Deformation and extensional exhumation of 1.9 Ga high-pressure granulites along the Wholdaia Lake shear zone, south Rae craton, Northwest Territories, Canada

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ABSTRACT

The origin of high-pressure granulites in the south Rae craton and Snowbird tectonic zone (STZ) is highly enigmatic. Current models for their formation and exhumation envisage continental collision at 2.55 Ga and intracratonic orogenesis at 1.9 Ga, or collision and exhumation at ca. 1.9 Ga. As an attempt to reconcile these disparate models, we conducted a regional and detailed mapping program along a geophysical discontinuity 100 km west of the STZ within the south Rae craton of the Northwest Territories, Canada. This work presents the discovery of a new crustal-scale shear zone, the Wholdaia Lake shear zone (WLsz), which deformed and transposed host rocks into a 20-km-wide and 300-km-long ductile high-strain zone. U-Pb zircon geochronology was utilized to establish host-rock crystallization ages, timing of deposition of metasedimentary rocks, and age constraints of metamorphism and ductile shearing. Hanging-wall metasedimentary rocks have a new depositional range of 1.98–1.93 Ga and contain abundant metamorphic zircon at 1.91 Ga. The protoliths of the footwall mafic granulite orthogneisses crystallized at 2.6 Ga and were metamorphosed at 1.9 Ga, which extends the known footprint of 1.9 Ga metamorphism 100 km west of the STZ. During and after 1.9 Ga metamorphism, the WLsz began progressively exhuming footwall rocks in three distinct stages, associated with (1) normal-oblique shearing at high-pressure granulite-facies conditions, (2) normal-oblique shearing accompanied by mylonitization at amphibolite-facies conditions, and (3) normal-oblique shearing with ultramyxolite development at amphibolite- to greenschist-facies conditions. Ductile shearing was waning by 1.86 Ga, based on ages obtained from late syn- to postkinematic crosscutting dikes. Collectively, the WLsz in concert with other regional structures aided both extensional and thrust-sense exhumation of a large high-grade terrane at 1.9 Ga in the south Rae craton.

INTRODUCTION

Deeply eroded ancient orogens provide insights into the formation and exhumation of stable continental masses involved in supercontinent amalgamation. The Rae craton of the western Churchill Province (Fig. 1), northern Canada, is thought to be the locus of amalgamation for the Paleoproterozoic supercontinent Nuna (Hoffman, 2014), yet the current paucity of regional data means its Proterozoic assembly is poorly understood. Several large high-pressure granulite domains occur along the Rae craton’s eastern margin (e.g., Sanborn-Barrie et al., 2001; Baldwin et al., 2007; Berman et al., 2007; Martel et al., 2008) within the Snowbird tectonic zone (STZ), which separates these rocks from the lower-grade Archaean Hearne craton. In southeasternmost Northwest Territories (NWT) and northern Saskatchewan, high-grade rocks of the Rae craton are particularly well exposed and preserve teconothermal events at 2.55 Ga (Baldwin et al., 2006; Mahan et al., 2006a), 2.5–2.3 Ga (Berman et al., 2013), 1.92 Ga (Martel et al., 2008; Bethune et al., 2013), and 1.90 Ga (Baldwin et al., 2004; Berman et al., 2007), followed by exhumation and tectonic stabilization of the crust by 1.75–1.70 Ga (Rainbird et al., 2005; Pett et al., 2014). A lack of overall understanding of the nature and significance of 2.55 Ga and 1.9 Ga high-pressure metamorphism in the southeastern Rae craton results in competing tectonic models that appear incompatible. Much of the high-pressure granulite-facies metamorphism along the STZ occurred at 1.9 Ga (Fig. 1) and is thought to relate to collision with the Hearne craton (Berman et al., 2007), which produced eclogites (Baldwin et al., 2004, 2007) and buried Paleoproterozoic metasedimentary rocks (Martel et al., 2008; Bethune et al., 2013). Alternative models suggest that local occurrences of high-pressure granulite-facies rocks along the STZ that have been dated at 2.55 Ga (Davis et al., 2006; Baldwin et al., 2006; Mahan et al., 2006a; Mills et al., 2007; Flowers et al., 2008; Dumond et al., 2015, 2017) indicate a Neoarchean collision between the Rae and Hearne cratons. In this scenario, high-pressure rocks deeply buried in Neoarchean time resided in the deep crust until a thermal pulse accompanied by shearing resulted in high-pressure metamorphism and exhumation at 1.9 Ga (Flowers et al., 2006a; 2008; Dumond et al., 2015).

In order to better understand the nature of these high-grade metamorphic rocks and their tectonometamorphic evolution, we undertook a combined study of the crustal-scale shear zones, deformed supracrustal assemblages, and deeply exhumed crust within the southeastern Rae craton. Herein, we document detailed structural, metamorphic, and
The Paleoproterozoic cover and Archean substrate were subsequently metamorphosed into mafic plutons (Pehrsson et al., 2013; Davis et al., 2015; Regis et al., 2017b). Various younger ca. 2.17–1.95 Ga Paleoproterozoic metasedimentary packages (Bostock and van Breeman, 1994; McDonough and McDonough, 1993; Martel et al., 2008; Ashton et al., 2013, 2017a; Shiels et al., 2016; Ply, 2016; Thiessen et al., 2017) overlie the basement rocks. The Paleoproterozoic cover and Archean substrate were subsequently variably deformed and intruded by 1.8–1.75 Ga magmatic suites (Peterson et al., 2002).

Bounding the south Rae craton to the west, there is the 1.99–1.91 Ga Taltson magmatic zone (Fig. 1), an orogen that contains high-grade Rae craton basement gneisses intruded by voluminous I-type (1.99–1.95 Ga) and S-type (1.94–1.92 Ga) plutonic rocks. Peak metamorphism in the Taltson magmatic zone occurred at 1.94–1.93 Ga (Henderson et al., 1990; McDonough et al., 2000; McNicoll et al., 2000). Recent work has also recognized that the western margin of the south Rae craton was affected by a significant pre-Taltson orogenic event, the 2.5–2.3 Ga Arrowsmith orogeny (Berman et al., 2013; Bethune et al., 2013). Following Arrow-Smith and Taltson magmatic zone orogenesis, the south Rae craton was affected by 1.9 Ga Snowbird (e.g., Berman et al., 2007) and ca. 1.85–1.79 Ga Trans-Hudson orogenesis (Corrigan, 2012). The latter overlapped in age with deposition of unmetamorphosed volcano-sedimentary rocks of the Baker Lake Group in a series of transtensional basins (Fig. 1; Rainbird et al., 2006; Hadliari and Rainbird, 2011).

Separating the south Rae craton from the Hearne craton to the east, there is the STZ, a >2000-km-long major crustal structure in the Canadian Shield that contains a network of northeast-trending shear zones, transposed host-rock panels, and late faults active between 1.9 and 1.75 Ga (Mahan et al., 2003; Mahan and Williams, 2005; Flowers et al., 2006b; Berman et al., 2007; Martel et al., 2008; Card, 2016). The latest ductile deformation on the STZ is constrained to be 1.85 Ga (Mahan et al., 2006a), although brittle-ductile faulting occurred as late as ca. 1.75 Ga (Rainbird et al., 2005; Davis et al., 2015).

**South Rae Craton Lithotectonic Domains**

A recent geochronological reconnaissance transect, which was aided by a high-resolution aeromagnetic survey (Kiss and Coyle, 2012), was undertaken across the largely unmapped south Rae craton in NWT. The transect resulted in the delineation of northeast-trending lithotectonic domains (Davis et al., 2015; Pehrsson et al., 2015; Percival et al., 2016; Regis et al., 2017b) that correlate well with similar domains recognized in Saskatchewan (e.g., Ashton et al., 2007).

The Tantato domain of northern Saskatchewan (Fig. 2) is underlain by abundant 2.6–2.55 Ga mafic to felsic orthogneiss (e.g., Hamer, 1997; Baldwin et al., 2003, 2006) and paragneiss (Hamer, 1997; Dumond et al., 2015). Along its eastern margin adjacent to the Hearne craton, the shear-bounded Chipman subdomain (Hamer et al., 1994; Mahan et al., 2006b) is composed of a distinct ca. 3.4 Ga tonalite batholith intruded by the 2.1 Ga Chipman dike swarm (Regan et al., 2016). A separate shear-bounded block of 3.4 Ga tonalite in NWT (Martel et al., 2008) is correlative with the Chipman subdomain in Saskatchewan. Their offset is due to dextral strike-slip displacement along the Grease River shear zone (Fig. 2), which was active between 1.9 and 1.8 Ga (Lafrance and Sibbald, 1997; Mahan and Williams, 2005; Dumond et al., 2008). To the west and east of the Chipman domain in NWT, there are two high-strain zones (Fig. 2), called the Striding mylonite zone and the Chipman shear zone (Regan et al., 2014) and Legs Lake shear zone in Saskatchewan (Mahan et al., 2003). The Chipman and Legs Lake shear zones form the surface trace of the STZ in this area (Fig. 2) and are interpreted to have accommodated exhumation of the Chipman panel to midcrustal levels via east-vergent thrusting over the Hearne by ca. 1.85 Ga (Mahan et al., 2006b). Similar to the strain documented on the Striding mylonite zone, the Cora Lake shear zone exhibits normal-oblique sinistral shear sense (Chipman domain up to the northeast) at ca. 1.89–1.87 Ga (Regan et al., 2014). North of Kasba Lake, an undeformed Nuelin intrusion cuts the...
West of the Snowbird-Dodge domains, separated by the Ryckman Bay shear zone (Ashton and Card, 1998) and the WLSz, there are the correlative Train (Saskatchewan) and Firedrake (NWT) domains (Fig. 2). These domains contain mafic-intermediate orthogneisses and minor paragneisses that are 2.7–2.57 Ga and abundant migmatitic granite that is ca. 1.85–1.81 Ga (Ashton et al., 1999, 2009; Davis et al., 2015; Regis et al., 2017b). The Firedrake domain is also characterized by magnetic anomalies that are tens-of-kilometer scale; curvilinear refolded patterns are high-intensity magnetic patterns (red colors in Fig. 3; Kiss and Coyle, 2012) that underlie the characteristic magnetic pattern of high and low magnetic anomalies (Fig. 3) that highlight kilometer-scale type-2 fold interference patterns (Fig. 3; e.g., Ramsay, 1967; Martel et al., 2008). Orthogneissic and metaplutonic rocks (clinopyroxene [Cpx] ± orthopyroxene [Opx]–bearing; mineral abbreviations according to Whitney and Evans, 2010) that underlie the high-intensity magnetic patterns (red colors in Fig. 3B) have ca. 2.73–2.70, 2.66, 2.63, and 2.54 Ga crystallization ages (Martel et al., 2008; Regis et al., 2017b). The low-intensity magnetic patterns (blue colors) are typically underlain by pelitic-psammitic to calc-silicate metasedimentary rocks with a maximum depositional age of 2.07 Ga (Martel et al., 2008). In the western Snowbird domain, metasedimentary rocks originally inferred to be Archean in age (Martel and Pierce, 2006) were recently determined to have Paleoproterozoic detrital zircon ages (Thiessen et al., 2017). These units are tentatively correlated with the Rae cover sequence (Martel et al., 2008; Rainbird et al., 2010) and the upper Murmac Bay Group (Fig. 1) to the southwest (Ashton et al., 2013). The Snowbird domain metasedimentary rocks and basement gneisses are unconformably overlain by undeformed, unmetamorphosed ca. 1.83 Ga Christopher Island Formation volcaniclastic rocks (Figs. 1 and 3) of the Baker Lake Group on Snowbird Lake (e.g., Roscoe and Miller, 1986; Peterson et al., 2002), indicating that the final exhumation of the Snowbird domain by this time.

At the latitude of Wholdaia Lake, high-pressure rocks were originally thought to be confined to an oblong domain of low-resolution magnetic anomalies (Dods et al., 1987) termed the Selwyn lozenge of the STZ (Hammer et al., 1994). The western margin of the lozenge was investigated by Hammer (1997) and Krikorian (2002), and although a distinct horizontal gravity gradient anomaly is also apparent along the boundary (Sharpton et al., 1987), and high-strain zones were recognized, they concluded the structure to be discontinuous and to have accommodated discrete shearing within a singular larger tectonic block. Ages for late-kinematic granites and metamorphic monazite suggested ca. 1.9 Ga high-grade metamorphism was associated with this deformation (Hammer, 1997; Krikorian, 2002).

When analyzed more regionally, a broad zone of rectilinear, northeast-trending, low aeromagnetic character (Fig. 3; Kiss and Coyle, 2012) occurs for 300 km from southwest of Angikuni Lake to the Saskatchewan border (Fig. 1). Reconnaissance mapping by helicopter focused along this discontinuity (Davis et al., 2015) found that it separates the more highly magnetic and geologically distinct Snowbird and Firedrake domains; is coincident with transposed, tectonized, and mylonitized units along its length; and importantly contains scattered high-pressure assemblages that are also noted west of the break. Our research into the nature of this structure was focused on its best exposures along an ~100 km segment in the vicinity of Wholdaia Lake and is discussed in detail in the following section.

West of the Snowbird-Dodge domains, separated by the Ryckman Bay shear zone (Ashton and Card, 1998) and the WLSz, there are the correlative Train (Saskatchewan) and Firedrake (NWT) domains (Fig. 2). These domains contain mafic-intermediate orthogneisses and minor paragneisses that are 2.7–2.57 Ga and abundant migmatitic granite that is ca. 1.85–1.81 Ga (Ashton et al., 1999, 2009; Davis et al., 2015; Regis et al., 2017b). The Firedrake domain is also characterized by magnetic anomalies that are relatively high and tens-of-kilometer-scale curvilinear refolded patterns and kilometer-scale shear folds (Fig. 3).

Previous Pressure-Temperature-Time Work in the Southeastern Rae Craton

The bulk of the research on the pressure-temperature-time (P-T-t) conditions of south Rae craton metamorphism has come from the subdomains
Figure 3. (A) Wholdaia Lake shear zone (WLsz) geology highlighting high-grade ($S^r_1$; Grt + Cpx) and lower-grade ($S^l_2$; Hbl + Bt) domains, with mineral abbreviations according to Whitney and Evans (2010). Yellow circles are newly dated mafic granulites, the dark-blue circle is a newly dated metasedimentary sample, and the white circles are newly dated felsic dike samples. Pink polygon within the WLsz represents a weakly deformed 1.9 Ga syenite (sy) body. Reference lakes are: SbL—Snowbird Lake, WL—Wholdaia Lake, and SL—Selwyn Lake. Numbered annotations (e.g., 4bcd) refer to corresponding photographs in Figures 4 and 5. (B) Total field magnetic map (Kiss and Coyle, 2012) of the WLsz highlighting geological boundaries (black lines) and samples from Figure 3A. Higher-intensity magnetic signatures are red, lower-intensity signatures are blue. "EB" refers to easterly bends of the WLsz and CIF refers to the Cristopher Island Formation, as discussed in text.
mylonitic paragneiss. Abundant folded and sheared granitic dikes crosscut felsic orthogneiss and mylonite to ultramylonite, metagabbro, and local 2). The WLsz is composed of amphibolite- to granulite-facies, mafic to felsic orthogneiss with relict Cpx + Opx, which occur adjacent to the Grt + Cpx + Opx layers (Hbl + Qz + Pl + Bt). The younger than 2.07 Ga Snowbird domain metasedimentary rocks yield peak conditions of 0.76–0.9 GPa and 800–840 °C at 1.92–1.90 Ga (U-Pb zircon; Heaman et al., 1999).

The tectonic significance of the 2.55 and 1.9 Ga events has been extensively debated (Baldwin et al., 2004, 2006; Berman et al., 2007; Flowers et al., 2006a, 2008; Mahan et al., 2006a, 2006b, 2008; Dumond et al., 2010, 2015; Regan et al., 2014). The Chipman subdomain in Saskatchewan and NWT also shows pressures of 1.1 GPa and 795 °C at ca. 1.9 Ga (Mahan and Williams, 2005; Martel et al., 2008; Thiessen et al., 2017). The younger than 2.03 Ga Grollier Lake metasedimentary rocks (Ashton et al., 2017a) and upper Murmac Bay Group (Fig. 1) also exhibit significant 1.94–1.90 Ga metamorphism at grades up to granulite facies (Knox et al., 2011; Bethune et al., 2013; Ply, 2016), and in the far west, the potentially correlative Rutledge River basin units were deformed and metamorphosed to lower-granulite-facies conditions during foreland basin closure at ca. 2.10 Ga prior to 1.99–1.91 Ga plutonism within the Tallon magmatic zone (Bostock and van Breeman, 1994).

The Train domain, which lacks a younger sedimentary record, contains widespread (Grt + Hbl)–bearing orthogneiss with relict Cpx + Opx assemblages that have been heavily retrogressed and migmatized by injection of anatectic melt (Ashton et al., 1999; Heaman et al., 1999; Davis et al., 2015). Estimates for the timing of high-grade (Grt + Cpx, Grt-Hbl) metamorphism are ca. 1.81 Ga (U-Pb zircon; Heaman et al., 1999).

The tectonic significance of the 2.55 and 1.9 Ga events has been extensively debated (Baldwin et al., 2004, 2006; Berman et al., 2007; Flowers et al., 2006a, 2008; Mahan et al., 2006a, 2006b, 2008; Martel et al., 2008; Dumond et al., 2010, 2015; Regan et al., 2014). While the P-T conditions of the Paleoproterozoic metasedimentary sequences clearly require renewed crustal loading in the Proterozoic (e.g., Martel et al., 2008; Bethune et al., 2013), the extent to which the high-pressure events signify Proterozoic or Archean collisional orogeny is controversial. Eclogite-facies basaltic sills mapped across the southern Tantato domain (Knox and Lamming, 2015) record 1.9 Ga high-pressure (1.6 GPa, 750 °C) tectonometamorphism (Baldwin et al., 2004, 2007), but they have been also argued to have formed at ca. 2.55 Ga, based on the metamorphic record of their host metasedimentary rocks (Dumond et al., 2015, 2017).

**GEOLGY OF THE WHOLDAIA LAKE SHEAR ZONE**

New Mapping

The following description of the WLsz and its adjacent wall rocks comes from boat, helicopter, and foot traverses conducted during this study. Regional geological and predictive mapping (magnetic character) suggests the WLsz is at least 300 km long (Fig. 2), although it is best constrained with good exposures for ~100 km along strike in the vicinity of Wholdaia Lake. At Wholdaia Lake, the WLsz is roughly 20 km wide and has a foliation that strikes consistently at 030°. South and north of this central segment, the foliation changes to 070°, and the width of penetrative deformation is restricted to <10 km (easterly bends [EB] in Fig. 2). The WLsz is composed of amphibolite- to granulite-facies, mafic to felsic orthogneiss and mylonite to ultramylonite, metagabbro, and local mylonitic paragneiss. Abundant folded and sheared granitic dikes crosscut the WLsz (Fig. 3), as do late east-west-trending weakly deformed lamprophyre dikes.

D1–S1

The central WLsz is a composite structure with two distinct lithotectonic and structural domains, both of which trend northeast and dip southeast (Fig. 3). The earlier, high-grade fabric, S1, consists of (Grt + Cpx + Opx)–bearing mafic and felsic granulite gneiss and mylonite that typically display centimeter-scale annealed gneiss banding. In contrast to the polydeformed, variably oriented structures that characterize the Firedrake and Snowbird domains, S1 is consistently aligned into a northeast-striking, southeast-dipping (~60°) transposition foliation with a rarely preserved shallowly southwest-plunging lineation (10°–30°).

A gneissic monzogranite and associated mylonite (red unit in Fig. 3A) occur as a distinct ~800-m-wide band of highly magnetic, highly sheared rocks within the easternmost margin of the central WLsz (Fig. 3B). To the south, the magnetic signature of this unit is traceable for over 50 km, but farther to the north and south, its signature becomes more cryptic. Additionally, the gneiss outlines a kilometer-scale z-fold indicating dextral folding of these gneissic and mylonitic units. The gneiss contains banded leucocratic layers (Hbl + Pl + Qz + Opx + Cpx + Grt) and mesocratic layers (Hbl + Qz + Pl + Bt).

Mafic units of S1 that occur west of the monzogranite gneiss have motted to weakly linear magnetic patterns (Fig. 3) that commonly display higher magnetic signatures. Although these units occur predominantly along the easternmost WLsz, such occurrences are also found on the western margin adjacent to the Firedrake domain. Rocks in this unit consist of meter-scale amphibolite to granulite pods (Fig. 4A) injected by centimeter-scale, northeast-trending anastomosing tonalite to granite veins. These pods are strained and well foliated, yet they do not display gneissic texture. The tonalite to granite migmatite display southwest-plunging stretching lineations (10°–30°). Metabasites and mafic gneisses also occur with assemblages of Grt + Cpx + Opx + Pl + Qz + Hbl + Bt. On southern Wholdaia Lake, these gneisses contain poorly defined compositional layering striking northeast and dipping moderately (~60°) to the southeast. Their peak assemblage contains coarse-grained (>2 mm) Grt + Cpx + Opx + Pl + Qz overprinted and replaced by Opx + Pl + Hbl + Bt.

Along strike, on the northern Wholdaia Lake, mafic gneiss units have discrete layers of Grt + Cpx + Pl + Qz (Fig. 4B) and Grt + Opx + Pl + Hbl + Qz (Fig. 4C). Layers with Grt + Cpx + Pl + Qz contain two texturally distinct assemblages. The first assemblage (M1) occurs as 1–2 cm elongate pods containing 0.5–1.0-cm-sized, partially disaggregated Grt, crystals and coarse Cpx + Pl with minor quartz, ilmenite, and pyrite (Figs. 4C–4D). These large garnet crystals host small (<10 µm) ilmenite inclusions, but otherwise they appear mostly inclusion free. A minor amount of quartz occurs as small, rounded crystals interstitial to Cpx. Plagioclase forms coronas around coarse garnet grains and occurs within late fractures. Based on the observed microtextures, we suggest that plagioclase is a retrograde phase overprinting a coarse-grained, high-pressure, granulite-verging on eclogite-facies assemblage; however, we cannot rule out entirely that a small portion of plagioclase in the fractures originated as inclusions in Grt.

A second texturally distinct assemblage (M2) of Grt + Cpx + Pl + Qz + Ilm (Figs. 4C–4D) has a finer granoblastic texture (<1 mm) and envelopes the older M1 domains. Garnet grains (Grt) in this assemblage contain small round inclusions of quartz, plagioclase, and apatite.

Additional centimeter-scale layers and pods of Grt + Opx + Pl + Hbl + Qz, which occur adjacent to the Grt + Cpx + Pl + Qz + Ilm layers...
Figure 4. (A) Northeast-trending amphibolite pods and tonalitic veins of S₁, with late east-west–oriented fracturing. (B) (Grt + Cpx)–bearing mafic gneiss of S₁. (C) Lenses of Grt + Opx + Pl + Hbl + Qz and Grt + Cpx + Pl + Qz (M1) enveloped by a fine-grained Grt + Cpx + Pl + Qz + Ilm (M2) assemblage. (D) Cross-polarized photomicrograph of the texturally distinct M1 Grt + Cpx + Pl + Qz assemblage and M2 Grt + Cpx + Pl + Qz + Ilm assemblage in sample 15ET249. (E) Highly sheared mylonites of S₂, dipping steeply to the southeast with a shallowly southwest-plunging lineation. (F) Photomicrograph of S₂, tonalite mylonite showing granoblastic textures in plagioclase consistent with late thermal annealing. Mineral abbreviations are according to Whitney and Evans (2010).
(Fig. 4C), may represent a retrograde assemblage or an early assemblage that did not equilibrate at higher pressures. Garnet in the (Opx + Hbl)–bearing layers contains many rounded Qz + Pl inclusions. This (Opx + Hbl)–bearing assemblage does not display granoblastic textures, yet it has straight grain boundaries and appears to be in textural equilibrium. The boundary between compositional layers is sharp with no overlap of assemblages. East-west–oriented hairline fractures associated with late overprinting green-amphibole occur within these gneissic units, suggesting later retrogression.

D2–S\textsubscript{T2}

A package of penetratively transposed (Hbl + Bt)–bearing mylonites (S\textsubscript{D2}) defined by a northeast-trending, low-magnetic linear pattern overprints the higher-grade gneissic rocks of S\textsubscript{T1} (Fig. 3). These rocks are typically dioritic to tonalitic in composition, with minor occurrences of paragneiss. The S\textsubscript{D2} rocks include millimeter- to centimeter-scale, rectilinear banded, highly sheared mylonites to ultramylonites striking northeast and dipping ~85° to the southeast (Fig. 4E). Although these rocks are highly sheared, they do show signs of grain coarsening and development of granoblastic textures indicative of thermal overprinting (Fig. 4F). Abundant dextral shear sense indicators are present in the horizontal plane, including winged porphyroclasts, z-folds in crosscutting dikes, and C- and C'-type shear bands (Fig. 5A). On the S\textsubscript{D2} foliation, a 10° to 60° southwest-plunging stretching lineation occurs, and oriented thin sections in the x-z-plane indicate displacement of the Snowbird domain down to the southwest. In southeast Wholdaia Lake, progressively overprinted mineral assemblages and structural fabrics are observed, and a continuum exists between higher-grade, (Grt + Cpx + Opx)–bearing gneiss tectonites of S\textsubscript{T1}, and lower-grade, (Hbl + Bt)–bearing mylonites of S\textsubscript{D2}.

A distinctive (Hbl + Bt) ultramylonite-mylonite (Figs. 5C–5E) occurs within the northern mapped reaches of the WLsz, where foliations dominantly strike more easterly and the magnetic signature is very low (refer to easterly beds [EB] in Figs. 2–3). This domain has a dip of ~85° (Fig. 5C), contains a southwest-plunging stretching lineation and dextral sense of shear, and has a minimum mapped width of 1 km, although the geophysical signature is ~5 km wide (Fig. 3). This unit has relatively similar structures and metamorphic grade compared to the S\textsubscript{D2} mylonites of central Wholdaia Lake; however, it preserves very fine-grained ultramylonites (Fig. 5D) with a lower degree of high-temperature static overprinting textures (Fig. 5E). For example, recrystallized quartz crystals show recovery textures akin to low-temperature mylonites (200–500 °C; Trouw et al., 2009), whereas mylonites elsewhere in S\textsubscript{D2} typically contain static recrystallization textures and chessboard-quartz subgrains indicative of temperatures consistently above 550 °C (Stipp et al., 2002). Although these rocks preserve lower-temperature microstructures than the bulk of S\textsubscript{D2}, we still consider them part of the S\textsubscript{D2} structural domain.

Minimum age constraints on development of S\textsubscript{D2} come from variably crosscutting mafic and felsic dikes. North of Wholdaia Lake, lamprophyre dikes cut the mylonitic fabric in an east-west orientation and are considered part of a regional swarm dated at ca. 1.83–1.81 Ga (Rainbird et al., 2006; Ashton et al., 2009). Numerous felsic dikes obliquely cut S\textsubscript{D2} and are themselves weakly folded, foliated, and locally sheared (Fig. 5B). Enveloping surfaces for these folded dikes are generally oriented northwest-souttheast (perpendicular to S\textsubscript{D2} foliations; Fig. 5B) and have subvertical dips, whereas axial planar foliations in the dikes are parallel to the S\textsubscript{D2} foliations. The dikes display the penetrative southwest-plunging lineation associated with S\textsubscript{D2}, indicating late-kinematic emplacement. Crystallization ages for two of these felsic dikes are presented in the results section.

Snowbird Domain Hanging Wall

Rocks of the Snowbird domain that underlie the arcuate low-magnetic-anomaly immediately east of the WLsz S\textsubscript{T1} (Fig. 3) contain a migmatitic metasedimentary gneiss with the assemblage Grt + Sil + Qz + Kfs + Bt + Rt. The sampled unit (15ET273b) is relatively homogeneous in texture, with millimeter-scale compositional bands of Grt + Sil + Bt and Qz + Kfs, is strongly foliated, and displays ribbons of quartz and foliation-parallel prismatic sillimanite (Fig. 5F). The high amount of garnet suggests this unit is a highly residual rock that has lost a significant amount of melt.

U-Pb GEOCHRONOLOGY

In order to constrain the protolith ages and timing of metamorphism in the WLsz and its hanging wall, five samples were selected (Fig. 3) for zircon U-Pb isotopic analysis. Samples 15ET249 and 15EM68b are mafic granulites from within the S\textsubscript{T1} portion of the WLsz in which zircon grains preserve both igneous and metamorphic crystallization ages. Samples 15ET253a and 15ET260d are felsic dikes that crosscut the S\textsubscript{T2} portion of the WLsz, and their igneous crystallization ages provide minimum ages for ductile shearing. The fifth sample, 15ET273b, is a metasedimentary gneiss from the Snowbird domain previously analyzed by SHRIMP (Thiessen et al., 2017). During this initial analysis (Thiessen et al., 2017), metamorphic zircon crystals were targeted; however, five detrital zircon cores were identified and analyzed, with resulting ages as young as 1.96 Ga. Follow-up LA-ICP-MS work (presented herein) was conducted to better constrain the maximum depositional age through dating additional detrital zircon and to characterize the metamorphic and detrital zircon populations using trace-element analyses.

Analytical Methods

U-Pb SHRIMP geochronology of zircon was conducted at the Geological Survey of Canada’s J.C. Roddick Ion Microprobe Facility in Ottawa, Ontario, while U-Pb LA-ICP-MS geochronology of zircon was conducted at the Isotope Geology Laboratory at Boise State University, Boise, Idaho. Refer to the Data Repository supplemental material1 for detailed descriptions of sample preparations, analytical techniques, and the treatment of geochronological data, including common-Pb corrections and results of primary and secondary reference materials. Zircon separated were mounted in epoxy grain mounts, polished to expose grain centers, and imaged using either backscattered electron (BSE) or cathodoluminescence (CL) imaging for targeting of spot analyses. U-Pb data are presented in Tables 1–2 and Figures 6–9. Interpreted ages are reported using the calculated weighted mean 30\textsuperscript{th} Pb/206\textsuperscript{th} Pb date at a 95% confidence level (2σ) unless otherwise noted.

U-Pb GEOCHRONOLOGY RESULTS

Sample 15ET249: Mafic Granulite (SHRIMP)

This sample was collected from northern Wholdaia Lake (Fig. 3), where gneissic rocks exhibit a northeast-striking S\textsubscript{T1} foliation. This mafic granulite gneiss contains Grt + Cpx + Opx and occurs adjacent to (Grt

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1GSA Data Repository Item 2018268, which includes (1) a text document of U-Pb analytical methods, trace-element analyses, and detrital zircon probability density diagrams of 2.0 Ga south Rae craton basins discussed in the text, (2) Table S1 of trace-element data, and (3) Excel versions of Tables 1 and 2 from the main text, providing SHRIMP and LA-ICP-MS U-Pb data, is available at http://www.geosociety.org/datalibrary/2018, or on request from editing@geosociety.org.
Figure 5. (A) C- and C’-type dextral shear bands developed in S$_2$ tonalite mylonites from an island shown in B. (B) Aerial view of deformed dikes within S$_2$ tonalite mylonite. (C) Ultramylonite developed along the northern easterly bend (EB of Fig. 3) of the Wholdaia Lake shear zone. (D) Highly deformed interlayered mafic and felsic components of an ultramylonite close to area in C. (E) Photomicrograph of an ultramylonite (from part C) that preserves fine-grain sizes and lacks substantial thermal annealing textures. (F) Outcrop photo of Snowbird domain migmatitic metasedimentary gneiss 15ET273b sampled for detrital zircon geochronology. Mineral abbreviations are according to Whitney and Evans (2010).
TABLE 1. U-Pb SENSITIVE HIGH-RESOLUTION ION MICROPROBE (SHRIMP) ZIRCON DATA

| Spotted zircon | Apparent ages (Ma) |
|----------------|--------------------|
| 10.1 CL bright 46 | 8.37 11.14 15.122 18.60 | 0.11 0.457 1.10 0.988 0.1624 0.22 4272 22 4813 3 |
| 26.1 CL bright 46 | 10.16 13.98 18.76 | 0.11 0.457 1.10 0.988 0.1624 0.22 4272 22 4813 3 |
| 35.2 CL bright 46 | 11.12 15.122 18.60 | 0.11 0.457 1.10 0.988 0.1624 0.22 4272 22 4813 3 |

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| Spot | Zircon domain | U (ppm) | Th (ppm) | Th/U | Hf/Yb | Err (%) | 206Pb/238U | Isotopic ratios |
|------|---------------|---------|---------|------|--------|---------|-------------|----------------|
| 30.1 | c oz 1152    | 1178    | 678     | 0.59 | 34     | 0.08    | 343         |                |
| 30.1 | c oz 1375    | 1395    | 778     | 0.59 | 34     | 0.08    | 343         |                |
| 30.1 | c oz 1352    | 1378    | 807     | 0.59 | 34     | 0.08    | 343         |                |
| 30.1 | c oz 1279    | 1279    | 542     | 0.59 | 34     | 0.08    | 343         |                |
| 30.1 | c oz 1265    | 1265    | 100     | 0.59 | 34     | 0.08    | 343         |                |
| 30.1 | c oz 1175    | 1175    | 100     | 0.59 | 34     | 0.08    | 343         |                |

Note: Spot names follow the convention y.z, where y—grain number and z—spot number. Uncertainties are reported at 1σ (%) and were calculated by propagation of all known sources of error using Squid version 2.22. Errors in ages in 1σ are absolute. Data are referenced to Pb, Th, and U abundances and Th/U and Hf/Yb ratios. Isotopic compositions are reported on the SPb method common for the Rb, Sr, Sm, and Nd systematics.
### TABLE 2. LASER ABLATION–INDUCTIVELY COUPLED PLASMA–MASS SPECTROMETRY (LA-ICP-MS) U-Pb ZIRCON DATA

| Spot name | U (ppm) | Th (ppm) | Pb (ppm) | Th/U | 206Pb/204Pb | 207Pb/204Pb | 208Pb/204Pb | 206Pb/238U | 207Pb/235U | 208Pb/235U | 230Th/235U | 232Th/235U | 238U/235U | 235U/238U | 238U error (Ma) | 235U error (Ma) | 238U error (Ma) | Discord. (%) |
|-----------|---------|----------|----------|------|-------------|-------------|-------------|------------|------------|------------|-------------|-------------|----------|-----------|----------------|----------------|----------------|------------|
| Core 1 | 352 186 218 0.53 1274 0.0171 11.6 6.61 | 3.9 9.574 5.6 | 0.4257 4.0 | 0.0697 2.39 | 4.0 0.1631 3.9 | 1.236-16 | 2489 65 | 2287 77 8 |
| Core 2 | 241 240 161 0.08 6465 0.1347 3.3 | 6.666 | 1.5 8.413 4.7 | 0.4067 4.4 | 9.273 2.49 | 5.4 0.1500 5.1 | 1.577-10 | 2346 25 | 2230 83 6 |
| M1 1 | 257 263 128 0.16 1571 0.0152 5.0 | 6.607 | 2.2 8.564 4.6 | 0.4215 4.7 | 6.271 2.35 | 2.4 0.1491 2.5 | 0.000+00 | 2334 42 | 2286 46 2 |
| Core 3 | 288 313 252 0.18 12106 0.1613 10.8 | 18.05 | 1.7 8.349 4.3 | 0.2846 3.9 | 7.962 1.88 | 4.4 0.1548 4.74 | 2169 30 | 2119 77 7 |
| Core 4 | 353 592 433 0.35 6034 0.0094 2.0 | 7.039 | 1.7 8.316 6.6 | 0.4348 6.4 | 9.607 2.30 | 6.4 0.1421 1.7 | 6.645-10 | 2431 27 | 2387 125 3 |
| M1 5 | 886 220 453 0.25 3776 0.1262 4.2 | 7.050 | 1.9 8.241 4.2 | 0.4124 4.4 | 6.117 2.37 | 3.6 0.1418 1.9 | 0.000+00 | 2250 34 | 2267 87 1 |
| L239 | 516 512 127 0.12 1662 0.1294 5.9 | 7.078 | 1.0 8.335 3.9 | 0.5674 3.6 | 6.824 1.48 | 4.4 0.1431 1.4 | 2433 17 | 2441 77 7 |
| M1 6 | 152 179 34 0.26 920 0.1554 10.5 | 7.276 | 2.7 7.412 4.3 | 0.3913 3.3 | 7.858 2.56 | 3.3 0.1374 2.7 | 0.000+00 | 2195 47 | 2129 60 3 |
| M1 7 | 850 636 0.23 5.8 | 7.298 | 1.6 8.508 7.7 | 0.4499 7.6 | 9.755 2.23 | 7.6 0.1372 1.6 | 2.97-15 | 2152 27 | 2208 252 2 |
| L244 | 722 365 371 0.51 13175 0.1486 13.1 | 7.700 | 3.4 6.139 4.8 | 0.3429 3.2 | 6.647 2.91 | 3.3 0.1329 3.4 | 3.576-10 | 2069 60 | 1990 54 9 |
| M1 8 | 201 411 0.93 3.9 | 8.103 | 1.8 8.330 4.7 | 0.5529 4.1 | 8.212 2.4 | 1.7 | 0.1270 1.7 | 0.000+00 | 2143 17 | 2151 77 7 |
| Core 9 | 458 161 224 0.35 16096 0.1351 13.4 | 7.971 | 1.6 6.335 3.9 | 0.3663 3.2 | 8.781 2.73 | 1.0 | 0.1254 1.6 | 3.326-10 | 2035 29 | 2021 61 1 |
| M1 9 | 590 199 429 0.21 3332 0.1201 9.3 | 8.134 | 1.4 6.456 4.6 | 0.3868 4.6 | 9.512 2.62 | 4.6 | 0.1229 1.4 | 2.81-10 | 2080 92 64 2 |
| Core 10 | 616 266 0.69 1.4 | 8.311 | 1.3 6.439 4.4 | 0.3477 3.3 | 8.064 2.42 | 3.7 | 0.1336 1.7 | 0.000+00 | 2046 38 | 2077 64 8 |
| L247 | 146 127 650 0.04 4709 0.1519 10.2 | 8.250 | 1.1 6.305 4.8 | 0.3566 4.1 | 8.982 2.8 | 3.4 | 0.1212 1.1 | 6.576-10 | 1974 20 | 1966 82 0 |
| M1 11 | 624 449 0.86 2.3 | 8.357 | 1.8 6.557 4.7 | 0.4336 3.6 | 8.879 2.8 | 4.8 | 0.1277 1.2 | 0.000+00 | 2146 20 | 2167 76 4 |
| M1 12 | 619 46 250 0.07 3785 0.1364 22.7 | 8.570 | 2.7 8.786 2.4 | 0.3512 3.8 | 8.850 2.4 | 8.0 | 0.1240 2.0 | 6.700-10 | 1963 35 | 1940 63 1 |

### Note:
- Isotope ratios and ages are NOT corrected for initial common Pb. Isotope ratio and apparent age errors do NOT include systematic calibration errors of 0.3009365857×10^-4%. (206Pb/204Pb), 0.6195676262×10^-3%.(207Pb/204Pb), (208Pb/204Pb) (all 1 sigma). Isotope ratios and ages were estimated using a measured secondary standard and calculated using the relative isotope ratio method. Since byssoweberite breakdown fraction of U-Pb isotopic calculation was corrected using the Sn/Zr fractionation. Backgrounds were monitored between sieves 12 and 32. Sample counts were integrated from sweeps 35 to 60. A laser firing repetition rate of 10 Hz. NAD83 UTM—Universal Transverse Mercator North American Datum 1983.
Figure 6. (A) Cathodoluminescence (CL) images of representative zircon analyzed in mafic granulite sample 15ET249 (spots are 20 µm for scale) that show igneous-zoned cores and bright-CL recrystallized rims. (B) Concordia plot with calculated igneous (filled red) and metamorphic (filled blue) crystallization populations highlighted. Open red ellipses correspond with bright-CL zircon analyses that are discordant or old. (C) Th/U vs. Hf/Yb ratios of all analyses color-coded by 207Pb/206Pb age. Trend of decreasing Th/U and increasing Hf/Yb with decreasing age is apparent.

Figure 7. (A) Zircon core and rim analyses from mafic granulite sample 15EM68b (backscattered electron [BSE] images, spots are 20 µm for scale). (B) Concordia plot showing igneous (2.6 Ga) and metamorphic (1.9 Ga) populations (filled red ellipses were used for weighted mean calculation). (C) Th/U vs. Hf/Yb ratios of all analyses color-coded by 207Pb/206Pb age. Younger analyses have relatively high Hf/Yb and generally lower Th/U values.
+ Cpx)-bearing intermediate to felsic gneiss. Zircon crystals in sample 15ET249 were irregular to subrounded, and stubby to elongate (2:1–3:1 aspect ratio), with few crystal facets, and varied from light brown to pale pink to clear in color. CL imaging revealed three internal zonation patterns (Fig. 6A) that were targeted for SHRIMP analysis: (1) moderate to widely spaced oscillatory zoning (e.g., Fig. 6A, #115, #48, #109, #12) that was relatively darker under CL, (2) dark- to moderately bright-CL sector and patchy zoning (e.g., Fig. 6A, #50, #34, #44), and (3) bright-CL overlaminations/rims on oscillatory and sector/patchy zoned zircon cores (e.g., Fig. 6A, #34, #44, #10, #22). Oscillatory and sector/patchy patterns in zircon commonly exhibited both faintly blurred or faded zoning (e.g., Fig. 6A, #34) as well as crisp well-defined zoning (e.g., Fig. 6A, #12). Bright-CL zones and rims varied from homogeneous zones that truncated primary zoning to gradational boundaries within zoned zircon. These rims were generally 10–20 µm wide and exhibited sinuous transgressive inner margins, whereas their outer margins usually had rounded to irregular morphologies.

Thirty-four analyses were conducted on 25 individual zircon grains yielding $^{207}$Pb/$^{206}$Pb dates from 2608 to 1858 Ma (Fig. 6B; Table 1), with two prominent age populations at ca. 2.6 Ga and at ca. 1.9 Ga. Analyses of 20 oscillatory and sector/patchy zoned zircon span dates of 2608–1921 Ma and lie along a discordia line anchored at ca. 2.6 Ga, with a lower intercept at ca. 1.9 Ga (black unfilled ellipses, Fig. 6B). The youngest oscillatory zoned zircon analysis (#115) was concordant and yielded a $^{207}$Pb/$^{206}$Pb date of 1921 ± 13 Ma ($\sigma$). The Th/U values for this group were all between 0.2 and 0.38, and Hf/Yb values were between 27 and 120 (Fig. 6C). The youngest 1921 Ma analysis had a Hf/Yb value of 202 owing to its low Yb concentration of 39 ppm. No duplicate analyses were performed, so Pb-loss within individual grains could not be assessed, yet the discordia array implies that Pb loss affected this population. The five oldest analyses yielded a $^{207}$Pb/$^{206}$Pb date of 2592 ± 16 Ma (mean square of weighted deviates [MSWD] = 2.2). Analyses younger than the oldest five were excluded due to greater discordance and the presence
of homogeneous low-U rims, suggesting these younger analyses were more significantly affected by Pb loss. The single oscillatory zoned zircon (Fig. 6A, #115) with a 1.9 Ga concordant date is interpreted to have been completely isotopically reset by 1.9 Ga metamorphism, but it preserves relict oscillatory zoning or “ghost texture” (Hoskin and Black, 2000) from igneous crystallization.

The second group consists of analyses within 14 bright-CL domains with $^{207}\text{Pb}/^{235}\text{U}$ dates ranging from 2569 to 1858 Ma, with a cluster at ca. 1.9 Ga and two concordant single analyses at 2569 ± 27 Ma and 2372 ± 44 Ma (1σ; red-rimmed ellipses in Fig. 6B). This group had Th/U values between 0.14 and 0.26 and Hf/Yb values between 118 and 393, with the oldest 2569 Ma analysis having a Hf/Yb value of 60 (Fig. 6C). These higher Th/U values are very similar to the values of the older 2592 Ma population. The Hf/Yb values, however, are much higher than the younger population, except for the oldest 2569 Ma analysis, which has a similar value. These older dates (2.57 Ga, 2.37 Ga) in bright-CL domains exhibit large errors and have higher Hf/Yb values, possibly indicating growth during a separate event, after Yb depletion of the host rock. We interpret these older, bright-CL zircon domains to be younger metamorphic recrystallization domains overgrowing 2.6 Ga zircon that underwent incomplete isotopic resetting at 1.90 Ga. No clear discordia array exists within this population, nor are there any significant differences in chemistry or zoning characteristics. A cluster of the youngest eight analyses with comparable chemistry yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean date of 1901 ± 8 Ma (MSWD = 1.8). Younger analyses were excluded due to increasing discordance and age clustering. Owing to the low U (34–62 ppm) of these bright-CL domains, their dates are less precise relative to the dark-CL domains with higher U (57–444 ppm). Nevertheless, these data show two distinct zircon populations at ca. 2.6 Ga (igneous crystallization) and 1.9 Ga (metamorphic recrystallization), distinguished by date, internal zoning patterns, and chemistry. Additionally, a regression anchored at 2.6 Ga has a lower intercept of 1.9 Ga, within error of the four youngest analyses.

Sample 15EM68b: Mafic Granulate (SHRIMP)

This sample was collected 80 km southwest of Wholdaia Lake within the WLSz (Fig. 3) and contained (Grt + Cpx + Opx + Hbl)—bearing metagabbro and metadiorite gneiss (Regis et al., 2017b). These gneissic rocks display strong overprinting by the $S_T1$ foliation. Zircon crystals in sample 15EM68b were subrounded, and stubby to elongate (2:1–3:1 aspect ratio), with few crystal facets, and were generally light brown in color. BSE imaging revealed two primary internal textures (Fig. 7A) that were targeted for SHRIMP analysis: (1) homogeneous to oscillatory zoned cores and whole zircon crystals, and (2) 5–50 µm homogeneous darker rims. Rims exhibited sharp truncations with zircon cores and had sinuous transgressive inner margins overprinting cores. Thirty-one analyses were conducted on 26 individual zircon grains with $^{207}\text{Pb}/^{206}\text{Pb}$ dates that ranged from 2604 to 1894 Ma (Table 1). Like sample 15ET249, two prominent age populations can be observed in the data, ca. 2.6 Ga and 1.9 Ga (Fig. 7B). Twenty-two oscillatory zoned and homogeneous zircons ranged between 2604 and 2184 Ma, which consisted of a concordant grouping at ca. 2.6 Ga and clear discordia toward ca. 1.9 Ga. The Th/U values for this older group decreased with age from 0.5 to 0.34, and Hf/Yb values were consistently <100 (Fig. 7C). No duplicate analyses were performed, so Pb loss within individual grains was not assessed; however, the discordia array does imply that Pb loss affected this population. The three oldest concordant analyses yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ date of 2602 ± 8 Ma (MSWD = 1.8). Younger analyses were excluded due to increasing discordance and presumed Pb loss.

The second group consisted of five homogeneous rim domains and four homogeneous cores that spanned dates from 1974 to 1894 Ma, with a cluster at ca. 1.9 Ga. This group had Th/U values between 0.20 and 0.61 and Hf/Yb values mostly >100. The three oldest homogeneous zircon cores had Yb values of 140–266 ppm, whereas the remaining six grains yielded Yb values of 36–86 ppm. The cluster of six youngest concordant analyses yielded a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date of 1901 ± 8 Ma (MSWD = 3.2). A regression anchored at 2.6 Ga had a lower intercept of 1.9 Ga, within error of the six youngest analyses. These two zircon
populations are interpreted to represent igneous crystallization at 2.6 Ga and metamorphic (re)crystallization at 1.9 Ga.

Sample 15ET253a: Felsic Dike (SHRIMP)

This sample represents a felsic dike of granitic to granodioritic composition that crosscuts mylonites of S_{12} (Fig. 8A) in the central WLsz. The sampled portion was 50 cm wide, was locally folded, and contained a weak axial planar foliation parallel to S_{12}. Only 64 zircon crystals were recovered from a 5 kg sample, which were poor in quality, irregular to subhedral, and brown. BSE imaging revealed mostly altered, yet prismatic and oscillatory zoned crystals, as well as larger less-altered homogeneous blocky subhedral crystals (Fig. 8B). Twenty-two analyses were conducted on 20 separate zircon crystals. The results fall into two age clusters, one at ca. 2.60 Ga and one at ca. 1.86 Ga (Fig. 8C; Table 1). Zircon crystals that form the older cluster were blocky to prismatic and homogeneous to weakly oscillatory zoned. This group of 13 crystals yielded 207Pb/206Pb dates between 2628 and 2595 Ma, Th/U values between 0.1 and 1.0, and Hf/Yb values between 34 and 128. Zircon crystals that form the younger population were prismatic (4:1 aspect ratio) with wide oscillatory zones (Fig. 8B) and Th/U values between 0.01 and 0.1, Hf/Yb values between 10 and 100, and U values of 819–4228 ppm. This group of nine grains had a range of 207Pb/206Pb dates between 1867 and 1858 Ma. These nine prismatic oscillatory zoned zircon yielded a 207Pb/206Pb weighted mean date of 1864 ± 2 Ma (MSWD = 0.39). The low Th/U values associated with the young population are due to anomalously high U concentrations, which are not typical of metamorphic zircon (Hoskin and Schaltegger, 2003), and which result in slightly reversely discordant (~1.5%) analyses. Clustering of analyses at 1.86 Ga and the well-defined internal oscillatory zones are indicative of magmatic crystallization and, therefore, we interpret ca. 1864 Ma to record the igneous crystallization age of the dike and the ca. 2.6 Ga zircon population to represent an inherited population that matches the host-rock ages (Thiessen et al., 2017).

Sample 15ET260d: Felsic Dike (SHRIMP)

Northeast of Wholdaia Lake, this 30-cm-wide alkali-feldspar granite dike intruded obliquely across the S_{12} mylonitic host rocks at a low angle (Fig. 8D). Zircon crystals were spherical to prismatic and mostly brown in color. BSE imaging displayed altered prismatic to spherical zircon crystals (Fig. 8E, #42, #35) that showed homogeneous to faint oscillatory zoning and rare core-rim relationships (Fig. 8E, #66). Zircon rims were thin (10–20 m) and had homogeneous to oscillatory zoning. Thirty-one analyses on 23 individual zircon grains yielded 207Pb/206Pb dates between 2482 and 1862 Ma (Fig. 8F; Table 1). The results fall into three age clusters at 2.4 Ga, 1.93 Ga, and 1.87 Ga. The oldest age cluster is between 2482 and 2350 Ma and contains 10 mostly concordant analyses with large errors of ~20 m.y. (1σ). These zircon crystals are oscillatory zoned or homogeneous whole grains. The age cluster at 1.93 Ga consists of eight analyses of homogeneous whole zircon, homogeneous rims, and one oscillatory zoned zircon. These analyses ranged from 1994 to 1919 Ma; all but one had Th/U below 0.1; and all had Hf/Yb values above 100. The youngest age cluster at 1.87 Ga consists of nine mostly concordant analyses of homogeneous to oscillatory zoned whole zircon and rims with 207Pb/206Pb dates ranging from 1883 to 1862 Ma (Fig. 8F). The Th/U values were above 0.1 (one analysis was 0.09), and the Hf/Yb values were consistently above 100. These nine youngest analyses yielded a weighted mean 207Pb/206Pb date of 1871 ± 5 Ma (MSWD = 1.9). The oscillatory nature of the 1871 Ma population, coupled with their igneous chemistry, leads us to interpret 1871 Ma as the igneous crystallization age of the dike. All older results from this sample are considered inherited. Interestingly, no 2.6 Ga zircon crystals were analyzed despite the ca. 2.6 Ga nature of the local host rocks.

Sample 15ET273b: Migmatitic Metasedimentary Gneiss (LA-ICP-MS)

Sample 15ET273b was collected to determine a maximum depositional age for this metasedimentary rock and to relate its metamorphic history to the WLsz. This migmatite-bearing Grt + Sil + Kfs + Bt + Qz metasedimentary gneiss (Fig. 5F) was collected from southeastern Wholdaia Lake (Fig. 3) within the Snowbird domain adjacent to the WLsz. These rocks have the same low magnetic response as the metasedimentary rocks 50 km to the east described in Martel et al. (2008). Zircon crystals from sample 15ET273b were largely spherical and rounded with minor elongate 3:1 crystals and varied from brown to almost clear. BSE imaging showed two distinct zircon zonation textures (Fig. 9A), (1) mostly homogeneous spherical zircon (Fig. 9A, #16, #34, #24, #20), and (2) rare oscillatory zoned cores (Fig. 9A, #17, #27, #36). CL imaging of the homogeneous spherical zircon crystals revealed bright and dark, well-developed, fir-tree (e.g., Rubatto, 2017) and sector zoning (Fig. 9A, #16, #34, #24, #20, #25) with inclusions of sillimanite. Zircon cores were resorbed but retained oscillatory zoning in both BSE and CL and luminesced both bright and dark. The few elongate rounded crystals displayed faint thin oscillatory zoning in BSE and CL and were commonly rimmed by fir-tree zoned zircon (e.g., Fig. 9A, #218).

Due to the paucity of preserved zircon cores, only 18 were observed that were amenable for analysis. These analyses are mostly discordant and yielded 207Pb/206Pb ages between 2.48 and 1.96 Ga (Table 2). In general, the data have large errors and display more discordance among older ages; however, two distinct populations are evident—an older population with 207Pb/206Pb ages between 2.50 and 2.20 Ga and a younger population with ages between 2.10 and 1.96 Ga (Fig. 9B). Broadly, the older population exhibited oscillatory zoning and both dark and light CL. The younger population commonly exhibited dark and homogeneous CL with rare bright-CL cores. Due to the high metamorphic grade and paucity of zircon core domains in the sample, a conservative approach was used for assessing the maximum depositional age (YC2tα+ method, which uses the youngest three [or more] grains overlapping within 2σ uncertainty; Dickinson and Gehrels, 2009). The mean age of the youngest five concordant zircon analyses that overlapped with 2σ uncertainty yielded a date of 1983 ± 19 Ma (MSWD = 1.3), and this is considered a conservative estimate for the maximum depositional age.

The fir-tree and sector zoned spherical zircon population was also analyzed (Table 2). This population yielded 207Pb/206Pb ages between 1.96 and 1.90 Ga with 2σ errors of ~20–30 m.y. per analysis. This population was previously well characterized by U-Pb SHRIMP analyses and discussed by Thiessen et al. (2017), who obtained an age range of 1.93–1.88 Ga and a 207Pb/206Pb weighted mean of 1915 ± 4 Ma (MSWD = 2.7).

Zircon crystals were also analyzed for trace elements (see Data Repository material and Table S1 [footnote 1]). A clear distinction existed in the Th/U values (Fig. 9C), total rare earth element (ΣREE) values, and chondrite-normalized REE values, and chondrite-normalized REE values, and by observing the covariance of U versus Th between older core domains and younger fir-tree zoned overgrowths. Zircon cores are interpreted to be detrital and had Th/U and ΣREE above 0.1 and 100 ppm, respectively, whereas metamorphic fir-tree zoned zircon domains consistently showed opposite trends. Zircon core domains had chondrite-normalized REE values that are typical for normal igneous zircons, displaying a moderately positive slope, positive Ce* anomaly, and a negative Eu* anomaly (Hoskin and Schaltegger, 2003). Metamorphic zircon domains differed from core analyses by having lower...
total concentrations of REEs, a steeper positive slope among the light (L) REEs, a slightly larger negative Eu* anomaly, and a shallow negative slope within the heavy (H) REEs (Fig. 10). The zircon cores lacked a distinct age population, were commonly oscillatory zoned, and had variable U versus Th, further supporting a detrital origin.

**DISCUSSION AND INTERPRETATION**

**Neoarchean Magmatism**

Our study and previous work (Davis et al., 2015; Thiessen et al., 2017) on WLSz host rocks show that these rocks formed predominantly in the Neoarchean, at ca. 2.6 Ga. As noted herein, the distinct discordia arrays, along with the well-clustered oscillatory and sector zoned zircon populations (with high Th/U values) in mafic granulate samples 15ET249 and 15EM60b, are interpreted to represent igneous crystallization at 2.6 Ga for the related protoliths. This Archean magmatism, termed the Snow Island Regionally, is well documented throughout the Rae craton as occurring during a major crust-forming event (e.g., Davis et al., 2015; Peterson et al., 2015; Regis et al., 2017a). Previously, mappable exposure of the mafic components of this event were primarily identified northeast of the Thelon Basin (Peterson et al., 2015); however, the units in the WLSz and adjacent Firedrake domain (Davis et al., 2015) expand the distribution of their setting.

**Snowbird Domain Metasedimentary Rocks**

**Constraints on Timing of Deposition**

The migmatitic metasedimentary gneiss (15ET273b) collected in the hanging wall of the WLSz has a restricted range of detrital and metamorphic grains. A conservative estimate for maximum depositional age, from the five youngest concordant analyses that overlap within 2σ uncertainty, is 1983 ± 19 Ma. The oldest metamorphic U-Pb zircon analysis from the same sample is 1932 ± 5 Ma (1σ; Thiessen et al., 2017), and so the maximum depositional range of ages for this package is constrained to 22–80 m.y. between 1983 ± 19 Ma and 1932 ± 5 Ma.

**Regional Correlation and Provenance**

The 1.98–1.93 Ga depositional age, the range of ages between 2.4 and 1.96 Ga, and the absence of Neoarchean zircon in sample 15ET273b restrict the possible sources of detritus. A viable potential source includes the 1.99–1.95 Ga I-type intrusions of the Taltson magmatic zone, which bound the south Rae craton to the west and south. Rocks of this age to the east of the present-day Hearne craton within the THO were not accreted to its southeast margin until ca. 1.87 Ga (Martel et al., 2008), which was well after basin inversion, tectonic thickening, and metamorphism of the Snowbird domain metasedimentary rocks. Rocks of the 2.1 Ga Clearwater anorhösite (Card et al., 2014) and Chipman dikes (Regan et al., 2016), west of the STZ, constitute another potential proximal source of detritus; however, these mafic units contain very few zircon of this age, and therefore they probably did not contribute significant 2.1 Ga zircon to nearby sedimentary basins.

The lack of Archean dates and zircon as young as 1.96 Ga lead us to suggest that a large proportion of the detritus came from the Taltson magmatic zone between 1.98 and 1.93 Ga. This is consistent with surface uplift and erosion of the Taltson magmatic zone (Fig. 11A) coeval with presumed crustal thickening during metamorphism at ca. 1.93 Ga (McDonough et al., 2000).

Our detrital zircon results have similar depositional and metamorphic ages compared to other south Rae craton metasedimentary rocks and rocks farther north in the central Rae craton with ages of ca. 2.0 Ga. Detrital samples from the south Rae craton have interpreted maximum depositional ages of ca. 2.0 Ga, with minimum depositional ages of 1.97 Ga Ma (Knox and Ashton, 2016; Ashton et al., 2017a), 1.99 Ma (Ply, 2016), and 1.93 Ga (Martel et al., 2008), and all have abundant metamorphic zircon at 1.91 Ga. Additionally, the zircon data of Martel et al. (2008), Shiels et al. (2016), and Ashton et al. (2017a) are remarkably similar to our detrital data, in that the majority of data plot between 2.4 and 2.0 Ga, with an exceedingly small Neoarchean signature. Collectively, these metasedimentary rocks have similar depositional time ranges to sequence 3 of the Rae cover sequence, as documented by Rainbird et al. (2010). These similar assemblages of shale, carbonate, banded-iron formation, and sandstones are interpreted to have been deposited on a thinning lithosphere developing on a passive margin related to a rifting event, with detrital sources from the west-central Rae craton (Rainbird et al., 2010).

The presence of major sedimentary basins deposited between ca. 2.0 and 1.95 Ga, the voluminous anorthosite intrusions in the southeast Rae craton (Card et al., 2014), and the 2.1 Ga age recently acquired for the Chipman mafic dikes along the margin of STZ in NWT (Regan et al., 2016) together suggest a rifted margin likely was situated along the southeast Rae craton. The rift margin developed prior to sedimentation and subsequent tectonometamorphism related to the Taltson magmatic zone, or collision of the Hearne craton along the STZ (Berman et al., 2007; Martel et al., 2008; Bethune et al., 2010, 2013). West of our study area, the southwest Rae craton has a similar, but more poorly constrained rifting event between 2.34 and 2.13 Ga, prior to closure of the Rutledge River basin and accretion of the Buffalo Head and Slave cratons at ca. 2.0–1.93 Ga (Bostock and van Breeman, 1994; McDonough et al., 2000). Taken together, these results support a model of widespread sedimentation along the southeastern and southwestern margins of the south Rae craton that followed a rifting event at ca. 2.2–2.1 Ga (Fig. 11A).

**WLSz Tectonometamorphic History from 1.93 to 1.86 Ga**

Our data show that metamorphism of WLSz units and hanging-wall units of the Snowbird domain occurred at 1.9 Ga, allowing us to compare the tectonometamorphic history associated with the WLSz at middle- to lower-crustal levels. The primary metamorphic assemblages (M2) in
WLsz mafic granulites contain Grt + Cpx + Pl + Qz + Ilm ± Opx, which are commonly ascribed to the mid- to high-pressure granulite field (Patton, 2003), and are consistent with data from Krikorian (2002), who reported a peak P-T estimate of ~1.1 GPa and 900 °C for the northern Wholdaia Lake granulites. Since these assemblages define the S_{T1} foliation, pervasive shearing must have occurred at granulite-facies conditions prior to static recrystallization and development of granoblastic textures (Fig. 4D). Within sample 15ET249, the coarse-grained Grt_1 + Cpx + Qz ± Pl assemblage provides evidence for an earlier high-pressure event (M1) not presently associated with a structural fabric. If all plagioclase in this assemblage is retrograde, as suggested here, then M1 may have equilibrated at pressures above ~1.2 GPa, which are characteristic of the eclogite facies (e.g., Patton, 2003). These two assemblages may have developed by (1) protracted growth of the coarse M1 assemblage at an earlier time (e.g., 2.55 Ga), followed by shearing and decompression to form the enveloping M2 assemblage (e.g., at 1.9 Ga), or (2) during a continuous process where high-pressure coarse M1 growth was followed immediately by shearing and decompression, forming the M2 assemblage. The low-U, bright-CL zircon overgrowths and rims in the mafic granulate samples (Figs. 6–7) are interpreted to represent high-grade metamorphic recrystallization at ca. 1.9 Ga that overprinted ca. 2.6 Ga igneous zircon. The metamorphic zircon growth was likely related to shearing on the WLsz and attendant cooling and possible decompression, which have been shown to be a viable mechanism for zircon growth in high-pressure mafic rocks (e.g., Kohn et al., 2015; Yakymchuk et al., 2017). We therefore interpret the S_{T2}-defining M2 assemblage to have formed at 1.9 Ga, which was also coeval with movement on other northeast-trending structures along the STZ. The age of the earlier M1 high-pressure assemblage remains unconstrained.

Within the hanging wall of the WLsz, sample 15ET273b contains a (Grt + Sil + Kfs)-bearing assemblage that is consistent with P-T conditions of ~800 °C and <0.9 GPa (Bucher and Grapes, 2011). Metamorphic zircon is characterized by abundant fir-tree and sector zones with low Th/U values (Fig. 9) and shallow HREE profiles (Fig. 10), indicative of growth in equilibrium with stable garnet. The 1915 ± 4 Ma (2σ) mean age (Thiessen et al., 2017) for these zircon domains likely corresponds
coupled with the dextral horizontal and x plane shear-sense indicators crosscutting dikes (samples 15ET253a, 15ET260d). Taken together, dis-

ST2-ultramylonites may have accommodated post–1.85 Ga exhumation ages across the WLsz suggest that amphibolite- to greenschist-facies focused deformation within the WLsz postdated the S T1 and ST2 defor-

3). These relationships suggest that younger than ca. 1.85 Ga and are dextrally transposed into the WLsz at the kilometer scale (Fig.

the Firedrake domain are associated with refolded magnetic anomalies resulting in their current configuration.

An important question is: Do the distinctly zoned structural and meta-
morphic facies within the WLsz (S1, S2-mylonites, S3-ultramylonites) preserve a record of punctuated or continuous deformation at progressively shallower crustal levels during exhumation? Field relations and timing constraints bracket formation of the first two events between ca. 1.90 and 1.86 Ga. The nature and age of the ultramylonites are more elusive; they are relatively thin (~1–5 km) and are observed north of Wholdaia Lake, where the WLsz bends to a more easterly strike (Figs. 2 and 3). South along the WLsz, the structure again deflects to more easterly strikes, and the WRz may deflect into the north-

1.85 Ga (Flowers et al., 2008) interpreted these upper and lower intercepts of 2.55 Ga to correspond to two metamorphic events, as previously interpreted by Mahan et al. (2008). Other mafic granulites from the Tantato domain also show similar U-Pb zircon arrays (e.g., Baldwin et al., 2003), where it is suggested that ca. 2.55 Ga upper intercepts may represent metamorphic recrystallization of the protolith.

In contrast, mafic granulite 15ET249 contains two texturally distinct high-grade assemblages that are remarkably similar to those of sample 02M133a in Mahan et al. (2008); however, it only contains one well-constrained metamorphic zircon growth episode at 1.9 Ga. Oscillatory zoned zircon cores, high Th/U values, and the presence of 2.6–2.55 Ga granitoïd and mafic magmatism throughout the south Rae craton (e.g., Hamer, 1997; Mahan et al., 2008; Davis et al., 2015; Regis et al., 2017b) lead us to suggest that the upper intercepts of Baldwin et al. (2003) and Flowers et al. (2008) may be better interpreted as igneous crystallization rather than metamorphic ages.

Although zircon does not record widespread 2.55 Ga regional metamorphism in the Snowbird domain (obtained by SHRIMP, Martel et al.,

Regional Exhumation Systems

High-pressure granulites of the south Rae craton have been previously documented to have been exhumed in multiple phases at ca. 1.9–1.85 Ga along major STZ–related structures (Mahan et al., 2006b; Regan et al., 2014), with the 1.85 Ga component involving east-directed Rae over Hearne craton displacement on the Legs Lake shear zone and Chipman shear zone. Other structures farther west from the Rae–Hearne boundary include the Striding mylonite zone, Cora Lake shear zone, and the WLsz, which record slightly older ca. 1.87 Ga normal-oblique displacement and are responsible for exhumation of the Snowbird, western Tantato, and Firedrake domains. The overall style and nature of the 1.9 Ga tectonism remain unclear; however, it appears that syn– to post–1.9 Ga normal-oblique sense displacement on multiple crustal-scale structures within the south Rae craton was actively exhuming high-pressure rocks to 0.8–1.0 Ga conditions.

Continued exhumation along the STZ (Mahan et al., 2006b) brought southeastern Rae craton rocks to ~0.5 Ga at 1.85 Ga. This must have been followed by rapid normal faulting on the Striding mylonite zone to explain the removal of 15–20 km of crust before brittle-dextral faulting and deposition of the ca. 1.83 Ga subaerial Baker Lake Group volcanic rocks on granite basement at Snowbird (Fig. 3) and Kamilukjuak Lakes (e.g., Roscoe and Miller, 1986; Rainbird et al., 2005; Flowers et al., 2006b). Late extension, final erosion, and uplift were complete by ca. 1.74 Ga, with brittle faulting on the STZ and deposition of the younger Wharton Group in restricted subbasins (Hadlari and Rainbird, 2011). Detritus removed during the uplift events likely ultimately found its way into the alluvial–fluvial Nonacho-Thluicho (Aspler and Donaldson, 1985; Bethune et al., 2010) or Kiyuk Groups (Aspler et al., 2002; Davis et al., 2005) and intracontinental basins of the Dubawnt Supergroup (Rainbird et al., 2006).

Comparative Zircon Ages in the Southeastern Rae Craton

One of the puzzles from the regional studies of Martel et al. (2008), Davis et al. (2015), and our results herein is the lack of evidence for 2.55 Ga metamorphism, which is prevalent in the Tantato domain (e.g., Baldwin et al., 2006; Mahan et al., 2006a; Flowers et al., 2008; Dumond et al., 2015). Precise U-Pb thermal-ionization mass spectrometry (TIMS) zircon data from Tantato domain mafic granulites highlight clear discordia arrays with upper intercepts at ca. 2.55 Ga (Th/U > 0.2) corresponding to dark-CL oscillatory and sector zoned zircon and lower intercepts at ca. 1.9 Ga corresponding to bright-CL overgrowths and rims. Flowers et al. (2008) interpreted these upper and lower intercepts of 2.55 Ga and 1.9 Ga (for sample 02M133a) to correspond to two metamorphic events, as previously interpreted by Mahan et al. (2008). Other mafic granulites from the Tantato domain also show similar U-Pb zircon arrays (e.g., Baldwin et al., 2003), where it is suggested that ca. 2.55 Ga upper intercepts may represent metamorphic recrystallization of the protolith.

Within the Firedrake domain west of the WLsz, pervasive migmatiz-
tion and metamorphism between 1.85 and 1.81 Ga (Davis et al., 2015; Regis et al., 2017b) imply that the Firedrake domain underwent a younger exhumation history. Basement gneisses and mylonite within the Firedrake domain are associated with refolded magnetic anomalies and are dextrally transposed into the WLsz at the kilometer scale (Fig. 3). These relationships suggest that younger than ca. 1.85 Ga and more focused deformation within the WLsz postdated the S1 and S2 deformation dated herein. Therefore, field relationships and metamorphic ages across the WLsz suggest that amphibolite- to greenschist-facies S3-ultramylonites may have accommodated post–1.85 Ga exhumation of the Firedrake domain.

LITHOSPHERE | Volume 10 | Number 5 | www.gsapubs.org

658

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2008; Regis et al., 2017a) and is questionable in many Tantato domain rocks (obtained by TIMS; Baldwin et al., 2003; Flowers et al., 2008), a few Tantato domain units have been interpreted to have 2.57–2.55 Ga metamorphic zircon obtained by LA-ICP-MS (Regan et al., 2017). One explanation for the disparate zircon record between the studies may lie in the method used. The LA-ICP-MS process may yield apparently concordant results for short mixing lines between 2.6 and 1.9 Ga, whereas more precise methods yield predominantly clear two-point isochrons. For example, an Archean garnetiferous orthogneiss from the Tantato domain recently analyzed by U-Pb zircon SHRIMP (Ashton et al., 2017b) yielded metamorphic zircon populations at 2.58 and 1.9 Ga; however, only the 1.9 Ga population contained high Hf/Yb values. This suggests that the younger zircon growth was associated with Yb depletion and probably garnet growth during high-grade metamorphism, whereas the older population, which overlaps in age with crystallization of basement protolith gneisses, may represent zircon growth during a thermal pulse related to documented magmatism. Regional 2.55 Ga metamorphism does not appear to be well documented by zircon ages; however, metamorphic monzinite in the Tantato domain (e.g., Baldwin et al., 2006; Mahan et al., 2006a) does yield concordant ages at ca. 2.55 Ga. This phenomenon can be explained by the fact that monzinite is a sensitive recorder of metamorphic reactions (e.g., Dumond et al., 2015) and is more likely to grow episodically throughout the P-T history of a rock compared to zircon growth (e.g., Taylor et al., 2016).

The general model for deep-crustal residence between 2.55 and 1.9 Ga for the Tantato domain granulites and 1.9 Ga high-pressure metamorphism due to igneous intrusion (e.g., Flowers et al., 2008; Mahan et al., 2008) is inconsistent with the evolution of the Snowbird-Dodge domains and upper Murmac Bay Group metasedimentary rocks; the latter two were unconformably deposited on high-grade basement gneisses that were exhumed to the surface before ca. 1.98 Ga and subsequently reburied to granulate-facies conditions at ca. 1.93 Ga (Bethune et al., 2013). Based on these results, regions north and west of the Tantato domain appear to have been affected by a widespread 1.93–1.90 Ga loading event, and it stands to reason that the Tantato domain also underwent this history. However, until younger than 2.0 Ga metasedimentary rocks are documented in the Tantato domain, this inferred history remains speculative.

SUMMARY AND CONCLUSIONS

We have refined the timing of deposition of Paleoproterozoic metasedimentary rocks in the south Rae craton to 1.98–1.93 Ga, constrained by detrital zircon analyses and timing of oldest metamorphic zircon growth. These metasedimentary units may have been sourced from localized uplift and erosion of the Taltson magmatic zone (Fig. 11A) prior to peak metamorphism therein at 1.93 Ga. Metamorphic zircon growth occurred within these metasedimentary rocks at 1.93–1.88 Ga and within high-pressure mafic granulites at a deeper structural level at ca. 1.9 Ga (Fig. 11B). Significantly, we have documented a newly discovered 300-km-long, crustal-scale shear zone, the WLsz, which partially accommodated exhumation of high-pressure mafic-granulites at 1.9–1.86 Ga during normal-oblique sense displacement (Fig. 11C). Shearing initiated at granulate-facies conditions and was followed by amphibolite-facies shearing and mylonite development (Fig. 11D), constrained by variably deformed crosscutting 1.86 Ga dikes (Fig. 8). Synchronous displacement along the Coral Lake shear zone and WLsz partially exhumed large regions of middle to lower crust in the southeastern Rae craton between 1.90 and 1.85 Ga by normal-oblique sense displacement, while the Legs Lake shear zone accommodated exhumation through thrusting and erosion. Lower-grade shearing along more easterly trending segments of the WLsz may correspond with post–1.86 Ga dextral offsets associated with the Grease River shear zone and other congruent structures.

Our results document a significant 1.93–1.90 Ga tectonometamorphic burial event, but evidence for metamorphism at 2.55 Ga is lacking, and so the extent and significance of this older event remain elusive within the Snowbird domain. Furthermore, although the Taltson magmatic zone (1.93 Ga), which flanks the west and south side of the south Rae craton, and the Snowbird orogen (ca. 1.90 Ga), which lies along its eastern margin, appear closely linked in time, the relationship and possible interaction between these Paleoproterozoic orogenic systems remain to be resolved.

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