids, which are particularly common (Hanmer et al., 1994; Martel et al., 2008; Hartlaub et al., 2007; Sanborn-Barrie et al., 2014a). There are rare and isolated regions of >3.0 Ga basement (Hanmer et al., 1994; Henderson and Theriault, 1994; Martel et al., 2008; Sanborn-Barrie et al., 2014a), and dominant Neoarchean supracrustal belts that contain distinctive komatiite successions, some associated with fuchsite-bearing quartzites (Machattie, 2008; Corrigan et al., 2013; Pehrsson et al., 2013; Sanborn-Barrie

et al., 2014b). Cover sequences include the Paleoproterozoic Murmac Bay and Amer-Chantry groups, and large tracts of undifferentiated, infolded, paragneiss (Aspler and Chiarenzelli, 1998; Aspler et al., 2001; Martel et al., 2008; Rainbird et al., 2010; Ashton et al., 2009;2013).

The Rae domain contains abundant evidence for reworking during the 1980-1900 Ma Taltson-Thelon Orogen, focused along the western margin of the domain (Hanmer et al., 1992; Bethune et al., 2013; Ashton et al., 2009; 2013; Hildebrand et al., 2010; Card et al., 2014). Structural evidence of this event decreases from west to east, towards the central STZ. In contrast, the metamorphic grade of cover sequence rocks increases from west to east. This metamorphism is generally interpreted to postdate the Taltson-Thelon tectonism, and is generally attributed to the 1.9 Ga Snowbird event (Berman et al., 2007; Ashton et al., 2009; Bethune et al., 2013). In the easternmost portions of the Rae domain, closest to the STZ, metamorphic grades in the cover rocks are similar to those in the Archean basement (granulite), making it difficult to distinguish Archean and Proterozoic rocks and events. This region has been designated as the Athabasca granulite terrane, which outlines a region that records high P-T metamorphism at ca. 1.9 Ga (Dumond et al., 2008; Williams et al., 2009)

The Hearne domain, southeast of the STZ, is predominately characterized by Paleoproterozoic sedimentary sequences (Hurwitz Group), Neoarchean basement exposures within antiformal windows (Aspler et al., 2002a), and one of the largest autochthonous greenstone belts in North America, the 2.7 Ga Ennadai -Rankin- greenstone belt (Aspler and Chiarenzelli, 1996). Metamorphic grade is generally in the greenschist or amphibolite facies (Hanmer et al., 2004; Davis et al., 2004; Mahan et al., 2003; Cousens et al., 2004). Archean basement of the Hearne domain differs from that of the Rae domain, in that it contains a greater amount of Archean supracrustal rocks, and is seemingly more juvenile (Aspler and Chiarenzelli, 1996). Aspler and Chiarenzelli (1996) presented detailed stratigraphy and petrology within the Henik segment of the central Hearne domain and inferred a Neoarchean history for the Ennadai-Rankin greenstone belt. They suggested that its development was far removed from a continental influence (at 2.70 Ga). Geochemical, isotopic, and geochronological analyses suggest that bimodal

volcanism was caused by partial melting of a depleted mantle source (Cousens et al., 2004), and rhyolitic flows were derived from melting of older basaltic crust (Cousens et al., 2004; Davis et al., 2004). These authors interpreted an infant arc setting for the central Hearne domain during the Archean, akin to the modern day western Pacific Ocean (Cousens et al., 2004; Davis et al., 2004; Hanmer et al., 2004).

The Athabasca Granulite Terrane

The Athabasca granulite terrane exposes >20,000 km² of granulite facies rocks, and has been interpreted as an isobarically-cooled terrane that is representative of lower continental crust (Mahan and Williams, 2005; Williams and Hanmer, 2006; Williams et al., 2009). The eastern margin of the Rae domain and the STZ are particularly well exposed in the eastern Athabasca granulite terrane (Dumond et al., 2008; Mahan et al., 2008). The East Athabasca mylonite triangle (Tantato domain of Gilboy, 1980) is a high grade region of nested shear zones exposed along the easternmost margin of the Rae domain and Athabasca granulite terrane is particularly well characterized, and will be the focus of this contribution (Hanmer et al., 1994; Hanmer et al., 1995; Mahan and Williams, 2005; Williams et al., 2009).

One major characteristic of the eastern Athabasca granulite terrane is a block style architecture, with distinct lithotectonic subdomains divided by thick ductile shear zones (*briefly discussed below*; Flowers et al., 2006b; Leslie et al., 2015). Detailed work has demonstrated that the steeply to moderately dipping shear zones that divide subdomains are Paleoproterozoic in age (Williams et al., 2009; Mahan et al., 2006a; Dumond et al., 2013; Regan et al. 2014). The major focus of this contribution will be the older tectonic events, evidence for which is heterogeneously preserved internal to each lithotectonic block or subdomain (Mahan et al., 2008), within the well characterized East Athabasca mylonite triangle (Fig. 2).

The Chipman subdomain is primarily underlain by the heterogeneous ca. 3.2 – 3.4 Ga Chipman tonalite (Chipman panel of Martel et al., (2008))(Hanmer, 1994, 1997). Despite the rarity of similar exposures throughout the mapped Rae domain, Mesoarchean crust and lithosphere has been interpreted to underlay much of the Rae domain (Aspler and Chiarenzelli, 1996; Petts et al., 2014). Along the western

flank of the Chipman subdomain, there are meter to decimeter scale lozenges of Archean mafic granulite and associated felsic granulite (Mahan et al., 2008; Flowers et al., 2008), which record evidence of polymetamorphism interpreted to have both Archean and Paleoproterozoic components (Table 1; Flowers et al., 2008; Mahan et al., 2008). The Chipman mafic dike swarm cross-cuts gneissosity and typically has a strong tectonic fabric defined by a granulite facies metamorphic assemblage (hornblende – plagioclase – garnet – clinopyroxene (Williams et al., 1995; Mahan and Williams 2005; Flowers et al., 2006a; Mahan et al., 2008; Regan et al., 2014).

The Northwestern subdomain is characterized by Neoproterozoic plutonic rocks with subordinate felsic granulite screens (Fig. 2). The subdomain is bounded by the Cora Lake and Grease River shear zones. Plutonic rocks of the Northwestern subdomain range in composition from felsic to intermediate, with subordinate mafic lithologies (*see below*). The largest, and best studied component is the Mary granodiorite (Hanmer, 1994; 1997; Williams et al., 2000; 2014; Dumond et al., 2010). Felsic granulite (classically the “Reeve diatexite” after Hanmer, 1994) commonly forms concordant, semi-continuous screens completely engulfed by the meta-plutonic rocks. Detailed structural work has focused within charnockitic (opx-bearing granitoid; terminology after Frost and Frost, 2008b) lithologies, which preserve evidence for both Neoproterozoic and Paleoproterozoic deformational phases (Dumond et al., 2010; Regan et al., 2014). Early, subhorizontal gneissosity (S_1) has been interpreted to represent Neoproterozoic crustal flow (Dumond et al., 2010), while subsequent folding and the development of a heterogeneously developed axial planar (S_2) fabric developed throughout the region is interpreted to represent Paleoproterozoic (ca. 1.9 Ga) deformation (Regan et al., 2014).

The Upper Deck (also referred to as the southern subdomain (Baldwin et al. 2004; Flowers et al., 2006b)) consists of abundant migmatitic paragneiss traditionally referred to as felsic granulite, mafic granulite, and minor eclogite (Snoeybosch et al., 1995; Baldwin et al., 2003; 2004). The dominant rock type is a garnetiferous quartzofeldspathic gneiss, interpreted to represent a deformed migmatite that has had variable amounts of melt loss and is referred to as the “white gneiss” (Snoeybosch et al., 1995;

Baldwin et al., 2004; Dumond et al., 2015). The white gneiss contains an assemblage of ternary feldspar, quartz, kyanite, garnet, and rutile +/- sillimanite (Snoeyenboes et al., 1995). Although production of eclogite assemblages was initially interpreted to be Paleoproterozoic and approximately 700 my after migmatization (Baldwin et al., 2006), recent work by Dumond et al. (2015) suggests that eclogitization was synchronous with anatexis during the Neoarchean.

Shear zones delineate the major lithotectonic blocks (subdomains) of the east Athabasca mylonite triangle. The Cora Lake shear zone separates the Northwestern subdomain from the Chipman subdomain and is thought to have been active during the waning phases of Paleoproterozoic granulite facies deformation (ca. 1.89-1.87 Ga; Regan et al., 2014). The 1.86 – 1.84 Ga Legs Lake shear zone represents the southeastern margin of the Athabasca granulite terrane, and accommodated ca. 20 km of vertical throw of high pressure granulites of the Chipman subdomain above mid-crustal Hearne domain rocks (Mahan et al., 2003; 2006a,b). The Grease River shear zone marks the northern margin of the Northwestern subdomain, and accommodated ca. 110 km of right-lateral slip at ca. 1.80 Ga (Mahan and Williams, 2005). Recent work by Dumond et al. (2008; 2013) suggests that the Grease River shear zone may also have an older history (ca. 1.9 Ga), but the relative importance and magnitude remains unclear. Based on the kinematics and timing of each shear zone, Regan et al. (2014) interpreted them as having formed during, and as a result of, sequential orogenesis along the western and southern margins of the western Churchill Province.

ANALYTICAL METHODS

Geochemistry

Major and trace element compositional analyses were carried out by Bureau Veritas analytical laboratories in Vancouver, British Columbia. Samples were cut into slabs, and weathered edges were trimmed. Samples were crushed into powders at Bureau Veritas analytical facilities. Samples were

analyzed by X-ray fluorescence (major elements) and inductively coupled plasma - mass spectrometry (trace and rare earth elements; *for detection limits view brochure*).

In-situ monazite U-Th-total Pb geochronology

All monazite grains were identified from full thin section wavelength dispersive spectrometry (WDS) maps, made on a Cameca SX-50 electron microprobe, using a 300 nA current at 15 KeV. The step size was 35µm, with a defocused beam approximately 30 µm in diameter. Grains were found with a spectrometer set on the Ce Mα x-ray line, along with a minimum of two other maps (base maps) of common elements (typically Mg Kα and Ca Kα; Williams and Jercinovic, 2012). Grains were subsequently mapped at high resolution with a focused 200 nA beam, using beam rastering technique with a fixed stage. This provides submicron compositional resolution (see Williams et al., 2006).

In-situ U-Th-total Pb monazite geochronology was performed at the University of Massachusetts, Amherst, on the Cameca SX-100 “ultrachron” electron microprobe (for analytical procedures see Williams et al., 2006; Dumond, 2008). Five spectrometers measured U, Th, Pb, S, Ca, K, Sr, Si, Y, and P, and the full suite of REEs. Each reported date consists of one background analysis and four to eight peak analyses all made within a single compositionally homogeneous domain, thus a single date is the result of 4 – 8 analyses. Background for U, Th, and Pb were acquired using a multipoint method, where 8-10 measurements were taken in positions of no known spectral interference (Allaz et al., 2011). An exponential fit was used to regress background through the peak position to evaluate the background curvature. Lead was measured by integrating counts from two VLPET (very large PET) crystals (*see Williams et al., in review*).

Standardization was done using both natural and synthetic standards. Calibration was performed on pyromorphite (PbPO₄), Brabanite (ThPO₄), and UO₂ for Pb Mα, Th Mα, and U Mβ, respectively. Peak positions were routinely updated on synthetic phosphates for P, Ce, La, and Nd prior to each analytical session. Prior to, and after each analytical session, as well as several times within a session, an in-house

consistency standard, Moacyr (506 $\pm$ 1Ma; Dumond et al., 2008), was analyzed using the same analytical procedure to evaluate the consistency of analyses.

IN SIMS

In-situ secondary ionization mass spectrometry (IN-SIMS) was performed on one sample during a brief visit to University of California, Los Angeles. Zircon was found in the same way as monazite (see above), except with one spectrometer set to Zr $L\alpha$. Several grains were imaged with BSE and CL on a Carl Zeiss SEM at the University of Massachusetts, Amherst. Individual grains were drilled out with a 3 mm drill bit and mounted with a standard (Duluth Gabbro). Analytical setup follow procedures presented in Chamberlain et al. (2010; *and references therein*) for micro zircon using a CAMECA 1270 SIMS at the University of California, Los Angeles. The field aperture was set to 8 μ m, and measurements were taken over thirty cycles on the electron multiplier.

ICP-MS

Laser ablation inductively-coupled-plasma mass spectrometry was performed on zircon separated from one sample. Zircon grains were separated and imaged at the Laserchron center at the University of Arizona in Tuscon, and imaged with a Hitachi 3400N SEM using backscattered detection and cathodeluminescence with a Gatan Chroma CL system. Isotopic analyses were performed on the Thermo Element 2 with 110 cycles at 10 Hz. Precleaning of individual spots was performed with 3 cycles and a 30 μ m spot, and full analyses were performed with a 20 μ m beam. For full analytical procedures see Gerhals and Pecha (2014).

RESULTS

Constraints on the Northwestern subdomain plutonic rocks

Table 2 displays existing U-Th-Pb geochronologic data from the east Athabasca mylonite triangle that pertain to emplacement ages of metagneous rocks. The majority of constraints are presented in

Hanmer et al. (1994; Hanmer, 1997), and were performed via multi-grain TIMS. Although the majority of felsic igneous rocks appear slightly older than the analyzed mafic rocks, given the multi-grain TIMS method employed, we can not account for metamorphic grains developed within mafic endmembers, potentially lowering the aggregate age slightly. Collectively, plutonic rocks of the Northwestern subdomain can be thought of as a ca. 2.6 Ga composite batholith.

Archean plutonic rocks of the northwestern subdomain contain evidence for at least two periods of penetrative deformation (Dumond et al., 2010; Regan et al., 2014). The earliest tectonic fabric has a subhorizontal enveloping surface throughout the entire northwestern subdomain (Dumond et al., 2010; Regan et al., 2014). This strong tectonic fabric parallels compositional variability present within the composite batholith. This major structural fabric, however, has been modified by subsequent, ca. 1.9 Ga open, upright folds that plunge shallowly to the southwest (Dumond et al., 2010; Regan et al., 2014). Unfolding of F_2 results in a southwest dipping S_1 ($\sim 12^\circ$). We assume that this dip, at least to some degree, represents the pre D_2 configuration S_1 (ie. dipping slightly to the southwest). Therefore, the early compositional layering is currently tilted, and thus the Northwestern subdomain may represent an oblique section of the ca. 2.63-2.60 Ga composite batholith with deeper portions exposed in the northeast.

The northeastern part of the plutonic belt is dominated by intermediate and subordinate mafic units of the bohica suite (Hanmer, 1994). Toward the southeast, these units grade into granodioritic to granitic units, which are largely textural variants of the Mary batholith (Hanmer, 1994). Therefore, the Neoarchean composite batholith grades from more mafic endmembers at structurally lower ends to an increasing felsic component with shallower structural depth. The lack of discrete mylonite zones at compositional boundaries, and the pervasive high strain internal to S_1 compositional layering indicates that compositional endmembers were not tectonically juxtaposed, but were emplaced and deformed together as one coherent composite batholith.

Rocks of the Neoarchean plutonic suite have a broad compositional range, but contain similar textural characteristics. A vast majority of the rocks contain plagioclase porphyroclasts, with variable amounts of orthopyroxene, hornblende, clinopyroxene, garnet, +/-kspars +/- quartz. Accessory phases include apatite, ilmenite, biotite, and zircon. However, both equigranular and megacrystic, pyroxene-absent, granites are also common (ie. Beed granite and correlatives, see below).

Geochemistry

A sampling of nine ca. 2.63 – 2.60 Ga plutonic rocks along a northeast to southwest transect, oblique to $S_{0/1}$ strike have been used to characterize the Neoarchean plutonic complex (Table 3). SiO_2 contents range from 58.01 – 73.71 weight percent. Plots of major oxides (Harker diagrams) plotted against SiO_2 display tight, linear trends (Fig. 5). With increasing SiO_2 , all samples decrease in CaO, Fe_2O_3 , TiO_2 , P_2O_5 , and MnO, but MgO displays no significant variation.

Rocks range from ferroan to magnesian with respect to the Fe index ($Fe_2O_3/MgO+Fe_2O_3$) of Frost and Frost (2008a; Fig. 6a). Samples define a linear trend with a shallow slope on modified alkali lime index ($Na_2O+K_2O - CaO$ against SiO_2) diagrams and plot as calc-alkalic to alkalic-calcic (Fig. 6b). All samples plot within the volcanic arc field in all four of the tectonic discrimination diagrams based on Y, Yb, Rb, Nb, and Ta (Fig. 7; Pearce et al., 1984). Some of the higher SiO_2 samples (Beed Granite), plot near the boundary of rift-related granite fields.

All samples are LREE enriched, and HREE depleted, and show a strong, concave up, negative slope when normalized to chondrites (Fig. 8a; Sun and McDonough, 1989). That is, samples are characterized by a LREE negative slope, and a fairly flat and depleted HREE slope. La_N/Sm_N (chondrite normalization of Sun and McDonough, 1989) ranges from 4.31 – 12.10. Incompatible element trends display strong depletions in high field strength elements, and enrichments in large ion lithophile elements and Pb when normalized to primitive mantle (Fig. 8b; Sun and McDonough, 1989). Figure 9 displays a comparison between classic Andean arc rocks presented in Bryant et al. (2006), and the 2.60 Ga plutonic

rocks from the study area, which show many similarities. These trends are most commonly interpreted in terms of derivation from a metasomatized lithospheric source (Cousens et al., 2001; Chiarenzelli et al., 2010).

Published Nd isotopic analyses from the northwestern subdomain show remarkable similarity (Fig. 10; Hanmer, 1997). Samples range from -1.84 to 1.55 $\epsilon_{\text{Nd}(t)}$, and are slightly below contemporaneous depleted mantle (Hanmer, 1997). Model ages (T_{DM}) range from 2686 to 3019 Ma, and thus the northwestern subdomain plutonic rocks are likely derived from both juvenile and older crustal sources or experienced variable amounts of contamination from Mesoarchean crust. On Nd evolutionary diagrams, all samples share similar slopes and a relatively small ϵ_{Nd} range. Differences may reflect varying LREE concentrations due to fractionation (Fig. 8a). When compared with the whole rock geochemistry presented above, it seems likely that the entire northwestern subdomain consists of a large and petrogenetically linked, Neoarchean composite arc batholith, that was built in ca. 30 my, or less.

Comparison with other 2.6 Ga plutonic rocks

The majority of the Rae domain of the western Churchill Province is underlain by Meso to Neoarchean plutonic rocks (Hoffman, 1988). However, published geochemical data for Neoarchean rocks in the Rae domain are sparse. Two major suites of granite are exposed along the southeastern margin of the Chipman subdomain, and into the Legs Lake shear zone: The Fehr and Stevenson granites (Fig. 2; Mahan et al., 2003). Data from the Fehr Granite (Koteas et al., 2010) plot collinearly on Harker diagrams with the rocks from the northwestern subdomain (Fig. 5). The Fehr granite also plots in the volcanic arc granite field on tectonic discrimination diagrams (Fig. 7), and identical trace element patterns (Fig. 6). These data suggest a potential petrogenetic relationship between the Fehr granite and plutonic rocks exposed within the Northwestern subdomain (ca. 2.6 Ga; Hanmer, 1997; Koteas et al., 2010; *supplement*). These data suggest that the Fehr granite may be a member of the Neoarchean plutonic complex exposed in the Chipman subdomain. There is no existing whole-rock geochemistry or zircon

geochronology on the Stevenson granite, but it appears to be older (ca. 2.68 Ga) based on new data (Regan et al., 2016). Ca. 2.7 – 2.68 Ga magmatism is common in both Rae and Hearne domains, and is interpreted as a separate magmatic event from the plutonic complex studied herein.

Data from Angikuni Lake, 450 km NE along the Rae-Chesterfield-Hearne boundary (Cousens, 1998; Aspler et al., 1999; Berman et al, 2007) within the vicinity of the geophysically defined Snowbird Tectonic zone provide an additional spatial comparison to test if the Neoarchean plutonic complex exposed in the east Athabasca mylonite triangle continues along the margin of the Rae domain. Little geochronologic data exist for metaigneous rocks at Angikuni Lake, but preliminary data suggest ages ranging from 2.7 – 2.6 Ga (Cousens, 1998; Aspler et al., 1999). The older populations are likely related to the initial growth of the Ennadaidai – Rankin greenstone belt (Aspler et al., 1999), however, ca. 2.6 Ga plutonic rocks that straddle the geophysically defined STZ (zone 2, 6, and 7; Fig. 11; Cousens et al., 1998; Aspler et al., 1999). On Frost diagrams (Frost et al., 2001; Frost and Frost 2008a), the rocks plot as ferroan to magnesian, and form linear trends on the modified alkali-lime index, plotting from alkali-calcic to calc alkalic. Similar to the Neoarchean plutonic rocks in the Athabasca granulite terrane, they plot in the volcanic arc granite fields on all Pearce diagrams (Pearce et al., 1984), and contain enrichments in their LILE and depletions in HFSE. Sm-Nd isotopic systematics are also quite similar to published data from Hanmer et al. (1994), with rocks of the east Athabasca mylonite triangle having very similar, time integrated, ϵ_{Nd} values. In all respects, the ca. 2.6 Ga plutonic rocks are similar to the east Athabasca mylonite triangle and suggest that the eastern Rae domain may be more broadly characterized by Noearechean arc magmatism.

In-situ monazite U-Th-total Pb geochronology

Four geochronology samples were chosen to constrain the timing of igneous (arc magmatism) event and to further limit the timing of peak phases of deformation/anatexis. Samples containing monazite were modelled (supplemental file; Regan, 2016) in an attempt to better link metamorphic

conditions to the regionally developed structure, specifically the development of S_1 (see supplement; Regan, 2016). Geochronology and major and trace element data from monazite is provided in Table 4.

Sample S823b

Sample S823b is located at East Hawkes Lake. It is a very low strain and equigranular sample of the Mary Granite, with little to no penetrative tectonic fabric, and an assemblage of plagioclase, K-feldspar, quartz, minor orthopyroxene and accessory zircon, rare monazite, ilmenite, and apatite. Four monazite grains were mapped, and three grains analyzed. The grains are generally unzoned with minimal compositional variation, with a very outerrim of m1 having slightly lower Y_2O_3 , (0.77 weight percent) compared to the remainder of the grains containing greater than 0.83 weight percent.

Five sets of analyses were performed on the monazite crystals. Five core and outercore analyses yielded an age of 2611 ± 30 Ma (2σ errors reported; MSWD: 2.5) interpreted to reflect the crystallization age. One analysis of the outerrim yielded an age of 2574 Ma, interpreted to represent metamorphic monazite growth.

Sample 10W-094g

Sample 10W-094g is from the southwestern shore of Cora Lake, along the southern flank of the Cora Lake shear zone (Fig. 2; preliminary results on monazite rims in leucosome presented in Regan et al., 2014). The sample is a garnet-rich migmatite adjacent to a mafic granulite boudin and a cross cutting Chipman dike (*for petrologic description, see supplement; Regan, 2016*). The sample was cut to provide a fold profile view. It is characterized by layers of garnetite interlayered with pods and layers of granitoid leucosome. A total of 20 monazite grains were mapped. Twenty new analyses of monazite domains were analyzed and dated from both the garnetite and leucosome.

The garnetite contains several complexly zoned grains. They typically have high Y_2O_3 and low ThO_2 cores, interpreted to have grown before anatexis, surrounded by a higher ThO_2 and lower Y_2O_3

rims, interpreted to have grown during melting. The leucosome contains monazite with three domains and several grains that have rims aligned parallel to S_2 (Regan et al., 2014). Commonly, monazite cores in the leucosome are characterized by high Y_2O_3 , and low ThO_2 content, interpreted to have grown during prograde metamorphism. High ThO_2 innerrims, that have low Y_2O_3 contents and vary in morphology are interpreted as syn-melting (see Dumond et al., 2015). Surrounding these domains are outer rims that are both low Y_2O_3 and ThO_2 , which are consistently parallel to S_2 fabric (syn- S_2 ; Regan et al., 2014).

Eighteen sets of analyses from the garnetite yield three distinguishable populations. Cores yield a poorly constrained age of 3138 ± 40 Ma. Two other sets of core analyses that are similar in composition, but with slightly higher ThO_2 yielded ages of 2962 ± 9 Ma and 2996 ± 57 Ma. High ThO_2 and low Y_2O_3 rims yield a wide spread of ages from 2590 Ma to 2525 Ma, and are interpreted to have grown during partial melt crystallization and tectonism. The leucosome contains a bimodal distribution. Both cores and inner rims yield indistinguishable dates of ca. 2565 Ma. Rims yield a weighted average of 1900 ± 3 Ma based on eight sets of analyses (Regan et al., 2014).

These data suggest that migmatization occurred during the Neoproterozoic. The drastic increase in ThO_2 and drop in Y_2O_3 is interpreted to reflect garnet growth during biotite dehydration melting similar to migmatites analyzed by Dumond et al. (2015) in the neighboring Upper Deck subdomain. This interpretation is also consistent with the lack of S_2 within the garnetite, and the corresponding lack of Paleoproterozoic ages. Analyses from the leucosome confirm that S_2 developed at 1.9 Ga (*see also* Dumond et al., 2008, 2010; Regan et al. 2014). The two Mesoproterozoic populations remain enigmatic. A ca. 3.2 Ga age coincides with some of the igneous grains from the Chipman tonalite (Hanmer et al., 1994; Martel et al., 2008). However, the 2.9 Ga ages do not correspond to any known igneous intrusive suite, or recognized widespread-metamorphic event within the immediate vicinity (Mahan et al., 2006a). Mahan et al. (2006a) noted monazite populations from a felsic granulite in the eastern Chipman subdomain, but the significance and regional extent remains largely untested. These ages could reflect: 1) the intrusion of mafic granulite bodies that remain unconstrained in age but have ca. 2.8-3.1 Ga depleted mantle model

ages (Sm-Nd; Flowers et al., 2008), 2) an earlier phase of metamorphism/metasomatism, 3) the presence of detrital cores, or 4) partial alteration of older grains.

Sample 12R-054

Sample 12R-054 is a sample of Beed granite (*also used for isochemical modeling, see supplemental document*; Regan, 2016), which is a megacrystic biotite granite with coarse K-feldspar and plagioclase porphyroclasts, but contains little garnet. The sample contains a strong S_2 with no observable S_2 , but is folded on the map scale (Regan et al., 2014). Twenty four monazite grains were mapped. Ten sets of analyses were attained on all monazite domains, three of which were presented in a separate contribution (Regan et al., 2014).

Four distinctive morphological populations are apparent: 1) Innercores (low ThO_2 , high Y_2O_3) are surrounded by 2) outcores (low Y_2O_3), 3) high ThO_2 outcores, and 4) innerrims containing moderate Y_2O_3 and high ThO_2 . Cores yielded a single population of 2608 $\pm$ 8.4 (MSWD: 2.5). Outcores and innerrims yielded separate and slightly younger dates averaging 2577 $\pm$ 22 Ma (MSWD: 6.8), and likely include more than one population. These results are interpreted to represent the timing of igneous crystallization and subsequent lateral flow of the Beed granite associated with a small amount of garnet growth. Lastly, a population of 1917 Ma is represented by very fine ($< 5\mu m$) rims (Regan et al., 2014).

Sample 11G-020

Sample 11G-020 is a biotite – garnet gneiss located on the northwestern shore of Reeve Lake, along the southern margin of the Grease River shear zone (Figure 2; Mahan and Williams, 2005). The outcrop contains a strong S_1 gneissic layering that is folded about a shallowly plunging fold axis with little to no S_2 development (Fig. 4a). The rock was cut to fold profile view, and contains plagioclase, quartz, potassium feldspar, garnet and biotite. Two thin sections were cut in a fold-profile orientation, and parallel to the regionally developed L_1 . Biotite is texturally late, and has overgrown garnet

porphyroblasts, which are fractured and highly embayed. Garnet ranges in size from ca. 0.5 mm to 3 cm and contains tonalitic leucosome aligned in the S_1 fabric.

M1 is compositionally distinct and partially armored by garnet. The crystal is a composite monazite grain, with several distinctive generations of monazite cement. Approximately six distinguishable cores make up the framework of the grain, while cement is made up of two compositionally distinct monazite generations. Composite grains consist of high Y_2O_3 innercores, and low Y_2O_3 outercores. Although, we refer to high Y and Th as innercores, this compositional population is only present towards the outer portion of the entire composite crystal, indicating that this monazite generation may have been a rim on a once more coherent grain. It is important to note, that aside from the innercores in the composite, grains, Y_2O_3 contents are relatively constant throughout the remainder of the monazite crystal. The innermost, or oldest generation of cement contains 7.9 wt percent ThO_2 . The outer generations of monazite cement contains 4.6-4.9 wt % ThO_2 .

Cores yield two Neoproterozoic populations, 2617 $\pm$ 10 Ma and 2572 $\pm$ 13 Ma. Unlike other samples in this study, the low Y_2O_3 population is 2617 Ma, and with the high Y_2O_3 innercores all yielding ages younger than 2590 Ma. Therefore, we interpret these data to indicate anatexis during prograde metamorphism, and emplacement of the 2.60 Ga plutonic rocks that surround this screen of diatexite. Subsequent tectonism at 2570 Ma caused deformation of the leucosomal pockets and biotite after garnet growth. The oldest generation of cement, which contains higher ThO_2 , yielded an age of 1929 $\pm$ 8 Ma. Two age populations were acquired during analysis of the outermost cement: 1900 Ma (2 sets of analyses) and 1867 $\pm$ 10 Ma (1 set of analyses). These populations correspond to all analyzed populations from the matrix grains, and weighted means are calculated including these analyses (*see below*). These data provide support to the interpretation of matrix grains below, and are consistent with published geochronology from the northwestern subdomain.

Matrix grains show a distinct series of compositional zones. Although compositional variation is slight within matrix grains, three distinctive compositional domains have been identified. Cores contain over 0.20 wt % Y_2O_3 and less than 3.5 wt % ThO_2 , while innerrims contain less than 0.05 weight percent Y_2O_3 and greater than 3.5 wt % ThO_2 . Outerrims are rare, approximately 8 microns wide, and typically contain less than 0.05 wt % Y_2O_3 and less than 3.5 wt % ThO_2 .

Thirteen analyses of matrix grains produced three distinctive, and compositionally consistent populations. Three sets of core analyses yielded an average of 1927 $\pm$ 5 Ma. Innerrims, or low Y_2O_3 high ThO_2 domains yield an average of 1899 $\pm$ 4 Ma based on eight analyses. Low Y_2O_3 and ThO_2 rims yield an average of 1862 $\pm$ 7 Ma. These data suggest that additional biotite growth, retrograde metamorphism due to static hydration, and garnet consumption initiated at ca. 1927 Ma due to the relative high Y_2O_3 contents relative to other populations, which is far lower than cores of M1. Two analyses of outerrims yielded an average of 1867 Ma, which are interpreted as the result of further hydration during contractional uplift and the associated peacemeal activity of several shear zones (Dumond et al., 2013), including the proximal and episodic Grease River shear zone (Dumond et al., 2008).

Zircon geochronology of 11G-020

11G-020 is a migmatitic paragneiss (“Reeve diatexite” of Hanmer, 1994) from Reeve Lake located along the southern margin of the Grease River shear zone. Garnet size is bimodal. Coarse garnet is generally associated with pods of tonalitic leucosome (Fig. 15a), whereas fine grained garnet occurs throughout the matrix. Both coarse garnet and associated leucosome is deformed, and garnet is fractured and dismembered along the S_1 fabric, indicating that anatexis occurred prior to, or during, deformation. Zircon was identified in-situ. Four zircon grains were drilled out of the thin section, two from the leucosome adjacent to coarse garnet, and three from the immediate host rock (results in Table 4).

Zircon grains from the leucosome have cores and rims (Fig. 15c). Cores are rounded, display oscillatory zoning, and give intermediate BSE and CL signals. Rims have similar BSE and CL signals,

and truncate internal zonation in cores. Cores are interpreted to be inherited from the protolith, and the rims are interpreted to be associated with crystallization of partial melt. Two rim analyses yielded $^{207}\text{Pb}/^{206}\text{Pb}$ dates of 2610 ± 6 Ma and 2608 ± 16 Ma. Analysis of a core (Fig. 15e) yielded a date of 2691 ± 11 Ma. Analysis of several other core grains from the matrix yielded two ages of 2789 ± 30 Ma and 2827 ± 22 Ma, consistent with a detrital origin. These results suggest that deposition of this rock occurred at or after ca. 2.69 Ga, and that it partially melted during emplacement of the arc batholith in which it is contained, consistent with monazite results above.

Two-hundred and seven U-Pb detrital zircon analyses were performed from a grain separate (full dataset in supplement). $^{206}\text{Pb}/^{207}\text{Pb}$ ages range from the Mesoarchean down to 1.9 Ga. A probability density diagram shows peaks, in decreasing relative probability (Fig. 15f), at 2650 Ma, 2692 Ma, 2790 Ma, 1905 Ma, 2538 Ma, 2895 Ma, 2372 Ma, 2120 Ma, and several smaller Mesoarchean peaks. However, owing to the samples complex history and polymetamorphic nature, we've plotted % concordance vs. age. Between 2.60 Ga and 1.9 Ga, data loosely define a u-shape indicative open system behavior between the Neoarchean and Paleoproterozoic (Fig. 15g), and data that fall between these two populations do not represent reliable detrital data (Flowers et al., 2008). Furthermore, considering the rock was partially melted during ca. 2.60 Ga plutonism, and that zircon data form a thick data cluster between 2780 Ma down to 2530 Ma, differentiating detrital zircon from grains that formed during anatexis is currently not attainable at the resolution of the single U-Pb analyses without detailed major and trace element chemistry of zircon grains. However, the data further demonstrate that deposition was quickly followed by tectonic burial and migmatization, and indicate proximity to Mesoarchean crust during deposition.

DISCUSSION

The case for a Neoarchean arc plutonic complex

The Athabasca granulite terrane is composed of Meso to Neoarchean plutonic rocks (Hanmer, 1994). The Neoarchean granitoids that dominate the entire northwestern subdomain, and varying amounts of adjacent Chipman and Upper deck subdomains, are broadly coeval (ca. 2.63 – 2.60 Ga)(Hanmer, 1997), and contain Nd isotopic systematics compatible with derivation from a relatively homogenous, metasomatized lithospheric source (Hanmer et al., 1994), with evidence for interaction with older, Mesoarchean, crust. The suite of rocks has a wide range of SiO₂ contents (> 15 wt %) and modal mineralogy, and grade from more mafic rocks in the northeast, to more felsic endmembers in the southwest. Composite compositional/S₁ layering is interpreted to have been tilted to the SW prior to or during to F₂ folding, and thus, grades from mafic to more felsic lithologies with increasing structural height.

Circa 2.6 Ga plutonic rocks within the Athabasca granulite terrane show marked similarity to modern arc plutons. They span a wide compositional range, form shallow calc-alkaline trends with respect to the modified alkali lime index (Frost and Frost, 2008a) and plot in the volcanic arc granite field on tectonic discrimination diagrams (Pearce et al., 1984). Trace element signatures display systematic enrichments in LILE and Pb, and corresponding depletions of HFSE when normalized to primitive mantle (Sun and McDonough, 1989), which is characteristic of high T metasomatism similar to that thought to occur in an overriding mantle wedge (Sun and McDonough, 1989; Chiarenzelli et al., 2010). These attributes are also seen in similar aged rocks in adjacent subdomains (Fig. 6; Kopf, 1999; Koteas et al., 2010). Given the range in composition and the ubiquitous major and trace element composition indicative of arc processes, we interpret that the ca. 2.6 Ga plutonic rocks to have formed above an active subduction zone, and that rocks represent a largely intact arc plutonic complex.

Structural measurements combined with new and published thermobarometric calculations (*see supplement*; Regan, 2016) provide insight into the original geometry of the composite batholith. S₁ is defined by both a strong planar gneissosity, and compositional layering. Only rare fold closures have been observed, but do not involve large recumbent folding, and are likely of a local significance.

Furthermore, there is no increase in strain across discrete compositional changes, or layers, suggesting that compositional heterogeneity was, at least in part, close in orientation to S_1 (Dumond et al., 2010; Regan et al., 2014). Layering currently dips shallowly to the southwest, parallel to the F_2 hinges, modified by a variably developed axial planar fabric (S_2 ; Dumond et al., 2010; Regan et al., 2014). Thermobarometry from the southwestern edge (more felsic) of the Northwestern subdomain yield estimates of 0.9 GPa and 750° C (Dumond et al., 2010). Calculations from the central Northwestern subdomain are nearly identical (East Hawkes Lake; Williams et al., 2000; Regan et al., 2014). Forward petrologic modeling of the Beed granite indicates it was deformed at higher pressures (~1.1 GPa) than those observed to the southwest (*supplement*; Regan, 2016). We suggest that compositional layering was tilted after the formation of a strong S_1 , and that the Northwestern subdomain is a tilted section (~12 ° dip) near the root of an Archean arc batholith.

Timing of events: implications for fingerprinting tectonic processes

The eastern Rae domain within the study region displays a three step evolution involving: 1) arc plutonism, 2) high grade metamorphism (Table 5), and 3) crustal flow (Dumond et al., 2010; 2015). Arc construction took place from ca. 2.63 – 2.60 Ga (Hanmer, 1997), and may extend to the Northeast along the Rae margin.

Timing and conditions of deformation

Paired forward petrologic modeling and in-situ geochronology of the Beed granite were undertaken to augment existing data from the southwestern portion of the Northwestern subdomain. Based on monazite geochronology, the Beed granite is interpreted to have crystallized (ca. 2.61 Ga) and to have been deformed at approximately the same time as the rest of the plutonic rocks of the Northwestern subdomain (ca. 2.57 Ga)(Dumond et al., 2010). Circa 2.57 Ga corresponds to the development of a penetrative fabric, with top to the east kinematics (Dumond et al., 2010), but forward modelling suggests the Beed granite experienced slightly higher pressures (ca. 1.1 GPa; see *supplement*;

Regan, 2016) than the felsic plutonic rocks to the southwest (Williams et al., 2000, 2014; Dumond et al., 2010). Based on a geobaric gradient of 1 GPa equal to 37 km, and based on a ca. 0.2 GPa difference observed within the subdomain, the vertical thickness represented by present exposures may have been on the order of ~ 7 km. Geochronology of the Reeve diatexite suggests that it was deposited after 2.68 Ga in close proximity to a continental influence, and partially melted during emplacement of the plutonic complex into the lower continental crust at ca. 2.60 Ga (Williams et al., 2001), indicating that tectonic burial occurred in less than 80 my. Furthermore, these data confirm that S_1 was roughly synchronous throughout the Northwestern subdomain.

Migmatites are common within the ca. 3.2 Ga Chipman batholith (Chipman subdomain; Mahan et al., 2008), particularly in the western segment where they are typically associated with mafic granulite (Mahan et al., 2008; Flowers et al., 2008). Mahan et al. (2008) determined that early, subhorizontal gneissic layering corresponds with the development of a clinopyroxene + plagioclase assemblage (M_1) that likely formed at ca. 1.35 GPa and temperatures in excess of 850 °C. This event was followed by a med-P event that involved the partitioned development of folds and an axial planar fabric ($M_{3a,b}$ of Mahan et al., 2008). Sample 10W-094G, presented here, contains two primary compositional components that define the migmatitic layering. Partial melting, and the growth of abundant garnet, occurred at P-T conditions greater than 1.3 GPa and temperatures exceeding 835 °C (see supplemental data; Regan, 2016). Monazite composition shows a drastic decrease in Y_2O_3 and increase in ThO_2 between 2.59 and 2.57 Ga. These are interpreted to have grown during partial melting and development of the S_1 migmatitic layering. The second deformation event occurred at pressures of ca. 1.15 GPa, approximately 0.2 GPa less than earlier phase, consistent with thermobarometry and zircon geochronology presented in Mahan et al., (2008; Flowers et al., 2008). We interpret that decompression of the Chipman subdomain relative to the neighboring Northwestern subdomain to have occurred between D_1 and D_2 , but timing of this phase of decompression remains unconstrained.

Crustal Flow

Neoproterozoic deformation in the Athabasca granulite terrane was widespread, and involved extensive partial melting. Furthermore, monazite geochronology suggests that anatexis and penetrative deformation postdated the igneous crystallization of the Neoproterozoic plutonic complex (ca. 2.59 – 2.54 Ga). Therefore, subsequent to voluminous plutonism, the region was subjected to increasing pressures and temperatures, which apparently resulted in subhorizontal flow (Dumond et al., 2010). Neoproterozoic sediments (< 2.68 Ga) are exposed within the region (11G-020), and thus suggests that some components of the region were at the surface prior to D₁. These data suggest that supracrustal rocks went from the surface to the deep crust in less than ca. 80 my, at least in part during the plutonic thickening. Flow occurred throughout the 2.6 Ga plutonic rocks at conditions ranging from 0.9 GPa to 1.1 GPa and 700 to 775°C, based on new and existing P-T data. Dumond et al. (2010) interpreted the flow as caused by pressure gradients similar to the modern Himalaya (Beaumont et al., 2004), but the possibility exists that lateral flow accommodated further crustal thickening of the Rae towards the Hearne domain. We suggest that the whole Northwestern subdomain may represent a several km thick zone of crustal detachment and flow that developed shortly after the plutonic rocks were emplaced, compatible with the consistent top to the east kinematics preserved throughout the structural thickness of the entire northwestern subdomain (Fig. 17).

Accretionary Model

The eastern Rae domain preserves evidence of a three step evolution. First, voluminous arc magmas were emplaced into Meso to Neoproterozoic basement. This was followed by a phase of deformation and high grade metamorphism at lower crustal depths showing evidence of increasing pressure through time (Dumond et al., 2015). This prograde metamorphism and tectonism is interpreted to reflect thrusting and shortening of the arc (Baldwin et al., 2004; Mahan et al., 2008; Dumond et al., 2010; Regan et al., 2014; Williams et al., 2014). Granulite facies metamorphism postdated magmatism. Finally, the juvenile lithologies of the Neoproterozoic arc plutonic complex accommodated subhorizontal lateral flow throughout the entire Northwestern subdomain. This may represent gravitationally driven

flow of the melt-weakened lower crust (Dumond et al., 2010), or lateral flow of the lower continental crust, the latter of which is compatible with the consistent kinematics from the top to the bottom of the composite arc batholith, and favored here.

The sequence of events recorded by rocks in the eastern Rae domain are most aptly categorized as having formed within an environment similar to a modern active margin. Arc construction was followed by crustal thickening, high grade metamorphism, crustal flow, and anatexis. The cessation of arc magmatism is interpreted to represent the onset of collisional orogenesis. Abundant migmatite formed during the waning phases of arc plutonism, and likely during this phase of collisional orogenesis (10W-094g; Baldwin et al., 2004; Mahan et al., 2006a; Dumond et al., 2010, 2015; Regan et al., 2014), and thus may represent the source of such aluminous and alkalic magmas indicative of lower crustal melting (Frost et al., 2001). Therefore, the region provides a tangible exposure of the transition of mantle-driven arc magmatism to a lower crust derived, syn-collisional type magmatism (Pearce et al., 1984).

Petrology and geochemistry of igneous rocks are two of the major tools for constraining the tectonic context of ancient terranes. Three major tectonic stages are inherent to the eastern margin of the Rae domain including: 1) arc magmatism, 2) deformation-metamorphism associated with crustal thickening (Dumond et al., 2015), and 3) flow of the melt-weakened lower crust (Dumond et al., 2010). The ca. 2.63 – 2.60 Ga magmatism contains the petrology and geochemical signature of arc magmatism. Second, subsequent to intrusion of the voluminous plutonic rocks of, the entire region underwent crustal thickening and high grade metamorphism/migmatization. Finally, lateral flow of the lower continental crust during granulite grade metamorphism with kinematics consistent over a large region encompasses both a large area and structural depth (throughout the Northwestern subdomain, > 0.1 GPa difference or ca. 7 km) suggesting top to the east flow. The combination of arc related geochemistry, subsequent thickening, partial melting, and lateral flow strengthens the case for a collisional boundary in the eastern Rae domain during the late Neoarchean.

Regional correlations and arc geometry

Arc rocks typically define linear arrays of plutonic or volcanic rocks and structures that delineate the general trend of a subduction zone. Identifying these trends in the Archean may be difficult due to probable modification during subsequent Proterozoic and/or Phanerozoic tectonism (Hoffman, 1988). The western Churchill Province is underlain by several Neoarchean plutonic complexes, but there are significant regions containing little to no geochronology or field data, making these correlations difficult and preliminary. Circa 2.6 Ga granitoids exist in the Snowbird Lake region to the Northeast along the geophysical lineament of the STZ (Martel et al., 2008), but there is no available whole-rock geochemical data to assess whether they too contain an arc signature. There are several ca. 2.60 Ga plutonic rocks to the northwest, but again geochemical data are very limited to constrain tectonic heritage (Hartlaub et al., 2005; Bethune et al., 2013). Circa 2.5 – 2.3 Ga plutonic rocks dominate in the Zemplin and Beaverlodge subdomains (Hartlaub et al., 2007; Bethune et al., 2013; Ashton et al., 2013), whereas the Hearne to the southeast contains both Archean and Proterozoic supracrustal rocks, with little evidence for plutonism at ca. 2.6 Ga (Aspler et al., 2002a; Van Breeman et al., 2007). Geochemical data from zone 2 and 7 of the Angikuni Lake region (Aspler et al., 1999) are chemically similar to those discussed above (Cousens, 1998), and are indicative of an arc setting. We tentatively suggest that a belt of arc-related plutonic rocks may continue along the eastern margin of the Rae domain to the Northeast, perhaps as far as Angikuni Lake, and that this plutonism was closely followed by a collisional event.

The Macquoid orogeny has been recognized throughout the northern Rae and Chesterfield (formerly northwestern Hearne) blocks of the western Churchill Province (MacLachlan et al., 2005; Berman et al., 2007; Pehrsson et al., 2013). It is thought to represent the collision between Rae and Chesterfield (formerly NW Hearne domain; Berman et al., 2007) following a protracted phase of crustal-scale deformation and magmatism from ca. 2.60-2.50 Ga (Davis et al., 2006). The Athabasca granulite terrane is ca. 475 km to the southwest of the Chesterfield block, but contains evidence of a similar age

and scale event within the eastern Rae domain, that was preceded by arc magmatism, and is interpreted as the southern continuation of this orogenic belt.

Finding a suture

The STZ has recently been interpreted to correspond with a major Paleoproterozoic suture (ca. 1.91 Ga; Berman et al., 2002; 2007), between the Rae and Hearne domains (Martel et al., 2010; Pehrsson et al., 2013). The resurrection of this model hinged, within the study region, on the presence of eclogite-facies assemblages recognized in Snoeyenbos et al. (1995), and interpreted by Baldwin et al. (2004) as having formed at ca 1.9 Ga. However, these data have been subject to rigorous petrologic modelling and in-situ geochronology that has firmly established a Neoarchean origin for these assemblages (Dumond et al., 2015). Therefore, eclogite facies metamorphism preserved within the upper deck subdomain formed during the ca. 2.59-2.55 Ga event discussed here (Dumond et al., 2015), and immediately post dated extensive arc plutonism, and can no longer be considered a Paleoproterozoic component of the region.

Many contributions have described that the eastern Rae domain and STZ corresponds to a large region of Paleoproterozoic nested shear zones (Mahan and Williams, 2005; Dumond et al., 2014; Regan et al., 2014). Many of these shear zones contain thick zones of mylonite, and accommodated major differential motion between lithotectonic subdomains (Mahan and Williams, 2005; Flowers et al., 2006a; Regan et al., 2014). However, none appear to correspond with a significant change in rock type, geochemistry, or geologic history. Ca. 2.6 Ga plutonism occurs both sides of each shear zone (Hanmer, 1994). 2.6 Ga granitoids have been recognized east of the Legs Lake shear zone within the Hearne domain, suggesting that if a suture is exposed on the ground, it existed at or prior to 2.60 Ga (Regan et al., 2016). If a Proterozoic suture is present, it must occur further east of the margin described here, and is therefore seemingly unrelated to the geophysically defined STZ.

Taken together, all these observations suggest that the Rae and Hearne domains, or something outboard of the Hearne domain, collided during the Neoarchean, during a phase of collisional orogenesis

at ca. 2.60 Ga – 2.55 Ga. Furthermore, the change from infant arc processes in the central Hearne at ca. 2.7 Ga, to a fully developed continental arc by ca. 2.65 Ga along the eastern Rae is consistent with a western Pacific modern analogue originally proposed by Cousens et al. (2004) and Davis et al. (2004). The growth of many volcanic edifices (Aspler and Chiarenzelli, 1996) as a result of foundering oceanic lithosphere (Cousens et al., 2004), progressed to more evolved subduction beneath the eastern Rae domain by 2.63 Ga (Hanmer, 1997). This model would collectively account for the differences between the Northwestern and central Hearne domains as well as the temporal variations in magmatism and tectonism. Furthermore, subduction beneath the eastern Rae domain would account for a lack of arc-related ca. 2.60 Ga granitoids in the central and northern Rae domain (Davis and Bleeker, 1999; Hinchey et al., 2009; *and references therein*), and a lack of plutons of this age in the neighboring Hearne domain.

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Figure Captions

Figure 1: Lithotectonic map of the western Canadian Shield (modified from Aspler et al., 2002b). The Snowbird Tectonic Zone defines the enigmatic boundary between Rae and Hearne domains of the western Churchill Province. Box outlines triangular region (east Athabasca mylonite triangle) depicted in figure 2. Angikuni Lake, 450 km to the northeast of the study region is outlined with a white box. Inset shows location of figure relative to North America.

Figure 2: Geologic map of the east Athabasca mylonite triangle (Gilboy, 1981; Hanmer, 1994) also displaying sample localities for this study. The region can be split into three distinctive lithotectonic domains that are divided by thick ductile shear zones. Geochronology samples labeled.

Figure 3: Photographs of S_1 and S_2 relationships from the Chipman and northwestern subdomains. A) Folded Chipman tonalite (ca. 3.2 Ga), where the early tectonic fabric (S_1) is overprinted by an axial planar fabric (S_2); B) folded Bohica intermediate gneiss from the northwestern subdomain (ca. 2.6 Ga), where feldspar porphyroclasts form ribbons along the early, subhorizontal S_1 , and are folded and variably aligned with S_2 (Pen oriented along the S_1 enveloping surface and the pencil is aligned with S_2); C) Meter-scaled fold of a Chipman dike that is cross cutting S_1 , but is folded about an F_2 fold axis; D) Early tectonic layering in the Bohica intermediate gneiss with little to no S_2 development showing abundant asymmetric porphyroclasts and S-C-C' geometries indicative of top to the east shear; E) Highly transposed Chipman anorthosite with a strong S_2 , with little to no evidence of earlier tectonic gneissic layering; F) Dextral S-C-C' kinematic indicator in the Bohica suite developed along a strong L_2 .

Figure 4: Photographs showing felsic granulite migmatites from various locations throughout the Chipman and northwestern subdomain. A) Reeve Lake diatexite (sample 11G-020) with a strong S_1 gneissic layering with little to no evidence of S_2 ; B) Cora Lake felsic granulite (sample 10W-094g)

displaying strong evidence for anatexis along S_1 , and a variably overprinting S_2 fabric that also defines the axial plane of the fold; C) Dismembered garnet porphyroblasts in felsic granulite from the Cora Lake shear zone (Jeanotte Lake); E) Brecciated garnet porphyroblast within a felsic granulite from the Cora Lake shear zone.

Figure 5: Harker diagram for Neoproterozoic plutonic rocks from the east Athabasca mylonite triangle.

Green squares are the Fehr granite in the Chipman subdomain (Koteas et al., 2010) and red circles are new data from the northwestern subdomain plutonic rocks.

Figure 6: Frost diagrams for Neoproterozoic plutonic rocks from the east Athabasca mylonite triangle. Left: Fe-index vs SiO_2 ; right: Modified alkali lime index projection.

Figure 7: Tectonic discrimination for ca. 2.6 Ga plutonic rocks from the east Athabasca mylonite triangle (Pearce et al., 1984).

Figure 8: Spider diagrams for Neoproterozoic plutonic rocks from the east Athabasca mylonite triangle.

Top: incompatible element diagram normalized to primitive mantle (Sun and McDonough, 1989); bottom: REE plot normalized to chondrites (Sun and McDonough, 1989).

Figure 9: Spider and MALI diagram comparing results from this study to Andean volcanic rocks provided in Bryant et al. (2006).

Figure 10: Neodymium evolutionary diagram for Sm-Nd isotopic results from plutonic rocks in the northwestern subdomain presented in Hanmer et al. (1994). Inset diagram is an isochron calculated from published data. No access to raw data prohibited statistical analysis, thus no error could be calculated. Also plotted on the diagram are results from Angikuni Lake from Cousens et al. (1998).

Figure 11: Geochemical comparison between Neoarchean plutonic rocks within the east Athabasca mylonite triangle and Neoarchean meta-igneous rocks from Angikuni Lake (Aspler et al., 1999).

Figure 12: Geochronology results for sample 10W-094g. Notice results from the garnetite are on the left, and results from the leucosome are on the right.

Figure 13: A) Summary of monazite textures, ages, and pressure temperature conditions for sample 10W-094g; B) schematic interpretation of the monazite chemistry with respect to growth and consumption of major phases.

Figure 14: Monazite geochronology and chemistry results for sample 11G-020.

Figure 15: IN-SIMS results for sample 11G-020. A) Mg $k\alpha$ full thin section WDS map with leucosome outlined; B) BSE image of zircon in leucosome, black circle delineates approximate location of drilling, zircon are labeled; C) BSE image of Z1 with analytical spots located and labeled with age; D) schematic displaying the function of limiting the field aperture within the transfer case of the mass spectrometer, for details see Chamberlain et al. (2010); E) results for core analysis on Z4; F) Probability density plot of detrital zircon U-Pb analysis ($n = 207$); G) age vs. % concordance for the detrital zircon analysis, notice u-shape profile between 2.60 and 1.9 Ga indicative of open system behavior of zircon.

Figure 16: Monazite geochronology summary for all samples analyzed. Weighted means of ages of a certain population listed to right. Below histograms are the metamorphic conditions calculated for individual samples corresponding to the development of a strong S_1 fabric. It is important to note that igneous activity is quickly followed by crustal thickening and pervasive deformation/migmatization.

Figure 17: Schematic tectonic model for ca. 2.60 Ga arc magmatism along the southeastern margin of the Rae domain. As drawn, oceanic crust between oceanic – plateaus and infant-arcs of the Hearne domain and continental Rae crust is subducting beneath the Rae margin. Collision of Hearne lithosphere caused collisional orogenesis, and thrusting to the southeast (present coordinates) of Rae crust towards the Hearne domain. STZ is delineated on picture to show approximate location of the geophysically defined STZ relative to the continental margin.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12

Figure 13

Figure 14

Figure 16

Figure 17

Table 1: Structural nomenclature (from Regan et al., 2014)

Structural Generation	Domain	Orientation	Sense of Shear	P-T	Age	Refs
$S_{N1}/M_{N1}/D_{N1}$	Northwestern hangingwall	Subhorizontal	Top to the ESE	0.90 GPa; 750°C	2577 Ma	(1); this study
$S_{C1}/M_{C1}/D_{C1}$	Southeastern footwall	Subhorizontal	n/a	1.35 GPa; 850°C	2557 Ma	(2)
$S_{N2}/M_{N2}/D_{N2}$	Northwestern hangingwall	Steeply dipping to the NW	Dextral on shallowly plunging lineation	0.90 GPa; 750°C	1900 Ma	(1)(3); this study
$S_{C2}/M_{C2}/D_{C2}$	Southeastern footwall	Steeply dipping to the NW	Dextral, with increasing sinistral component towards CLsz	1.17 GPa; 800°C	1900 Ma	(4)(5); this study
$S_3/M_3/D_3$	Both (localized)	Moderately to steeply dipping to the NW	Sinistral on a moderate to shallowly plunging lineation	1.06-0.82 GPa; 800-700°C	1880-1870 Ma	This study

(1) Dumond et al. (2010); (2) Mahan et al. (2008); (3) Williams et al. (2000); (4) Williams et al. (1995); (5) Flowers et al. (2006a)

Table 2: Existing U-Th-Pb geochronology

Suite	Subdomain	Member	Age (Ma)	Reference	Method	Phase
Mary-Bohica	Northwestern	Mary Granite	2606+13/-11	Hanmer, 1997	Multi grain TIMS	Zircon
Mary-Bohica	Northwestern	Godrey Granite	2618+/-4	Hanmer, 1997	Multi grain TIMS	Zircon
Mary-Bohica	Northwestern	Brykea Granite	2601+/-4	Hanmer, 1997	Multi grain TIMS	Zircon
Mary-Bohica	Northwestern	Bradley Granite	2584+40/-15	Hanmer, 1997	Multi grain TIMS	Zircon
Mary-Bohica	Northwestern	Rea Granite	2604+/-1	Hanmer, 1997	Multi grain TIMS	Zircon
Mary-Bohica	Northwestern	Hawkes Granite	2610+/-1	Hanmer, 1997	Multi grain TIMS	Zircon
Mary-Bohica	Northwestern	Bohica mafic complex	2596+/-12 2620+34/-21	Hanmer, 1997	Multi grain TIMS	Zircon
Mary-Bohica	Nothwestern	Felsic Lake leucogranite	2629+26/-20	Hanmer, 1997	Multi grain TIMS	Zircon
McGillivray	Chipman	McGillivray	2621 +/-3	Hanmer, 1997	Multi grain TIMS	Zircon
Fehr	Chipman	Fehr Granite	2598+/-3	Hanmer, 1997	Multi grain TIMS	Zircon
Fehr	Chipman	Bompass Granite	Ca. 2600	Hanmer, 1997	Multi grain TIMS	Zircon
Chipman batholith	Chipman	Anorthosite	3149+/-100	Hanmer, 1997	Multi grain TIMS	Zircon
Axis	Upper deck	Mafic Granulite	Ca. 2600	Hanmer, 1997	Multi grain TIMS	Zircon
Gabbronorite	Upper deck	Clut granite	2614 +9/-7	Hanmer, 1997	Multi grain TIMS	Zircon
Clut granite	Upper deck	Melby-Turnbull granites	2610+/-2	Hanmer, 1997	Multi grain TIMS	Zircon
Mafic Granulite	Chipman	Mafic Granulite	2554.5+/-4.3*	Flowers et al., 2008	Single grain TIMS/SHRIMP	Zircon
Chipman Dikes- Leucosome	Chipman	Migmatitic Dike	1896.2+/-0.3 *	Flowers et al., 2006a	Single grain TIMS	Zircon
Axis	Upper deck	Mafic granulite	2550-2520	Baldwin et al., 2004	Single grain TIMS	Zircon
Gabbronorite	Upper deck	Godfrey Granite	2610	Baldwin et al., 2003	Single grain TIMS	Zircon
Mary	Upper deck	Beed Granite	2608+/- 9	This study	EPMA: in-situ	Monazite
Mary-Bohica	Northwestern	Mary Granite	2611+/- 30	This study	EPMA: in-situ	Monazite

*Ages interpreted to be metamorphic, or a minimum age of igneous cooling.

Table 3: Whole rock major and trace element geochemistry

Sample	10w- 110a	11R- 077	06G- 017	12R- 031	12R- 054	11R- 112	12R- 032 b	12R- 068	11R- 081	12R- 072a	10W- 094gb	10W- 094gc
SiO ₂	73.71	67.39	65.81	59.25	68.35	58.01	59.59	64.24	64.85	45.85	84.08	72.12
Al ₂ O ₃	13.20	14.59	15.42	14.97	15.24	16.58	14.34	15.12	14.79	15.14	9.09	13.06
Fe ₂ O ₃	2.46	5.71	5.21	10.56	4.15	9.42	10.45	8.73	7.49	12.31	1.14	4.76
MgO	0.60	0.63	2.92	1.64	1.39	1.25	2.13	0.72	1.04	9.01	0.97	3.34
CaO	1.86	2.69	2.31	4.25	2.51	5.92	4.64	3.73	3.84	10.94	1.30	1.74
Na ₂ O	2.73	2.73	2.87	3.28	3.21	3.55	3.14	3.29	3.08	2.05	1.98	2.20
K ₂ O	4.31	4.58	3.27	3.10	3.40	2.68	2.34	2.98	3.41	1.70	1.05	1.65
TiO ₂	0.28	0.55	0.48	1.28	0.50	1.30	1.34	0.84	0.80	0.94	0.19	0.53
P ₂ O ₅	0.05	0.16	0.07	0.43	0.19	0.64	0.50	0.21	0.25	0.33	0.02	0.03
MnO	0.02	0.07	0.07	0.14	0.07	0.13	0.14	0.11	0.10	0.18	0.03	0.08
Cr ₂ O ₃	0.002	0.003	0.020	0.003	0.005	0.002	0.005	0.002	0.003	0.026		
Total	99.22	99.10	98.45	98.90	99.02	99.48	98.62	99.97	99.65	98.48	99.85	99.51
Cs	0.1	0.9	2.5	0.9	0.3	0.9	0.4	0.2	0.2	0.6		
Rb	53.7	138.6	46.9	51.2	83.9	36.0	32.2	47.9	58.0	27.0		
Sr	312	238	351	353	295	585	353	375	324	870		
Ba	1149	1203	947	1383	731	1380	1184	1547	1442	401		
Ta	0.1	0.9	0.3	0.9	0.5	0.9	1.0	0.6	0.6	0.9		
Nb	1.95	11.33	5.85	16.07	10.45	15.12	15.80	9.97	10.23	40.17		
Pb	22.12	37.28	17.89	19.82	28.16	19.38	17.93	18.91	19.90	12.59		
Hf	2.63	8.27	1.99	7.88	4.18	2.16	2.83	4.43	3.77	0.80		
Zr	96.8	323.0	72.8	321.7	146.5	72.0	123.4	203.7	160.8	29.7		
Y	4.6	27.5	9.3	38.3	15.0	34.3	33.6	23.9	23.6	15.4		
Sc	3.1	11.5	11.0	19.0	9.1	17.9	21.6	17.6	14.6	21.4		
Ni	4.9	6.3	63.6	9.3	14.2	4.2	15.9	4.9	5.6	42.4		
Zn	32.2	80.2	80.7	134.7	46.5	150.3	126.0	125.4	89.8	174.0		
V	52	26	96	73	54	46	95	23	57	212		
Cu	4.17	12.40	30.54	12.77	8.38	8.15	20.93	8.24	11.42	40.55		
Th	22.3	7.8	5.1	3.2	28.6	3.4	1.2	0.7	0.8	1.9		
U	1.0	1.5	0.5	1.2	2.2	1.1	0.5	0.4	0.4	0.4		
Ga	15.04	20.78	18.55	22.48	16.93	25.40	20.47	23.24	20.88	16.37		
La	154.58	84.43	51.77	98.31	118.74	104.35	79.63	48.17	58.16	91.72		
Ce	19.4	10.2	6.7	14.0	16.0	15.9	12.1	7.4	8.2	16.0		
Pr	55.4	34.6	21.0	47.7	49.3	53.8	42.4	27.0	29.5	57.0		
Nd	7.1	6.4	3.6	9.7	9.1	10.7	8.3	5.8	5.7	10.3		
Sm	1.5	2.1	0.9	2.9	1.1	4.3	3.0	3.3	2.4	3.0		
Eu	2.3	5.6	2.1	7.6	5.0	8.3	7.7	5.1	4.9	5.9		
Gd	0.4	0.9	0.4	1.5	0.9	1.4	1.3	0.8	0.9	0.8		
Tb	1.0	4.7	1.6	6.7	3.3	6.6	6.5	3.9	4.1	3.1		
Dy	0.2	1.1	0.4	1.6	0.6	1.5	1.4	1.0	1.1	0.6		
Ho	0.3	2.9	1.1	3.7	1.3	3.5	3.3	2.7	2.3	1.3		
Er	<0.1	0.5	0.1	0.6	0.2	0.6	0.5	0.4	0.4	0.2		
Tm	0.4	2.5	1.1	3.5	1.2	2.8	2.9	2.6	2.3	1.1		
Yb	<0.1	0.4	0.1	0.6	0.2	0.6	0.5	0.4	0.4	0.2		
Lu	154.58	84.43	51.77	98.31	118.74	104.35	79.63	48.17	58.16	91.72		

Table 4: EPMA monazite results

Sample	domain	UO2	ThO2	CaO	SO3	P2O5	Y2O3	SiO2	PbO	Age (Ma)	2 σ error
12R-054	m13_outercore1	0.547581	6.41353	1.27987	0.021327	29.67768	1.97982	0.404157	1.000564	2584	10
12R-054	m12_innercore1	0.612756	4.797385	1.190083	0.014745	30.17553	2.077418	0.158628	0.85679	2615	11
12R-054	m12_outercore1	0.584783	7.014313	1.610588	0.019481	29.84515	1.299607	0.240102	1.09768	2604	11
12R-054	m12_innerrim1	0.55805	6.592563	1.539185	0.021241	29.64473	1.574027	0.235433	1.037067	2607	9
12R-054	m13_innerrim1	0.562919	6.696695	1.521753	0.028107	29.8183	1.60549	0.262125	1.038743	2579	13
12R-054	m1_outercore1	0.206373	6.621407	0.839065	0.033765	28.80868	1.492458	0.870019	0.874166	2612	9
12R-054	m15_outercore1	0.594523	6.300905	1.419155	0.020117	30.30373	1.780353	0.261107	1.001867	2566	23
10W-094g_1	m1core1	0.134081	7.658608	1.165067	0.016957	29.74308	0.045535	0.845461	0.94851	2580	8
10W-094g_1	m2_outercore1	0.435077	2.587753	0.665859	0.011185	30.96018	1.563933	0.140128	0.508728	2599	14
10W-094g_1	m2_innercore1	0.373545	4.806685	1.17362	0.024574	30.7907	0.597151	0.147552	0.860286	2961	12
10W-094g_1	m2_innerrim1	0.288526	7.93177	1.20468	0.018629	30.12584	0.035685	0.961841	1.042439	2558	8
10W-094g_1	m3outercore1	0.50075	2.909715	0.797825	0.014791	31.08667	0.459473	0.140914	0.749698	3189	12
10W-094g_1	m3innercore1	0.457113	3.25613	0.865644	0.024015	30.63233	2.274542	0.125786	0.739238	3087	12
10W-094g_1	m4_upperrim1	0.639305	3.523196	1.025977	0.016525	30.69696	0.329312	0.096847	0.71081	2588	10
10W-094g_1	m4_innercore1	0.408152	3.192088	0.793082	0.016382	30.26618	1.097682	0.186186	0.721024	3159	14
10W-094g_1	m4lowerrim1	0.55809	2.939597	0.833398	0.026662	30.44608	0.420638	0.095094	0.603028	2582	10
10W-094g_1	m1rimlower1	0.176183	5.481374	0.9383	0.014798	29.40696	0.04891	0.509413	0.715086	2575	10
10W-094g_1	m1rcore2	0.099	8.587853	1.186718	0.009506	28.43983	0.029289	1.043552	1.028765	2555	8
10W-094g_1	m2innerrim2	0.308284	7.91644	1.20702	0.016628	29.6507	0.046561	0.97593	1.024695	2501	10
10W-094g_1	m6outerrim2	0.121935	3.139412	0.611526	0.010838	30.66504	0.076871	0.253973	0.399367	2530	18
10W-094g_1	m3outercore2	0.509332	2.964578	0.823992	0.019727	30.92963	0.437214	0.132564	0.739236	3114	14
10W-094g_1	m2innerrim3	0.286151	7.05663	1.119098	0.013378	29.81694	0.038573	0.812095	0.928382	2525	8
10W-094g_1	m7uppercore1	0.863397	3.692984	0.983073	0.016603	30.97354	2.601848	0.168465	0.82129	2552	8
10W-094g_1	m3_innerrim1	0.285916	2.795598	0.673709	0.011912	30.968	1.503593	0.134233	0.462452	2602	16
10W-094g_1	m2coreright2	0.664399	3.296285	0.934656	0.014501	30.49918	0.82367	0.17605	0.835789	2983	11
10W-094g_2	m9innerrim2	0.097437	2.31658	0.476759	0.009326	30.5217	0.041853	0.205318	0.311621	2564	22
10W-094g_2	m9_innercore1	0.566435	3.053693	0.892768	0.020351	30.44255	0.305176	0.081714	0.615337	2564	12
S823b	m2outercoreleft	0.406629	5.885394	0.174794	0.010788	27.9505	0.866943	1.549864	0.882983	2609	9
S823b	m3_core1	0.238565	9.883168	0.449991	0.017675	26.55878	0.968649	2.163833	1.275293	2618	8
S823b	m1_core1	0.754087	16.67246	0.184669	0.025365	22.60608	0.824671	4.41261	2.334636	2630	4
S823b	m1_rim1	0.602138	11.21105	0.236883	0.022871	25.66428	0.86355	2.910123	1.596798	2605	6
S823b	m1_core2	0.816215	18.44202	0.186796	0.022849	22.05058	0.92011	4.97252	2.533614	2596	4
S823b	m1_outerrim1	0.598904	14.0583	0.578352	0.019197	23.92173	0.777936	3.587325	1.901863	2573	5
11G-020	M2_core1	0.719813	1.412396	0.5425734	0.009846	29.87478	0.2242562	0.06433	0.3479488	1923	11

Table 4 continued

Sample	domain	UO2	ThO2	CaO	SO3	P2O5	Y2O3	SiO2	PbO	Age (Ma)	2 σ error
11G-020	M2_rimright1	0.1650775	3.473032	0.585702	0.015624	29.6002	0.050312	0.362035	0.334158	1857	11
11G-020	M2_core2	0.7287914	1.535068	0.562976	0.013455	29.49804	0.222298	0.052487	0.362872	1931	10
11G-020	M1_coreleft2	0.128466	14.94108	0.606166	0.025626	24.67938	0.238695	3.138782	1.812014	2612	6
11G-020	M1_innerrimtop1	0.140063667	7.942373	1.367233	0.015675	29.19677	0.062913	0.602147	0.719647	1928	8
11G-020	M12_core1	0.4498842	3.221576	0.856966	0.028751	30.17856	0.358964	0.144079	0.416562	1927	10
11G-020	M11_innercore1	0.219029333	3.615095	0.741518	0.013819	29.91722	0.048452	0.22107	0.371823	1904	10
11G-020	M11_outerrim1	0.145799	4.090962	0.736061	0.015693	29.60058	0.038689	0.346074	0.387278	1897	10
11G-020	M9_core1	0.4005874	4.008648	0.865637	0.021055	29.5868	0.104205	0.242012	0.458461	1893	8
11G-020	M13_innercore1	0.1961408	4.97197	0.961578	0.008501	29.38906	0.057894	0.268335	0.476491	1896	9
11G-020	M10_core1	0.317049333	3.826435	0.805021	0.020557	29.33053	0.066944	0.208983	0.420609	1905	9
11G-020	M9_innerrim2	0.2320605	5.653855	0.960426	0.008727	29.0903	0.059508	0.482391	0.545565	1899	9
11G-020	M9_outerrim1	0.161425333	2.61638	0.556675	0.004552	29.56817	0.071583	0.202072	0.269795	1903	18
11G-020	M11_innerrim1	0.184998167	4.309753	0.814206	0.0198	29.31377	0.037049	0.372458	0.419921	1906	9
11G-020	M11_outerrim2	0.143645667	3.195123	0.717971	0.021468	30.0452	0.039949	0.997355	0.307411	1860	16
11G-020	M1_innercorebot1	0.14982475	11.47347	0.714931	0.028505	26.26053	0.322811	2.05939	1.392659	2571	6
11G-020	M1_innerrimbot1	0.164624667	4.403887	0.763438	0.018386	29.4906	0.046534	0.35666	0.412958	1867	10
11G-020	M1_core1	0.1809156	4.677422	0.840856	0.01132	29.0206	0.051579	0.385763	0.448686	1903	9
11G-020	M1_coreright1	0.3250214	17.101	1.67984	0.025454	24.8839	1.36535	2.74979	2.118582	2566	4
11G-020	M1_coreleft1	0.05453625	9.593608	0.528855	0.014224	26.58345	0.008854	1.914223	1.154835	2623	8
11G-020	M1_rimtop1	0.1708438	4.990962	0.918861	0.016876	29.62394	0.073769	0.338146	0.470317	1896	9
11G-020	M1_outercorebot	0.09379675	8.277918	0.747883	0.016056	27.70868	0.032727	1.366993	1.015102	2615	9
11G-020	M1_coreright2	0.1681576	13.1723	0.897945	0.017472	25.83044	0.595402	2.433324	1.551626	2505	6
11G-020	M1_coreright3	0.1494452	14.33422	0.642483	0.023467	25.03518	0.394828	2.981992	1.706888	2552	5
11G-020	M1_coreleft3	0.0764656	8.500496	0.745599	0.014768	27.72838	0.042293	1.391938	1.037116	2622	8
11G-020	M1_coretop1	0.1139534	13.66222	0.902159	0.025006	25.94732	0.089871	2.497754	1.639022	2589	6
11G-020	M1_corebotright1	0.1319625	13.3144	0.627442	0.019605	25.29313	0.247349	2.751598	1.588455	2560	6
11G-020	M9_innercore1	0.05596425	8.338933	0.60596	0.021608	27.09978	0.014212	1.490505	0.989447	2577	9
11G-020	M1_coreright4	0.2635496	15.57768	1.238056	0.026719	25.20904	0.952557	2.701682	1.887624	2532	5

Table 5 IN-SIMS geochronologic results

	Isotope ratios						Age (Ma)					
	$^{207}\text{Pb}/^{206}\text{Pb}$	2σ	$^{207}\text{Pb}/^{235}\text{U}$	2σ	$^{206}\text{Pb}/^{238}\text{U}$	2σ	$^{207}\text{Pb}/^{206}\text{Pb}$	2σ	$^{206}\text{Pb}/^{238}\text{U}$	2σ	$^{207}\text{Pb}/^{235}\text{U}$	2σ
11G-020_z1_1	0.1754	0.0003	8.80	0.77	0.3636	0.0319	2610	3	1999	151	2317	80
11G-020_z1_2	0.1752	0.0008	8.56	0.85	0.3544	0.0353	2608	8	1955	168	2292	91
11G-020_z2	0.1645	0.0008	8.15	0.71	0.3592	0.0314	2502	8	1979	149	2247	79
11G-020_z4	0.1842	0.0006	8.45	0.67	0.3328	0.0262	2691	5	1852	127	2281	72
11G-020_z6_1	0.1952	0.0018	9.48	0.63	0.3521	0.0235	2787	16	1944	112	2385	61
11G-020_z6_2	0.2001	0.0014	11.25	1.07	0.4078	0.0390	2827	11	2205	179	2544	88

Notes: $^{207}\text{Pb}/^{206}\text{Pb}$ ages used for weighted means

Table 5: Existing Pressure-Temperature calculations in the study area

Subdomain	Pressure (GPa)	Temperature (°C)	Proposed timing (Ga)	Rock Type	Reference
Chipman	1.17	850	1.9	Chipman dike	Williams et al., 1995
Chipman	1.17	850	1.9	Chipman dike	Mahan and Williams, 2005
Chipman	1.06	800	<1.9	2.6 Ga Mafic Granulite	Mahan et al., 2008
Chipman	1.35	850	2.56	2.6 Ga Mafic Granulite	Mahan et al., 2008
Chipman	1.17	850	1.9	Chipman dike	Regan et al., 2014
Chipman	1.17	850	1.9	Chipman dike	Martel et al., 2008
Chipman (CLsz)	1.06	800	1.89	Mafic granulite	Regan et al., 2014
Chipman (CLsz)	0.9	775	1.88	3.2 Ga Anorthosite	Regan et al., 2014
Chipman (CLsz)	0.8	750	1.88	Mafic Granulite	Regan et al., 2014
Northwestern (CLsz)	0.8	775	1.88	Felsic Granulite	Regan et al., 2014
Northwestern	0.9	750	1.9	Mary Granodiorite	Regan et al., 2014
Northwestern	0.9	750	2.56	Mary Granodiorite	Regan et al., 2014
Northwestern	0.9	750	2.56	Mary Granite	Dumond et al., 2010
Northwestern	0.9	750	1.9	Mary granodiorite	Williams et al., 2000
Northwestern	1.0	750	1.9	Mary Granodiorite	Williams et al., 2000
Northwestern	1.05	725	2.56	Beed Granite	<i>This study</i>
Chipman	1.35	850	2.56	Felsic Granulite	<i>This study</i>
Chipman	1.17	815	1.9	Felsic Granulite	<i>This study</i>

* This study: see supplement (Regan, 2016)