Component geochronology in the polyphase ca. 3920 Ma Acasta Gneiss

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Abstract

The oldest compiled U–Pb zircon ages for the Acasta Gneiss Complex in the Northwest Territories of Canada span about 4050–3850 Ma; yet older ca. 4200 Ma xenocrystic U–Pb zircon ages have also been reported for this terrane. The AGC expresses at least 25 km² of outcrop exposure, but only a small subset of this has been documented in the detail required to investigate a complex history and resolve disputes over emplacement ages. To better understand this history, we combined new ion microprobe 235,238U–207,206Pb zircon geochronology with whole-rock and zircon rare earth element compositions ([REE]zirc), Ti-in-zircon thermometry (Ti_xln) and 147Sm–143Nd geochronology for an individual subdivided 1/C24 cm² slab of Acasta banded gneiss comprising five separate lithologic components. Results were compared to other variably deformed granitoid-gneisses and plagioclase-hornblende rocks from elsewhere in the AGC. We show that different gneissic components carry distinct [Th/U]zirc vs. Ti_xln and [REE]zirc signatures correlative with different zircon U–Pb age populations and WR compositions, but not with 147Sm–143Nd isotope systematics. Modeled D^3D^6zirc from lattice-strain theory reconciles only the ca. 3920 Ma zircons with the oldest component that also preserves strong positive Eu anomalies. Magmas which gave rise to the somewhat older (inherited) ca. 4020 Ma AGC zircon age population formed at ~IW (iron–wüstite) to ~FMQ (fayalite–magnetite–quartz) oxygen fugacities. A ca. 3920 Ma emplacement age for the AGC is contemporaneous with bombardment...
of the inner solar system. Analytical bombardment simulations show that crustal re-working from the impact epoch potentially affected the precursors to the Acasta gneisses. © 2014 Elsevier Ltd. All rights reserved.

### 1. INTRODUCTION

What is known about the Hadean–Archean transition on Earth at ca. 4 billion years ago comes from (i) analysis of the oldest detrital zircons in Western Australia (e.g. Harrison, 2009) and North America (e.g. Maier et al., 2012), (ii) abundances of anomalous daughter products of the extinct radionuclides $^{146}$Sm (Caro et al., 2006; Rizo et al., 2011; O’Neill et al., 2012) and $^{182}$Hf (Willbold et al., 2011; Touboul et al., 2012) in ancient rocks, (iii) geophysical models (e.g. Davies, 2007; O’Neill et al., 2013), and (iv) supporting information gleaned from lunar samples, meteorites, robotic space missions and astrophysical observations (Taylor, 2001). It is vital to better understand the physical and chemical regimes operative on Earth in the late Hadean because, it has been argued, life had already emerged by that time (e.g. Mojzsis et al., 1996; cf. Lepland et al., 2005; McKeegan et al., 2007). If the first biosphere came into being in the Hadean it would have done so in the context of asteroid and comet bombardment that should have strongly modified the early crust and by extension, affected the nascent biosphere (Abramov and Mojzsis, 2009). Yet, only after the onset of the Eoarchean at about 3850 Ma (Bleeker, 2004) does the geologic record yield direct examples of hydrosphere-crust interactions via the preservation of the oldest rocks of sedimentary protolith (Nutman et al., 1997; Cates and Mojzsis, 2006; Manning et al., 2006; Moszewska et al., 2012).

What became of the first crust? Indications are that some fraction of primordial enriched and depleted crust and mantle domains lingered on and continued to affect the compositions of rocks well into the Archean (Bennett et al., 2007; O’Neil et al., 2008, 2012; O’Neill et al., 2013; Rizo et al., 2011, 2013; Guitreau et al., 2013; Roth et al., 2013). This was still possible because the last vestiges of the Hadean world were not yet wholly recycled and geochemically homogenized with Earth’s bulk silicate reservoirs (reviewed in Caro, 2011).

In light of these observations it can be postulated that between the formation of the planet and the start of the preserved rock record the Hadean “primordial” crust hosted a wide array of rock types ranging from the products of ultramafic (komatite) to mafic (basalt) melts, to intermediate andesitic and other enriched felsic rocks such as tonalities, granodiorites, as well as granites (e.g. Harrison, 2009). Consistent with this view, the sedimentary products of crust–hydrosphere interactions and recycling in the first half billion years appears to have given rise to metaluminous to peraluminous melts as evidenced by the chemistry of the oldest terrestrial zircons (Maas et al., 1992; Hopkins et al., 2008, 2010). Indications are that the redox state of the Hadean-Eoarchean mantle was near present oxygen fugacities (e.g. Delano, 2001), and more cryptic evidence has been presented that the crust was affected the so-called Late Heavy Bombardment (LHB) of the inner solar system (Trail et al., 2007a; Abbott et al., 2012; Abramov et al., 2013; Bell and Harrison, 2013). A better picture about the true diversity of rock types and fate of the Hadean lithosphere necessitates expanded geochemical and geochronological analyses of the oldest gneissic complexes. One such example is the Acasta Gneiss Complex (hereafter, AGC) situated at the western margin of the Slave craton (Bowring et al., 1989) approximately 300 km north of Yellowknife (town) in the Northwest Territories of Canada (Fig. 1).

In the following sections, we show how comparison of the whole rock and zircon trace element geochemistry of different Acasta gneissic components – mapped and sampled individually at the hand sample scale – can be combined with zircon imaging, Ti-in-zircon thermometry (Ti$_{an}$), crystal chemistry and REE modeling to distinguish between inherited, igneous, metamorphic, and altered zircons to derive an internally consistent emplacement age estimate for the different AGC components. The oldest of the Acasta gneisses (3850–4200 Ma) overlap in age with the LHB, so we evaluate the possibility that the AGC preserves evidence for global scale impacts as epitaxial zircon overgrowths, and make recommendations for future work (Section 5.2). Our results elucidate the different emplacement ages of separate Acasta components, reveal the inheritance of older ages in younger lithologies that potentially carry information about the primordial Earth, uncover clues about the oxidation state of Acasta magmas, and clarify the protracted history of polyphase metamorphism and equilibration that disrupted the $^{147}$Sm–$^{143}$Nd isotopic system. This level of documentation is necessary to pave the way for new studies to seek for components of possible sedimentary protolith (paragneisses) that can help us make direct inferences about the nature of the surface zone of the Hadean Earth. It also sets the stage for Lu–Hf measurements (Guitreau et al., in press), and other emerging isotopic studies such as $^{142}$Nd/$^{144}$Nd and $^{182}$W/$^{184}$W, that will be performed on the same rocks.

#### 1.1. Geologic background

The Acasta complex is captured within the pre-2800 Ma basement of the Slave Craton and overlain by supracrustal rocks of the ca. 2700 Ma Yellowknife Supergroup. It was discovered in the early 1980s over the course of regional mapping directed by the Geological Survey of Canada (King, 1985; St-Onge et al., 1988 and references therein). The AGC has since been the object of occasional geochronological investigations which sought to better resolve relationships between the various gneisses to explore the state of earliest Earth’s crust and mantle (e.g. Bowring and Housh, 1995; Bleeker and Stern, 1997; Bowring and Williams, 1999;
Iizuka et al., 2007). That past work concluded that the oldest of the AGC components formed sometime between about 4050 and 3920 Ma (Stern and Bleeker, 1998) notwithstanding evidence from ca. 4200 Ma xenocrystic zircons for the assimilation of even older crust (Iizuka et al., 2006).

It was evident soon after their discovery that a pervasive and protracted history of thermal overprinting and crustal re-working has severely complicated what can be deduced about the origin(s) of the various AGC gneisses and what might be preserved from the time of their formation. Briefly, early magmatism that emplaced the oldest Acasta components was followed by Eoarchean thermal events at ca. 3850–3720 Ma and 3660–3590 Ma, which in turn were accompanied by igneous intrusions, crustal anatexis and recrystallization of the older units (reviewed in Bowring and Housh, 1995; Iizuka et al., 2006, 2007). Later intrusions and modifications continued into the Archean, including ca. 3300 Ma granitic sheets that resulted in leucosome development and minor neosome zircon growth. Moderately deformed ca. 2875 Ma granites are also present (Stern and Bleeker, 1998). The deposition of the Yellowknife Supergroup supracrustals over the Eoarchean basement was accompanied by a new period of granitic magmatism at ca. 2700 Ma, followed by the intrusion of multiple Paleoproterozoic dike swarms and ca. 1800 Ma syenites. This history culminated with regional metamorphism related to the Wopmay tonalite between about 1800 and 1900 Ma (Hodges et al., 1995; Bleeker and Stern, 1997; Sano et al., 1999; Iizuka et al., 2007).

1.2. Component geochronology of the Acasta gneisses

In the first zircon geochronological studies of the AGC, age estimates for its emplacement centered around 3960 Ma
This value was later refined upwards to 4020–4050 Ma based on a much larger zircon geochronology inventory collected from a wider spectrum of whole rock samples, and focused on more high-resolution multiple ion microprobe spot analyses of different domains within the oldest individual grains discovered (Bowring and Housh, 1995; Bleeker and Stern, 1997; Stern and Bleeker, 1998).

Insomuch as these various results point to the existence of different age domains within the AGC and within individual samples, it now makes sense to extend our investigation of the heterogeneity of different age components within AGC gneisses to the centimeter-scale of an individual sample as opposed to drawing inferences about an ill-defined terrane strictly from bulk chemical analysis of polyphase rocks. Highly transposed and overprinted banded gneisses can be dissected into different lithological components and analyzed individually within the same rocks (see, for example, Kroner et al., 1989; Kinny and Nutman, 1996; Nutman et al., 2007; Gordon et al., 2013). It is important to note that herein we make no distinction between the term “lithology” and “component” at the hand sample scale (<0.1 m).

Although long lauded as the “world’s oldest rocks” no comprehensive geochronological data sets for the Acasta gneisses have been available in the published literature. Furthermore, rather than being restricted to a small area of the Western Slave Province in the vicinity of the original discovery site, new mapping studies show that it exceeds 25 km<sup>2</sup> across several domal basement antiforms (see Supplementary Fig. S1). Only a few small areas of one of these domes have been investigated in the fine detail required to resolve complex field relationships to the geochronology and to understand the extent of compositional variations of the gneisses (e.g. Iizuka et al., 2006, 2007).

Conventional 147<sup>Sm</sup>-143<sup>Nd</sup> studies showed that those gneisses in the AGC thought to be older than about 3600 Ma preserve a wide range of both positive and negative ε<sub>Nd</sub> values (Bowring et al., 1990; Bowring and Housh, 1995). It was first postulated that these heterogeneities reflected early inhomogeneous mantle domains, and in particular that the negative ε<sub>Nd</sub> values point to the prior existence of much older evolved crust. Preliminary searches for anomalous 142<sup>Nd</sup>/144<sup>Nd</sup> that might be expected in rocks with such an ancient heritage have so far been unsuccessful (Caro et al., 2006). An 147<sup>Sm</sup>-143<sup>Nd</sup> “age” derived from a sample with ca. 4000 Ma zircon ages and reported in Bowring and Housh (1995) is only 3156 Ma. This estimate was later re-examined by Moorbeth et al. (1997) with the addition of further AGC samples of unknown geochemical affinity or age which provided a larger range of Sm/Nd values, and revised to 3371 ± 59 Ma based on a statistically uncertain scatterchron (MSWD of 9.2). Consequently, competing interpretations of published Sm-Nd data so far have led many workers to embrace the view that although some parts of the AGC may indeed host “the world’s oldest rocks”, their isotope systematics as a whole have been reset such as to preclude any possibility of inferring Hadean mantle heterogeneities.

Another serious complicating factor is that U–Pb zircon geochronological studies of multiply-deformed polyphase ancient gneiss terranes like the AGC face the dilemma of confidently identifying the original igneous zircon population of a rock that can be used to establish emplacement ages via U–Pb zircon geochronology. What constitutes either inherited zircons from incomplete assimilation of older material, and/or younger zircons grown under metamorphic conditions in complex and ancient poly-metamorphic terranes, is an old debate (e.g. Mojzsis and Harrison, 2002; Whitehouse and Kamber, 2005; Manning et al., 2006). As such, more robust assessments of the results of zircon U–Pb age studies from detailed mineral chemistry arguments (Section 5.1.2) are necessary because, as mentioned above, the outcomes of zircon geochronology studies are vital to the interpretation of other radiogenic systems.

2. LOCATIONS AND DESCRIPTIONS OF SAMPLES

A complete list of the AGC samples collected for this project, along with location information, is provided in Table 1. In Table 2 whole rock geochemical data are reported for the various lithologies (1–5) identified in sample BNB99-151B.

2.1. Polyphase banded gneiss (BNB99-151B)

Gneiss sample BNB99-151B is from the same general locality as SAB94-134 reported in Bowring and Williams (1999) and is within the so-called “layered gneiss series” of Iizuka et al. (2007; Fig. 1). This particular rock was chosen for study because previously reported zircon geochronology for it yielded concordant U–Pb ages up to about 4030 Ma (Bowring and Williams, 1999). Our sample comprises a 30 x 20 x 4 cm polished slab of highly transposed and tectonized banded grey gneiss (Fig. 2) with mm- to cm-scale banding typified by either mafic (labeled as Lithology 1) or felsic (Lithology 2) minerals with metamorphic textures reminiscent of stromatic migmatites. The banded gneiss is cut by a low-angle finer-grained dark grey orthogneiss (Lithology 3) that in turn hosts small melt segregations (Lithology 4). Finally, a coarse-grained felsic vein (Lithology 5) transects both Lithologies 1 and 2.

2.2. Granitoid gneisses (AG09008, AG09009, AG09014, -15, -16, -17)

Rock samples AG09008 and AG09009 were collected from what was judged to be near the original sampling sites (e.g. “BGXM”) of Bowring et al. (1989). Samples AG09014 through -17 are from the vicinity where Iizuka et al. (2006) reported a tonalitic gneiss that hosts ca. 4200 Ma xenocrystic zircons. It is also from this area that we deduced samples SAB91-37 and -38 of Bowring and Housh (1995) were collected which yielded abundant ca. 4000 Ma zircons. Guitreau et al. (2012) also documented ancient U–Pb ages based on zircon analyses of one of these same samples. Both AG09008 and AG09017 are very dark grey (barely 10% quartz), hornblende- and biotite-rich foliated tonalite gneisses that contain garnet in places (sub-sample AG09008g in Guitreau et al., in press). Sample AG09017 displays narrow plagioclase and quartz banding, whereas
AG09008 is so finely banded that it looks almost massive. Gneisses AG09014 through AG0916 are variably banded biotite-bearing tonalitic gneisses that comprise part of the “felsic gneiss series” of Iizuka et al. (2007). Granitoid gneiss AG09009 is more felsic than the other samples, and contains some K-feldspar and preserved granitic textures.

### Table 1
Whole rock geochemistry of select samples from the AGC. Major elements are reported in wt.% oxides and trace elements in ppm.

<table>
<thead>
<tr>
<th>Sample Code</th>
<th>Lithology</th>
<th>Latitude</th>
<th>Longitude</th>
<th>SiO2 (wt.% oxides)</th>
<th>TiO2 (wt.% oxides)</th>
<th>Al2O3 (wt.% oxides)</th>
<th>Fe2O3 (wt.% oxides)</th>
<th>MnO (wt.% oxides)</th>
<th>MgO (wt.% oxides)</th>
<th>CaO (wt.% oxides)</th>
<th>Na2O (wt.% oxides)</th>
<th>K2O (wt.% oxides)</th>
<th>P2O5 (wt.% oxides)</th>
<th>LOI (wt.% oxides)</th>
<th>Total (wt.% oxides)</th>
</tr>
</thead>
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<td>AG09008</td>
<td>Si-poor tonalite gneiss</td>
<td>65° 10.118’ N</td>
<td>115° 33.549’ W</td>
<td>71.62</td>
<td>0.37</td>
<td>14.30</td>
<td>1.69</td>
<td>0.20</td>
<td>0.79</td>
<td>1.72</td>
<td>0.79</td>
<td>1.72</td>
<td>0.79</td>
<td>1.72</td>
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<tr>
<td>AG09009</td>
<td>Si-rich tonalite gneiss</td>
<td>65° 10.213’ N</td>
<td>115° 33.761’ W</td>
<td>70.03</td>
<td>0.41</td>
<td>15.91</td>
<td>2.81</td>
<td>0.03</td>
<td>0.17</td>
<td>3.89</td>
<td>0.39</td>
<td>1.94</td>
<td>0.12</td>
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<td>AG09014</td>
<td>Intermediate tonalite gneiss</td>
<td>65° 9.615’ N</td>
<td>115° 32.839’ W</td>
<td>67.20</td>
<td>0.63</td>
<td>15.43</td>
<td>4.16</td>
<td>0.04</td>
<td>0.16</td>
<td>7.20</td>
<td>0.59</td>
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<td>115° 32.839’ W</td>
<td>66.23</td>
<td>0.39</td>
<td>16.99</td>
<td>3.36</td>
<td>0.04</td>
<td>0.14</td>
<td>5.07</td>
<td>0.89</td>
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<td>0.86</td>
<td>0.43</td>
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<td>1.01</td>
<td>13.63</td>
<td>1.77</td>
<td>0.33</td>
<td>1.52</td>
<td>13.15</td>
<td>1.52</td>
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<td>65° 10.050’ N</td>
<td>115° 33.549’ W</td>
<td>44.42</td>
<td>2.99</td>
<td>17.47</td>
<td>7.22</td>
<td>0.29</td>
<td>5.65</td>
<td>22.47</td>
<td>5.65</td>
<td>9.00</td>
<td>5.65</td>
<td>1.56</td>
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<td>Mafic gneiss</td>
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<td>115° 33.803’ W</td>
<td>44.50</td>
<td>1.86</td>
<td>17.82</td>
<td>7.22</td>
<td>0.17</td>
<td>4.99</td>
<td>2.22</td>
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<td>7.22</td>
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<td>4.99</td>
<td>2.22</td>
<td>4.99</td>
<td>1.56</td>
<td>98.3</td>
</tr>
</tbody>
</table>

1 Predicted Th/U ratio zircon crystallizing from a melt with the Th/U ratio of the rock. Assumes 0.2 partition coefficient.
2 Calculated M value for zircon saturation, using the formula of Watson and Harrison (1983).
3 Zircon saturation temperature for each lithology, using the formula of Watson and Harrison (1983).
Table 2
Whole rock geochemistry of lithologies from BN99-151B. Major elements are reported in wt.% oxides and trace elements in ppm.

<table>
<thead>
<tr>
<th>Sample</th>
<th>A</th>
<th>B</th>
<th>D</th>
<th>E</th>
<th>F</th>
<th>G</th>
<th>H</th>
<th>I</th>
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<th>M</th>
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<td>12</td>
<td>13</td>
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1 Predicted Th/U ratio zircon crystallizing from a melt with the Th/U ratio of the rock. Assumes 0.2 partition coefficient.
2 Calculated M value for zircon saturation, after Watson and Harrison (1983).
3 Zircon saturation temperature for each lithology, after Watson and Harrison (1983).

2.3. Plagioclase-hornblende schists (AG09017N and AG09020)

Sample AG09017 N was tentatively described during our field mapping as a “gabbroic schist” based on texture and field association; it hosts nearly equal parts hornblende and plagioclase with a marked tectonic fabric expressed in the strong foliation. This rock was collected just to the northwest of AG09014–017 described above. A similar rock (AG09020) comes from the vicinity of AG09008 on an island at the discovery site of King (1985) in the Acasta River in Fig. 1. Both “leucogabbros” (“l.s.”) are from places where
previous field observations have shown that several generations of intermediate to felsic granitoid gneisses enclose and dismember the hornblende-plagioclase schists (Stern and Bleeker, 1998). Bleeker and Stern (1997) used the cross-cutting field relations to argue that these are the oldest rocks of the AGC. If this interpretation is correct, the Acasta "leucogabbros" represent Earth’s oldest known poly-mineralic assemblage. From that observation alone, they warrant further investigation.

3. METHODS

3.1. Whole rock geochemistry

Selected whole-rock samples were crushed to fine powders at the Department of Geological Sciences of the University of Colorado Boulder (CU) in ceramic mortars pre-cleaned with quartz sand and subsequently conditioned with small sample aliquots before powders were made of the main sample masses. Un-powdered rock splits were set aside for archival purposes as well as for petrographic thin section preparations (see below). Care was taken to avoid contact with metal hammers, saws or other dividing apparatus that could compromise future analytical work on these samples. Splits from homogenized powders were further subdivided for archival storage in our sample library (~50% of the volume) and the remainder was reserved for major-, minor-, and trace element geochemistry, and for other on-going studies. Whole rock geochemical analyses were performed at the Centre de Recherches Pétrographiques et Géochimiques (CRPG) SARM Facility in Nancy (France).
3.2. Mineral chemistry

Thin sections used for in situ U–Pb zircon geochronology were prepared for electron microprobe elemental analysis. Individual mineral analyses on plagioclase and biotite grains were performed on the CU JEOL JXA-733 electron microprobe at 15 kV and 20 nA on a defocused beam (5 μm). Standardization was via natural minerals: Amelita albite for Na and Si; anorthite for Ca and Al (USNM #137041); Microcline for K (USNM #143966); Springwater olivine for Mg (USNM #2566); Fayalite for Fe (USNM #85276); Garnet P-130 for Mn; and Ilmenite for Ti (USNM #96189). All analyses were calculated using the integrated CITZAF matrix correction routines (Armstrong, 1988).

3.3. U–Pb zircon geochronology

Subdivided whole-rock samples reserved for zircon extractions were cut with diamond saws, crushed, sieved, passed through heavy liquids in stages, and processed through a Franz magnetic separator following our standard techniques (Cates et al., 2013). Zircons hand-selected from the densest nonmagnetic fractions were cast in epoxy molds along with grains of zircon standard AS3 (Paces and Miller, 1993; Black et al., 2003; Schmitz et al., 2003). The mounts were polished in stages to expose zircon cores, cleaned in 1 N HCl solution to reduce common Pb contamination, rinsed and ultrasonically cleaned in ultrapure water, oven dried in 4 N HCl with a few drops of concentrated HClO4 and of no consequence for the age constraints derived here. After complete dissolution of the rock powders, samples were passed through heavy liquids in stages, and processed through high-pressure Teflon bombs at 180 °C. Separate aliquots of sample 5, the result is virtually identical to that obtained using the classic dissolution technique, while sample 2.3 is slightly less radiogenic (Section 4.4). This small difference may result either from imperfect zircon dissolution or from powder heterogeneity. In any case, these differences are minor and of no consequence for the age constraints derived here. After complete dissolution of the rock powders, samples were dried in 4 N HCl with a few drops of concentrated HClO4 to ensure complete elimination of fluoride precipitates. The REEs were then separated from the rock matrix using Eichrom TRU-Spec resin and further purification of Sm and Nd from the other lanthanides was achieved by one pass through Eichrom Ln-Spec columns. The procedural blanks were 100 pg for Nd and 30 pg for Sm, and the recovery yields were typically 70–80% for both Sm and Nd.

Isotope analyses were conducted on aliquots of 500 ng Nd and 100 ng Sm on a Finnigan MAT262. Both Sm and Nd were loaded in HCl and H3PO4 on zone-refined Ta filaments and measured as positive ions using a double Ta–Re microprobe output data reduction and served as input to the ISOPLOT® v3.7 software package (Ludwig, 2008) to generate cumulative probability and concordia plots, and to model concordia, discordia intercept and weighted mean ages. Errors for individual age analyses herein are reported at the ±1σ level, and at the 2σ level for the concordia, discordia intercept and weighted mean ages. Individual geochronology analysis points are provided in the Supplementary Table T1 and spot locations are shown in Supplementary Fig. S2.

3.4. Zircon trace element geochemistry

All zircon [REE]<sub>light</sub> and Ti<sub>YH</sub> analyses were made on the UCLA Cameca ims1270 ion microprobe following the approach described by Schmitt and Vazquez (2006). Analysis spots were ~25 μm in diameter and positioned to overlap previous geochronology analysis points. Both REE and Ti intensities were normalized to 10<sup>6</sup>Si<sup>6+</sup> and corrected for interfering oxides. The trace element content of NIST standard glass 610 was used to determine the analytical sensitivity (Pearce et al., 1996), and Ti concentrations were internally normalized to a value of 6.3 ± 0.1 ppm for zircon standard SL13 (Harrison and Schmitt, 2007). Instrument stability was monitored by periodic REE and Ti measurements on standard zircon 91500. Two-dimensional kernel density estimates plots were constructed of the zircon trace element data (Section 5.1.3) as a quantitative means to evaluate different compositional populations.

3.5. Sm–Nd geochronology

Twelve sub-samples of mapped lithology components 1–5 were separated from banded gneiss sample BNB99-151B by diamond saws and analyzed for 146<sup>Sm</sup>–143<sup>Nd</sup> at CRPG (Nancy, France). Powder aliquots of 80–100 mg were mixed with a 146<sup>Sm</sup>–150<sup>Nd</sup> spike and dissolved in concentrated HF–HNO<sub>3</sub> in a temperature-controlled oven at 120 °C. Separate aliquots of sample 5 and 2.3 were dissolved in high-pressure Teflon bombs at 180 °C for 1 week. For sample 5, the result is virtually identical to that obtained using the classic dissolution technique, while sample 2.3 is slightly less radiogenic (Section 4.4). This small difference may result either from imperfect zircon dissolution or from powder heterogeneity. In any case, these differences are minor and of no consequence for the age constraints derived here. After complete dissolution of the rock powders, samples were dried in 4 N HCl with a few drops of concentrated HClO<sub>4</sub> to ensure complete elimination of fluoride precipitates. The REEs were then separated from the rock matrix using Eichrom TRU-Spec resin and further purification of Sm and Nd from the other lanthanides was achieved by one pass through Eichrom Ln-Spec columns. The procedural blanks were 100 pg for Nd and 30 pg for Sm, and the recovery yields were typically 70–80% for both Sm and Nd.
4. RESULTS

4.1. Protolith assignments based on whole rock major-, minor- and trace element geochemistry

4.1.1. Granite-granitoid gneisses

Based on major-element compositions, the granitoid gneisses in our sample suite can be separated into three groups: (i) a “silica-poor” tonalitic group, (ii) an “intermediate” tonalitic group, and (iii) a “granitic” group. In general, our samples have typical tonalite-trondhjemit-granodiorite (TTG) compositions, including LREE enrichments and HREE depletions (normalized to chondrite), with small negative Eu anomalies (Fig. 3a) and negative Nb and Ti anomalies relative to primitive mantle (Fig. 3b).

The “silica-poor” group (AG09008 and AG09017) has SiO2 < 60 wt.%, elevated Fe2O3 (~12 wt.%) and is low in Na2O and K2O (~4 wt.%) total when compared to CaO (~7 wt.%) Table 1). It also has relatively flat chondrite-normalized REE patterns so that (Ce/Yb)CN ~ 3, and weakly negative Nb (Nb/Nb* ~ 0.4) and Ti (Ti/Ti* ~ 0.5) anomalies.

The “intermediate” group (AG09014–AG09016) has higher SiO2 contents (65–70 wt.%) and lower Fe2O3 (<5.0 wt.%) and CaO (~3.0 wt.%) than its more mafic counterparts. It also has enhanced LREE/HREE ((Ce/Yb)CN ~ 25) and more pronounced Nb anomalies (~0.09) compared to the more mafic tonalities.

Granitic gneiss AG09009 is silica-rich (71 wt.%) and has the highest K2O content of any of the TTG gneisses sampled thus far (3.6 wt.%) Table 1). It also has the strongest LREE/HREE ratio ((Ce/Yb)CN ~ 70) and strong Nb (0.06) and Ti (0.48) anomalies.

4.1.2. Hornblende-plagioclase schists

Both hornblende-plagioclase rocks analyzed in this study (AG09017N and AG09020; Section 2.3) are silica poor (SiO2 ~44 wt.%) with elevated Fe2O3 but are relatively low in MgO (<5 wt.%), with Mg# (Mg/(Mg + Fe) × 100) around 27. Unlike the various granitoid gneisses investigated in this work, the hornblende-plagioclase rocks show comparatively flat LREE/HREE patterns ((Ce/Yb)CN ~ 2) and positive Eu anomalies (Eu/Eu* = 1.2–2.0) on chondrite-normalized plots (Fig. 3a). They also have subdued Nb anomalies (~0.5) but large positive Ti anomalies (~4) on a primitive mantle-normalized multi-element plot (Fig. 3b). These geochemical indicators (Table 1) suggest that the protolith of this rock was a cumulate.

4.1.3. Compositional heterogeneity in polyphase banded gneiss sample BNB99-151B

To probe each mapped component of our polished gneiss slab in detail, we used diamond saws to cut out a filament assembly. Typical running conditions for Nd and Sm were 0.5–1 × 10−11 A on 144Nd for 1 h 30 min and 0.5 × 10−11 A on 149Sm for 1 h, respectively. Isobaric interferences on 144Nd (from Sm) and 152Sm (from Gd) were found to be negligible for all measurements. Repeated analyses of the LaJolla standard over the course of these measurements gave 143Nd/144Nd = 0.511839 ± 0.03 ppm (2σ).

Fig. 3. Trace element geochemistry of AGC TTG and amphibolite gneisses other than sample BNB99-151B. (a) Anders and Grevesse (1989) chondrite normalized REE patterns. (b) McDonough et al. (1992) primitive mantle normalized multi-element plot.

5 cm wide section oriented perpendicular to the dominant layering, and subdivided each layer into labeled chips A–P (Fig. 2b). Only when sufficient sample was recovered for geochemical analysis could a particular component of BNB99-151B be used to relate to the geochemical analyses reported in Table 2. Lithologies 1, 2 and 3 were abundant enough that they could be analyzed over multiple subsamples to test for internal compositional variations within the component, whereas Lithology 4 was volumetrically too small to permit whole rock geochemical analysis, and there was only sufficient material for one geochemical assay of Lithology 5.

Although major element geochemistry and normative calculations for all lithologies within sample BNB99-151B are consistent with tonalitic (Lithologies 1–3) or trondhjemitic (Lithology 5) compositions (classification of Barker, 1979), trace element and textural evidence shows that the rocks have been substantially altered since their original formation. Plagioclase grains within Lithology 2 are concentrated in coarser bands bounded by (interstitial) biotite-rich layers (Lithology 1) that is strongly reminiscent
of stromatic migmatite banding. Conversely, it should be noted that the Archean and younger deformational history of the region (Section 1.1) suggests that much of the evidence for early deformation would have been overprinted and as such textural arguments as to origin are unconvincing. It is worth noting that plagioclase grains within Lithologies 1 and 2 are subhedral, whereas (rare) K-feldspar and quartz are dominantly anhedral, indicating that they crystallized in a loose framework of plagioclase. Biotite, while mostly euhedral to subhedral, can be observed in places to halo plagioclase grains, also indicating that plagioclase crystallized first.

Compared to average upper or lower crust, Lithologies 1 and 2 have strong positive Eu anomalies (Eu/Eu*: 2.1–4.7) and LREE over HREE enrichments, but are depleted in REE overall (Fig. 4a). Other key features of Lithologies 1 and 2 include LILE enrichments, elevated Zr contents (mostly 300–690 ppm) and moderate to strong negative Nb anomalies (Nb/Nb*: 0.18–0.70; Fig 4b).

Lithology 5 is a silicic (76 wt.% SiO₂) trondhjemitic dike that steeply cross-cuts the banding in Lithologies 1 and 2. This rock does not share the trace element characteristics of Lithologies 1, 2 and 5 as its overall REE content is not depleted, and has only a very small Eu anomaly. Nor is a strong Nb anomaly evident in Lithology 3, and the multi-element and REE plots closely resemble those for average mid- and upper-crust (Taylor and McLennan, 1985; Rudnick and Fountain, 1995) except for a pronounced enrichment in Zr (600–725 ppm; Fig. 4b).

4.2. U–Pb zircon geochronology

4.2.1. Granitoid gneiss AG09008

Although bulk analysis shows that this sample contains 216 ppm Zr (Table 1), its zircon yield was low with a mere 50 grains recovered from ~0.5 kg of crushed rock. The bulk trace element geochemistry of the sample has high Hf, but with low Lu and relatively low SiO₂. Zircons recovered are small (50–120 μm) with low aspect ratios (no more than 2:1). Most grains appear mottled in CL and BSE imagery and contain multiple small inclusions. None of the zircons from this sample which were imaged by BSE or CL display obvious oscillatory zoning (Supplementary Fig. S2), which leads to the first-order conclusion that zircons recovered from this sample are not part of the original igneous population. Our 25 analyses on 23 grains yielded three age populations, with one grain (duplicate analyses) at ca. 3805 ± 150 Ma. Because duplicate analyses yielded data that were ~12% discordant, we regard this result as an age estimate rather than a true age. Two younger populations were revealed at 3597 ± 17 (n = 4) and 3466 ± 23 (n = 7), and show evidence for minor recent Pb-loss such that analyses string out along a horizontal axis on the Tera–Wasserburg plot in Fig. 5a. Based on this observation, their ages are calculated as weighted means rather than isochrons. All other analyses yielded younger 206Pb/238Pb ages with significant (>20%) discordance.

4.2.2. Hornblende-plagioclase schists (“leucogabbros”) AG09017N, AG09020

Zircons are generally rare in gabbroic rocks, but it is possible for some to crystallize during initial emplacement (e.g. Grimes et al., 2007, 2009; Kaczmarek et al., 2008). From a compositional standpoint it resembles Lithologies 1 and 2, but with a slightly more sodic composition, and mafic phases are almost entirely absent. On both the normalized REE and multi-element plots, the same relative enrichments and depletions reported for Lithologies 1 and 2 are visible in this sample, but with about a 50% depletion in absolute concentrations.

Lithology 3 cross-cuts Lithologies 1 and 2 at a low angle and is less silicic (SiO₂: 55–61 wt.%) than these two, as well as fine-grained and melanocratic. It is also enriched in total alkalis relative to Lithologies 1 and 2 as reflected in a near twofold enrichment in CaO, which most likely is due to the dominance of hornblende as the mafic phase rather than biotite. Texturally, the rock is banded with a sense of foliation parallel to the regional trend. Plagioclase is interstitial to hornblende and biotite and there is little evidence for migmatization other than a small felsic (plagioclase + quartz) apophysis within the banding (Lithology 4). This rock does not share the trace element characteristics of Lithologies 1, 2 and 5 as its overall REE content is not depleted, and has only a very small Eu anomaly. Nor is a strong Nb anomaly evident in Lithology 3, and the multi-element and REE plots closely resemble those for average mid- and upper-crust (Taylor and McLennan, 1985; Rudnick and Fountain, 1995) except for a pronounced enrichment in Zr (600–725 ppm; Fig. 4b).
Field relations indicate that the AGC hornblende-plagioclase rocks pre-date the oldest generation in the Acasta gneisses (Section 2.3), which made this rock a special target for attempting zircon geochronology. The rocks are so poor in Zr (<40 ppm) that we chose to extract zircons from large sample volumes (>1 kg) in the quest for rare primary igneous zircons. Zircon yields for both units validated this approach as fewer than 20 grains from each sample were recovered. The grains are uniformly small (20–100 μm), anhedral to subhedral, with mottled or structure-less interiors as seen in BSE and CL images. Furthermore, CL shows that many of the grains have thin (<1 μm) and bright rim overgrowths, which are not obvious in the BSE images. The largest of these grains also typically have small (<5 μm) but abundant inclusions (Supplementary Fig. S2).

The 21 analyses from sample AG09020 on 10 grains yielded mostly young, discordant ages, although there appears to be a grouping of near-concordant analyses at about 3050 Ma (Fig. 5b). A single grain yielded three ages of ca. 3800 Ma, with the most concordant datum at 3770 ± 5 Ma. Similarly, sample AG09017N yielded only few concordant grains, all of which we interpret to be associated with metamorphic growth. Three concordant ages at 2618 ± 5 Ma came out of our 12 spot measurements on 9 grains and a near concordant group of four analyses was found at ca. 3500 Ma (Fig. 5c). If all but two highly discordant analyses are used in the regression, the data yield a concordia age with an (uncertain) upper intercept of 3776 ± 270 Ma and a lower intercept of 2606 ± 210 (MSWD = 1.7).

4.2.3. Banded gneiss BNB99-151B

Overall, zircons in BNB99-151B were generally larger than observed in our other AGC samples (200–600 μm) and have subhedral habit. The largest and oldest grains (>450 μm; ca. 4020 Ma) are, however, usually anhedral broken crystals with subdued CL zoning. Smaller zircons in this category have aspect ratios of 2:1 to 4:1 and tend to show relatively fewer small inclusions than the large zircons. Image analysis in CL and BSE shows that these zircons tend to have large structure-less or mottled cores mantled by thin rims. Rarely, faint- to well-defined internal oscillatory zoning was observed in some cores and rims, especially those from the ca. 3920 Ma population. Zircons from the nominally younger Lithologies 3 and 4 are smaller (<200 μm), have higher aspect ratios (4:1) and have well-defined internal sector zoning (Fig. 6; Supplementary Fig. S2).

![Fig. 5. Tera–Wasserburg concordia plots of zircon U–Pb zircon analyses from (a) tonalitic gneiss AG09008; (b) plagioclase-hornblende schist AG09020; and (c) plagioclase-hornblende schist AG09017N.](image)

![Fig. 6. Annotated CL images of representative population 1 and 2 zircons from sample BNB99-151B.](image)
When all analyses for all lithologies in this sample are grouped together, 150 spot measurements out of 217 on 157 separate grains are less than 15% discordant. On a cumulative probability plot (Fig. 7), the data reveal 6 apparent age “peaks” (C24/4010, 3920, 3880, 3750, 3550 and 3300 Ma); this distribution of ages is investigated further below, and a detailed explanation on how each zircon population was calculated as a separate grouping is presented in the Supplementary data file S3. A summary of the geochronological data is provided in Table 3.

The oldest population (here termed Population 1; Fig. 6) is composed of concordant analyses with an age of 4004 ± 13 Ma (MSWD of concordance + equivalence = 1.3). If less concordant analyses from the same zircons are included (n = 42), they describe a discordia line with an upper intercept at 4003 ± 26 Ma. Zircons from this population include some large rounded grains with little internal structure. Their Th/U values have a weighted mean of 0.85 ± 0.12. Population 1 zircons are present only in Lithologies 1 and 2 (Fig. 8a and b).

The next oldest population (Population 2) constitutes a majority our zircon results. The most concordant analyses center at 3920 Ma, and when all analyses are considered (n = 105) data yield discordia with an upper intercept of 3914 ± 20 Ma (MSWD = 2.6). The Th/U ratios for the most concordant grains (within 5% of concordia) have a weighted mean of 0.50 ± 0.13. Population 2 is present in Lithologies 1, 2 and 3.

Population 3 zircons occur in all lithologies and have an upper intercept discordia age of 3728 ± 13 Ma.
from the Acasta rocks are up to 6700 ppm (0.06–0.99), but generally the value is low with a weighted mean of 0.18 ± 0.07, whereas Population 3 zircons from Lithology 3 have higher Th/U values with a weighted mean of 0.88 ± 0.33 (Fig. 8c).

The youngest zircon population is also present in all lithologies and defines a discordia with an upper intercept at 3557 ± 20 Ma (MSWD = 1.8). The most concordant analyses from this group of zircons have a wide range of Th/U ratios from 0.15 to 1.13.

4.3. Zircon trace element geochemistry

A subset of zircons was chosen for trace element studies from previously analyzed concordant grains identified in sample BNB99-151B and part of Populations 1–3 as noted above. Some of these have younger rims that could be compared with zones that also preserve igneous textures in BSE and CL imagery. All REE analyses and Ti results translated into Ti\(^{39+}\) (Section 4.3.2) are reported in the Supplementary data , but will only be discussed in passing. A summary of zircon REE analyses is presented in Table 4.

4.3.1.2. Flat REE zircon cores.

Although the degree of enrichment varies greatly, [REE]\(_{\text{arc}}\) from the Acasta rocks are up to 6700 ppm ΣREE. All but one of the analyses we report here are super-chondritic and overall enrichments in HREE over LREE typical for zircon are evident. Together the grains preserve positive Ce anomalies, but have variable Eu anomalies. Based on combined geochronology and chondrite-normalized REE patterns, the zircons can be divided into three groups (Fig. 9): (i) HREE/LREE enriched zircon cores with either strong negative Eu anomalies or weak/absent negative Eu anomalies; (ii) those with flat REE zircon cores; and (iii) those with flat REE zircon rims. Analyses on spots with discordant ages from grains that also had concordant spot analyses were also taken and are reported in the Supplementary data

4.3.1.1. HREE/LREE enriched zircon cores. These zircons are largely from Population 1 (12 analyses on 10 grains), along with two grains arising from the Population 2 analyses (Sections 4.2.3). These are typified by HREE-enriched chondrite-normalized profiles ([La/Yb]

\(_{\text{CN}}\) < 0.022) and steep positive HREE slopes ([Gd/Yb]\(_{\text{CN}}\) = 0.02–0.09) with ΣREE concentrations of ~2000 ppm. The Eu anomalies from this group range from strongly negative to nearly flat (Eu/Eu\(^*\) = 0.084–1.02; Table 4, Fig. 9a). They may be further subdivided based on this chemical feature, since zircons with the strongest negative Eu anomalies also have strong positive Ce anomalies (Ce/Ce\(^*\) = 1.67–15.7). This group are also the least enriched LREEs (e.g. average La concentrations of 4.7 vs. 9.7 ppm) and correspond to zircons with the oldest ages (ca. 4020 Ma) within Population 1.

4.3.1.2. Flat REE zircon cores. This compositional grouping of zircons is found in Population 1 (n = 1), 2 (n = 4) and 3 (n = 1) grains. Although they are HREE-enriched (Table 4), they have a much more subdued chondrite-normalized slope than the high-HREE/LREE grouping described above (average [La/Yb]\(_{\text{CN}}\) = 0.10; average [Gd/Yb]\(_{\text{CN}}\) = 0.32). Overall, they are enriched in LREE and depleted in HREE compared to the high-HREE/LREE population,
and have elevated \( \Sigma \text{REE} \) (~2350 ppm). These zircons also have slightly positive Ce and Eu anomalies (Ce/\( \text{Ce}^{*} \) = 1.21–1.67; Eu/Eu* = 1.16–2.80; Fig. 9b).

4.3.1.3. Flat REE zircon rims. The zircons in this grouping are found in Populations 2 (\( n = 3 \)), 3 (\( n = 2 \)) and 5 (\( n = 1 \)). Their chondrite-normalized REE profiles are similar to the flat zircon cores with slightly more depleted LREE vs. HREE (average \((\text{La/Lu})_{\text{CN}} = 0.05\)), but are less enriched overall with an average \( \Sigma \text{REE} \) concentration of ~1200 ppm (Table 4). Relative to the “flat REE core” population (Section 4.3.1.2), the rims have both positive Ce and Eu anomalies, but the Ce anomalies are greater (Ce/\( \text{Ce}^{*} \) = 1.40–1.90) and the Eu anomalies are more subdued (Eu/Eu* = 1.08–1.62; Fig. 9b).

4.3.2. Zircon \( Ti^{4+} \) temperature estimates

The Ti-in zircon thermometer was developed by Watson and Harrison (2005), and further refined by Ferry and Watson (2007) based on improved thermodynamic analysis of Ti substitution. In short, the technique computes the temperature of crystallization of zircon for melts based on

### Table 4

Average chondrite normalized zircon REE analyses.

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<td>216</td>
</tr>
<tr>
<td>Gd 701</td>
<td>300</td>
<td>235</td>
<td>342</td>
</tr>
<tr>
<td>Tb 752</td>
<td>605</td>
<td>536</td>
<td>651</td>
</tr>
<tr>
<td>Dy 898</td>
<td>1027</td>
<td>958</td>
<td>1073</td>
</tr>
<tr>
<td>Ho 1273</td>
<td>1837</td>
<td>1766</td>
<td>1884</td>
</tr>
<tr>
<td>Er 1782</td>
<td>2693</td>
<td>2631</td>
<td>2734</td>
</tr>
<tr>
<td>Tm 2360</td>
<td>3625</td>
<td>3630</td>
<td>3621</td>
</tr>
<tr>
<td>Yb 3152</td>
<td>4618</td>
<td>4711</td>
<td>4556</td>
</tr>
<tr>
<td>Lu 4411</td>
<td>6126</td>
<td>6154</td>
<td>6107</td>
</tr>
<tr>
<td>Ce/( \text{Ce}^{*} )</td>
<td>1.43</td>
<td>1.63</td>
<td>1.64</td>
</tr>
<tr>
<td>Eu/Eu*</td>
<td>1.59</td>
<td>1.42</td>
<td>0.83</td>
</tr>
<tr>
<td>((\text{La/Lu})_{\text{CN}} )</td>
<td>0.312</td>
<td>0.406</td>
<td>0.076</td>
</tr>
<tr>
<td>((\text{Gd/Yb})_{\text{CN}} )</td>
<td>0.383</td>
<td>0.324</td>
<td>0.237</td>
</tr>
<tr>
<td>Total ppm*</td>
<td>2274</td>
<td>2717</td>
<td>2045</td>
</tr>
</tbody>
</table>

* Based on non-normalized concentrations of La to Lu.

Fig. 9. Anders and Grevesse (1989) chondrite normalized REE plots from zircon spot analyses. Zircons sorted according to their REE pattern and color coded to their respective population based on geochronology in Fig. 8. (a) HREE/LREE enriched zircons separated into zircons with strong Eu anomalies (Red) and zircons with small or absent Eu anomalies (Pink); (b) Zircon cores (darker colors) and rims (lighter colors) with flat HREE/LREE.
the log-linear relationship between temperature and Ti content in zircon and a co-existing Ti phase (usually rutile). The thermometer depends on the activities of both TiO$_2$ and SiO$_2$ and because of this it tends to underestimate the temperature of crystallization by up to 70 °C if the activity $a$TiO$_2$ is <1, and overestimate the temperature by equally as much if the activity $a$SiO$_2$ is <1. However, these effects tend to compensate for one another. Cracks, the grain/epoxy interface, and Ti-bearing micro-inclusions in zircons can elevate measured Ti concentrations and yield spurious temperatures (e.g. Hiess et al., 2008). These factors are mitigated by also measuring other tracers such as Fe in zircon since most contaminants (epoxy, ilmenite/sphene inclusions) are found to also increase Fe abundances (Harrison and Schmitt, 2007; Reid et al., 2011).

4.3.2.1. Thermometry estimates for HREE/LREE enriched zircon cores. This compositional cluster of zircons yielded a relatively narrow range of temperatures, with most Ti$^{46}$ thermometry results falling between 720 and 840 °C. If four anomalous data points are excluded (Supplementary Table T3), an average temperature of 770 °C emerges. While some of the grains gave very high Fe counts, we found no clear relationship between Fe and Ti concentrations in the data. Analysis of the data shows that the zircons with the highest Fe counts did not yield substantially higher temperatures.

4.3.2.2. Thermometry estimates in flat REE zircon cores. Zircons with flat REE patterns in their cores (Section 4.3.1.2) tend to have elevated Ti$^{46}$ thermometry results falling between 720 and 840 °C with an average of about 930 °C (Supplementary Table T3). Although the calculated temperatures based on these analyses are higher when compared to the HREE-enriched zircons (Section 4.3.1.1), it is notable that they are not more enriched in Fe (see above). We interpret this to mean that the elevated temperatures are due to intrinsically higher Ti temperatures for the zircons rather than to contamination. Support for this interpretation comes from a zircon with the lowest Fe concentration in our sample set (lower than most of the HREE enriched zircons), which nevertheless yields a crystallization temperature estimate of 789 °C.

4.3.2.3. Thermometry estimates in flat REE zircon rims. These zircons have Ti$^{46}$ temperatures that have a wide range from 662 to 991 °C with an average of about 820 °C, lower than that of the flat REE zircon cores (Supplementary Table T3). Unlike the previous two populations, however, these zircons show a correlative increase in Ti concentrations with increase in Fe content ($R^2 = 0.87$). Thus, the more elevated apparent Ti temperatures may be due to contamination. The lowest apparent temperature from this data set (~660 °C) provides a minimum temperature of crystallization for these samples.

4.4. Sm–Nd geochronology of AGC components

As shown in Fig. 11 (data in Table 5), the range of Sm/Nd ratios within each lithological component identified in this study and described in Section 4.1.3 is fairly limited, precluding precise age determinations from separate regressions within individual groups. When pooled, however, samples from Lithologies 1 and 2 define a linear array from which only one outlier (sample 6.2) is excluded. When these lithological types are considered together, a scatterchron yields an imprecise age of 3502 ± 320 Ma (MSWD = 29). Removing Lithologies 1 and 2 from consideration reveals a second linear array with shallower slope defined by the four samples from Lithologies 3–4–5. If considered separately, two samples of Lithology 5, and two samples from Lithologies 3–4 define similar ages of 2890 and 2749 Ma, respectively. Considered together, these four analyses define a precise isochron age of 2714 ± 37 Ma (MSWD = 1).

**5. DISCUSSION**

### 5.1. Chronology and petrogenesis of the Acasta Gneiss

Despite evidence that there are much older components within the AGC – such as in the form of some inherited 4200 Ma zircons – no zircons older than about 4020 Ma

---

**Table 5**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Lithology</th>
<th>[Nd] ppm</th>
<th>[Sm] ppm</th>
<th>147Sm/144Nd</th>
<th>+/− (%)</th>
<th>143Nd/144Nd</th>
<th>+/− (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>3-1</td>
<td>1</td>
<td>19.0</td>
<td>2.94</td>
<td>0.0935</td>
<td>0.11</td>
<td>0.510156</td>
<td>12</td>
</tr>
<tr>
<td>5-1</td>
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<td>6.9</td>
<td>1.08</td>
<td>0.0949</td>
<td>0.11</td>
<td>0.510152</td>
<td>24</td>
</tr>
<tr>
<td>7-1</td>
<td>1</td>
<td>8.0</td>
<td>1.16</td>
<td>0.0869</td>
<td>0.12</td>
<td>0.509985</td>
<td>10</td>
</tr>
<tr>
<td>11-1</td>
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<td>8.8</td>
<td>1.20</td>
<td>0.0819</td>
<td>0.12</td>
<td>0.509862</td>
<td>12</td>
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<tr>
<td>4-2</td>
<td>1</td>
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<td>0.0786</td>
<td>0.13</td>
<td>0.509828</td>
<td>12</td>
</tr>
<tr>
<td>8-2</td>
<td>2</td>
<td>5.8</td>
<td>0.84</td>
<td>0.0874</td>
<td>0.11</td>
<td>0.510022</td>
<td>14</td>
</tr>
<tr>
<td>10-2</td>
<td>2</td>
<td>22.4</td>
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<td>0.0652</td>
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<td>0.509475</td>
<td>16</td>
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<tr>
<td>6-2</td>
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<td>1.45</td>
<td>0.0969</td>
<td>0.10</td>
<td>0.510071</td>
<td>10</td>
</tr>
<tr>
<td>2-3</td>
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<td>29.3</td>
<td>5.67</td>
<td>0.1167</td>
<td>0.09</td>
<td>0.510568</td>
<td>12</td>
</tr>
<tr>
<td>2-3'</td>
<td>3</td>
<td>29.2</td>
<td>5.47</td>
<td>0.1143</td>
<td>0.10</td>
<td>0.510354</td>
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<tr>
<td>1-4</td>
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<td>0.0956</td>
<td>0.10</td>
<td>0.510185</td>
<td>29</td>
</tr>
<tr>
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<td>5</td>
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<td>0.0666</td>
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<td>5*</td>
<td>5</td>
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<td>0.0651</td>
<td>0.09</td>
<td>0.509669</td>
<td>12</td>
</tr>
<tr>
<td>12-5</td>
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<td>1.02</td>
<td>0.0771</td>
<td>0.13</td>
<td>0.509863</td>
<td>20</td>
</tr>
</tbody>
</table>

* Samples dissolved in Teflon bombs.
were found in this study. As described in Section 2.3, field relationships indicate that the plagioclase-hornblende schists are perhaps the oldest coherent rocks of the Acasta Gneiss Complex. Zircons extracted from two such units, however, yielded no grains older than 3856 Ma (20% discordant) and no concordant analyses older than 3770 Ma (Supplementary Table T1). Indeed, zircons in the “leucogabbro” samples are very rare, and, where present, are rounded or anhedral with internal structures (either massive or mottled) usually considered inconsistent with igneous growth (e.g. Hoskin and Schaltegger, 2003 and references therein). As we have argued, the “leucogabbro” zircons do not represent the crystallization age of these rocks, but rather reflect metamorphic zircon growth. The absolute age of the “leucogabbro” rock remains unknown.

Based on zircon U–Pb geochronology alone, there appear to be two distinct pre-3900 Ma populations within sample BNB99-151B: The oldest of these is 4004 ± 26 Ma and a younger definable population exists at 3914 ± 20 Ma (Table 3). The oldest population statistically overlaps the reported maximum age(s) of 4030–4050 Ma based on repeat analyses on the oldest zircons discovered from similar samples (Bleeker and Stern, 1997; Stern and Bleeker, 1998; Bowring and Williams, 1999). The next step is to determine which of these two zircon populations represents the actual age of the rock.

5.1.1. The Acasta Gneiss is a migmatite

A striking geochemical feature of the oldest lithologic components in sample BNB99-151B is the strong positive Eu anomaly as shown in Fig. 4a. This distinctive REE pattern is normally found in anorthosites as a result of plagioclase accumulation, but this rock type has thus far not been described from the AGC. Rarer felsic rocks with strong positive Eu anomalies have been found either as minor occurrences in Archean granulite terranes (e.g. Rudnick, 1992) or as leucosomes in migmatitic rocks (e.g. Johannes et al., 2003). Since there is no (preserved) evidence that the AGC was originally part of a granulate terrane, and younger rocks do not share the geochemical signatures found in sample BNB99-151B, we explored the possibility that this and other features of the Acasta Gneiss arose from an early migmatization event.

Leucosomes are the felsic melt portion formed during migmatization. If there is more than ~5% melt fraction, melt migration can occur resulting in a metatexite residual phase largely made of melanosome where a low degree of melt is retained, and a diatexite magma where the melt is concentrated. If the melt migrates away from the metatexite and accumulates, it forms a secondary diatexite. In migmatite systems where the leucosome, melanosome and parent paleosome/mesosome are all present and field relations are interpretable, mass-balance arguments can be made and variations in the geochemical signatures caused by migmatization can be understood. In such systems, leucosomes are overall depleted in REE, but their chondrite normalized patterns have large positive Eu anomalies compared to the parent, irrespective of the parent original composition (see, for example, Milord et al., 2001; Johannes et al., 2003; Bhadra et al., 2006). These characteristics are different from melanosomes, restite-rich mesosomes and larger-scale granitic melts related to migmatites. These systems have REE patterns that closely resemble the parent and either retain the Eu signature or have more negative Eu anomalies than the parent rock generally due to subsequent crystal fractionation processes.

A migmatitic interpretation for the origin of Lithologies 1 and 2 can be supported by the geochemistry, which shares the classic signature of leucosomes formed by either incongruent melting (Jung et al., 1998, 1999), the development of cumulate (Sawyer, 1987), or some combination of the two. All of the leucosomes we sampled shared an overall depletion in REE compared to the parent rock, but have large positive Eu anomalies. The shared geochemical signatures in Lithologies 1 and 2 also mean that Lithology 1 cannot be the paleosome from which Lithology 2 was extracted, or the accompanying melanosome. We consider it more likely that they both originated as a secondary diatexite (melt-dominated) migmatite which either formed as an incongruent continuous melt or as an anatectic melt that underwent crystal fractionation. Part of the extracted melt left a dominantly plagioclase residuum, and the banding now observed is a tectonic fabric that was imparted later.

We note that the positive Eu anomaly found in Lithologies 1 and 2 is absent in the oldest (ca. 4000 Ma) Population 1 zircons. These are comprised almost entirely of grains with negative Eu anomalies and HREE enrichments. The strongest negative anomalies (Eu*/Eu = 0.032–0.437) correspond to zircons with a collective age of 3988 ± 33 Ma (Population 1) vs. another group of 3947 ± 35 Ma zircons with Eu*/Eu = 0.35–1.10 (Fig. 9). All but two of the zircons with HREE enrichments are about 4000 Myr old, whereas those with flatter REE patterns and positive Eu anomalies (both cores and rims) come from Population 2 (ca. 3920 Ma) and Population 3 (ca. 3780 Ma) grains. When only core zircon analyses are considered, the weighted average age of the REE flat and positive Eu anomaly zircons is 3917 ± 57 Ma. This result is indistinguishable from the Population 2 U–Pb zircon age (3914 ± 20) reported in Table 5. Thus, it is likely that the Population 2 zircons formed during the migmatization process and Population 1 zircons were entrained into the melt from the original parent rock. This interpretation is further supported by the output from lattice strain modeling described in the next section.

5.1.2. Lattice strain models of zircon REE distributions discriminate between indigenous vs. inherited zircon

It has long been known that the crystal-melt partition coefficients for a series of isovalent cations such as the REEs show a near-parabolic dependence on cation radius (Onuma et al., 1968; Brice, 1975). Lattice strain models are based on the fact that crystal composition is a major factor controlling the partitioning of ions between crystal and melt (Shaw, 2006 and references therein). Owing to the fact that REEs in zircon have extremely low diffusion rates and that rates of diffusion are particularly low for the LREEs (Cherniak et al., 1997), contents in the Acasta zircons are likely robust and representative of values from which the zircon formed. The Population 1 and 2 zircons
have such different REE patterns, so we should be able to determine with lattice strain theory which of these is most likely to have crystallized in a melt with the composition of the host lithology (e.g. Trail et al., 2007b; Cates and Mojsis, 2009). With this task in mind, we undertook a comprehensive comparison using lattice strain models of each zircon REE pattern (excluding Ce and Eu due to their variable +3/+4 and +3/+2 valence states, respectively) against the REE patterns of the different slab components described in Section 4.1.3.

Partition coefficients were determined for each zircon in our database (REE$_{\text{ain}}$/REE$_{\text{WR}}$) and fitted using the Blundy and Wood (1994) lattice strain model. The results from Ti$_{\text{melt}}$ described in Section 4.3.2 were used as input parameters to the temperature term in the lattice strain calculations. The best fit parabola (described by its $R^2$ value) for the exponential function was then calculated iteratively over 1000 model runs. Examples of the results are presented graphically in Fig. 10 for all zircon populations discussed herein, and a summary of the output $R^2$ values for each fit function calculation are shown in Table 6 and the Supplementary Table T4.

A great majority of analyzed zircons in this study have REE patterns that generally conform to the expected zircon compositions for Lithologies 1 and 2, and fit the least well with Lithology 5. This is the expected result from the lithologies described in Fig. 2. At closer inspection, zircons with flat REE patterns (both cores and rims) have the best fit overall ($R^2 \approx 0.92$) with a slightly better fit for the more felsic Lithology 2. Zircons with enriched HREE patterns have the poorest fit functions ($R^2 \approx 0.86$) with Lithologies 1 and 2, and zircons that are both enriched in HREE and have large negative Eu anomalies fit best with Lithology 3, but this is still a relatively poor relation ($R^2 = 0.87$). Not surprisingly, discordant grains have a wide range of fits (Supplementary Table T4), but as expected the most discordant grains have the poorest apparent $P_{\text{WR}} \text{[REE]}$ correspondence.

Thus, Population 1 zircons with ages of about 4000 Ma have the poorest correspondence with the host-rock composition for Lithologies 1 and 2 ($R^2 = 0.86$), while the Population 2 zircons show a better fit ($R^2 = 0.91$; Table 6). Consequently, both differences in the general zircon REE chemistry and lattice strain models used to reconcile zircon composition with whole rock (host) composition support the interpretation that only the zircons from Population 2 (ca. 3920 Ma) crystallized from a melt of the composition of Lithology 1 (and lithology 2).

This analysis strongly suggests that the pre-4000 Ma zircons in the Acasta gneisses are inherited from a pre-existing rock that was destroyed during the formation of the oldest coherent component of the Acasta gneisses in a migmatization event (Section 5.1.1). Our interpretation is given further credence if one examines the whole rock “zircon saturation temperatures” (Watson and Harrison, 1983) calculated for Lithologies 1 and 2. Output yields average temperatures around 840 °C (labeled as “$M$ values” in Table 2), which is some 80 °C higher than that for average early Archean TTGs as derived from the compilation in Rudnick and Fountain (1995). When considered in context with the Ti$^{57}$ thermometry and elevated Zr contents for Lithologies 1 and 2 cited above, the argument for zircon inheritance in the oldest (pre-4000 Ma) AGC population has merit: The Ti$^{57}$ thermometry for Population 1 zircons yields average temperatures (803 ± 3°C) which are slightly cooler than the rock’s calculated zircon saturation temperature. If two suspect analyses (>1000 °C, that also have elevated Fe) from Population 2 zircons are excluded from this analysis, then the Ti$^{57}$ thermometry for Population 2 zircons conveniently matches the whole rock’s zircon saturation temperature at 844 °C. Taken together, the ca. 3920 Ma zircon age population best fits all the geochemical constraints and we deem that this represents the migmatization age of the Acasta gneisses.

It is unlikely that Lithologies 1 and 2 faithfully represent the magma from which the Population 2 zircons crystallized, because the entrainment of inherited zircon clearly shows that the melt was contaminated with some older rock. If the possible Eu anomaly observed in the lithologies is partly due to crystal fractionation, plagioclase accumulation and the removal of the fractionated portion of the melt during solidification, it is possible that the melt composition was quite different from the rock that now remains (see for example, Johannes et al., 2003). We can determine, however, the possible influence of plagioclase accumulation and restitic contamination on the composition of the mimarite if we can infer the composition of the original rock and know the composition of the cumulate plagioclase.

### 5.1.3. Compositional constraints on the parent rock to the oldest AGC zircons

Since we have not yet definitively sampled or observed a likely candidate for a restite or parent rock for Lithologies 1 and 2, our best opportunity to constrain the composition of the parent rock comes from the geochemistry of Population 1 zircons. The ca. 4000 Ma zircons have REE patterns distinct from all other zircons in the various Acasta Gneiss components, and it may be possible to broadly limit the possible magma compositions that gave rise to them (e.g. Belousova et al., 2002; Grimes et al., 2007). To explore whether trace element compositions for the old HREE/LREE-enriched zircon population (>4000 Ma) can be reconciled with any Acasta lithotype for which data are available, we performed element partition modeling of the oldest population’s [REE$_{\text{ain}}$ against [REE]$_{\text{WR}}$ of all available AGC whole rock data.

Ironically, zircons from the Population 1 group (ca. 4000 Ma) that possess the most pronounced negative Eu anomalies best match (albeit weakly; average $R^2 = 0.87$) mafic granitoid gneiss AG09008 for which no zircons of this age population were found. Another feature of sample AG09008 is that it is poorer in silica (57 wt.% SiO$_2$) than any of the current host rocks to the ca. 4020 Ma zircons (Lithologies 1 and 2 ~70 wt.% SiO$_2$), even while it is also a member of a younger generation of intrusions. This correspondence is hardly any better than the average fit ($R^2 = 0.86$) that the ca. 4000 Ma zircons define with their present host rock. When zircons with negligible or no Eu anomalies are compared to any other AGC lithologies in
Fig. 10. Onuma diagrams coupled with lattice strain partition modeling results of selected zircons, using Zircon REE/Average Lithology as the partitioning parameter. (a) Population 1 zircons; (b) Population 2 zircon cores; (c) Population 2 zircon rims.
While the compositional match determined by $D_{\text{zircon}}^{\text{REE}}$ modeling between much younger and more mafic tonalitic gneiss and the oldest zircons is reasonably good, this is an unsatisfactory result given the geochronology results presented in Section 4.2. We therefore pursued tests of the same zircon compositions against a series of model melt compositions derived from literature compilations following Trail et al. (2007a). The average compositions chosen for this comparison were N-MORB (Bevins et al., 1984); Archean granite and TTG (Rudnick and Fountain, 1995); anorthosite (Ashwal and Seifert, 1980); adakite, continental arc, Andean arc, island arc, low-Al and high-Al TTGs, and plagiogranite (Drummond et al., 1996). Results (Fig. 12, Supplementary Table T6) show that the more mafic compositions provided the best fit to our Population 1 zircons; average N-MORB, adakite and anorthosite all provided $R^2$ values over 0.86, and over 0.89 in the case of zircon 1_1 from mount 2.

To complement the lattice strain approach, we employed zircon discrimination diagrams derived by Grimes et al. (2007) to compare Yb to other trace cations as a constraint on likely source melt compositions. This is possible since U and Yb have different compatibilities in magmatic systems as more evolved magmas tend to be enriched in U but depleted in Yb. It also happens that U and Yb have nearly identical mineral/melt partition coefficients in zircon (Bea et al., 1994) and therefore preserve a reliable record of the U/Yb of the magma from which they crystallized. In this way it is possible to distinguish ocean crust zircons from continental zircon and, to a lesser extent, from island arc magmas. Grimes et al. (2007) assembled a large compositional database including zircons from mafic oceanic rocks, oceanic plagiogranites, and continental arcs. To this original dataset we added zircon analyses from other lithologies including diorites and aplite dikes (Belousova et al., 2006; Schulz et al., 2006; Kaczmarek et al., 2008). Acasta zircons that carry HREE/LREE enrichments and negative Eu anomalies plot in the same field as Ti-poor gabbros and diorite in this diagram (Fig. 13; Ballard et al., 2002). The plot shows that magma sources for the Population 1 zircons are distinct from plagiogranites and more felsic rocks: the former have Yb and U contents indistinguishable from more mafic oceanic zircons and the latter display lower Yb concentrations at comparable U concentration (Ballard et al., 2002; Belousova et al., 2006; Grimes et al., 2009).

As an internal check, we used the techniques described in Section 5.1.2 against the discrimination diagram in

---

Table 6
Summary of average $R^2$ values from lattice strain partition modeling.

<table>
<thead>
<tr>
<th>Averages</th>
<th>Average Lith 1</th>
<th>Average Lith 2</th>
<th>Average Lith 3</th>
<th>Lith 5 (Acasta E)</th>
<th>Ti-in-zirc (T)</th>
<th>$^{207}$Pb/$^{206}$Pb age (Ma)</th>
<th>$%^{206}$Pb*</th>
<th>Th/U</th>
<th>% Concord.</th>
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<tbody>
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<td>HREE zircons*</td>
<td>0.86</td>
<td>0.86</td>
<td>0.87</td>
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<td>799</td>
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<tr>
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<td>0.87</td>
<td>0.81</td>
<td>0.81</td>
<td>768</td>
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<td>99</td>
<td>0.7551</td>
<td>98</td>
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<tr>
<td>Flat core zircons</td>
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<td>0.88</td>
<td>0.89</td>
<td>908</td>
<td>3952</td>
<td>100</td>
<td>0.6998</td>
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</tr>
<tr>
<td>Flat rim zircons</td>
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<td>0.92</td>
<td>0.84</td>
<td>0.89</td>
<td>819</td>
<td>3743</td>
<td>99</td>
<td>0.2380</td>
<td>97</td>
</tr>
<tr>
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<td>0.88</td>
<td>984</td>
<td>3691</td>
<td>97</td>
<td>0.4843</td>
<td>64</td>
</tr>
</tbody>
</table>

* Zircons with strong Eu anomalies.
† Zircons without strong Eu anomalies.
the $R^2$ outputs of these analyses are presented in Fig. 14 and Supplementary Table T7. As expected from the discrimination diagram results, granitoid arc lithologies had poor fits to the lattice strain model with average $R^2$ ranging from 0.80–0.83. The MORB basalts and gabbros also had poor fits, with $R^2$ averages spanning 0.76–0.82. Volcanic arc and within-plate basalts fared somewhat better with average $R^2$ ~ 0.84. Of all the magma compositions we tested from these various databases, the best fit to the data comes from a Mg-rich andesite from the southwestern part of the Japan Arc (Tatsumi et al., 2006). Average $R^2$ values were 0.88 when an outlier with the least HREE/LREE enrichment and smallest Eu anomaly is omitted.

Compositions for the high-magnesium andesite Oto-Zan lava flows compared well to theoretical melt compositions for the Acasta HREE/LREE-enriched zircons. Of these, a sample from the lowest part of the Oto-Zan flow, considered to be the best representative of the original melt composition (OTO-1 of Tatsumi et al. (2006), serves as the best match to our data for the oldest AGC zircons. The Oto-Zan andesites are moderately LREE enriched (WR/CI ~ 60), with no Ce or Eu anomalies and are similar to the chondrite-normalized patterns for within-plate and volcanic arc basalts. These melts are likely sourced from direct melting of mantle peridotite caused either by the injection of fluids from the underlying slab or infusion and reaction with slab melts (Kelemen et al., 1995; Shimoda et al., 1998; Hanyu et al., 2002). We therefore propose a candidate protolith for the ca. 4050–4000 Ma grains of the Acasta Gneiss, and the parent rock for the diatectic migmatite preserved in Lithologies 1 and 2, to have been a mafic to intermediate rock with moderate LREE enrichments akin to those that form in plate boundary settings.

5.1.4. Migmatization of the Acasta protolith

Lithologies 1 and 2 do not plot near the cotectic for a haplogranitic melt at any pressure, which would be predicted for a small-volume initial equilibrium melt (Fig. 15;
They do fall, however, within the field experimental dehydration melts of mafic protoliths at high temperatures (>925 °C) and pressures (4–16 kbar; Fig. 15; Rushmer, 1991; Wolf and Wyllie, 1991; Springer and Seck, 1997). By measuring the elemental compositions of the major mineral components (Supplementary Table T8) and assuming that the Oto-Zan andesites are comparable to the parent rock, the influence of (i) melt contamination from the parent rock, (ii) plagioclase accumulation and (iii) fractional crystallization can be evaluated (see Sawyer, 2008 and references therein).

Cation vs. SiO₂ plots show that the various trends of accumulation or contamination occur as strong linear variations within Lithologies 1 and 2. Both felsic (CaO + Na₂O) and mafic (FeO + MgO) cations show that the trend within Lithologies 1 and 2 are towards contamination consistent with (but not proof of) a Mg-andesite (Oto-Zan-like) protolith with little involvement of plagioclase accumulation and or of biotite (Fig. 16). The potassium composition within Lithologies 1 and 2 is dependent on the amount of biotite and shows no significant influence from fractional crystallization. Based on major cation trends, it appears that Lithologies 1 and 2 represent anatectic melts from an intermediate source that did not undergo fractional crystallization, and plagioclase accumulation was not the driving source for the positive Eu anomalies. They could have formed under disequilibrium conditions as an incongruent continuous melt with variable contamination from the parent lithology.

5.1.5. Comparison to other ancient gneiss samples in the AGC

Bowring and Williams (1999) reported SHRIMP data for three AGC samples (SAB91-37, SAB91-63, and SAB94-134) and the oldest zircons they found were older than about 4000 Ma. No whole-rock geochemical data were provided in that study, nor were trace element data forthcoming for the zircons that could aid in evaluating inheritance. Retrospective examination of the zircon CL images accompanying those data shows that most of the oldest ages were for structureless or weakly zoned zircon cores. Conversely, the oldest strongly oscillatory zoned cores...
tended to be part of the ca. 3920 Ma age population. The (weak) internal zones mentioned in the Bowring and Williams (1999) report resemble what have been termed “ghost textures” by Hoskin and Schaltegger (2003), and these structures can be consistent with inherited grains. Similarly, Iizuka et al. (2006, 2007) analyzed over 100 zircons from a variety of Acasta samples collected over a large area and some of these also had ages greater than 3920 Ma. Whole rock geochemistry reported for sample BGXM of Bowring et al. (1990) shows a positive Eu anomaly, just like the oldest (ca. 3920 Ma) rocks in our collection. As described above, a positive Eu anomaly is a fingerprint of the oldest migmatized granitoid gneissic component of the AGC.

On the other hand, Lithology 3 from sample BNB99-151B does not show nearly as strongly a Eu anomaly as Lithologies 1 and 2, nor does it contain any zircons older than about 3922 Ma; the oldest age in Lithology 3 comes from one out of three spots collected from a CL-dark zircon population for zircon growth may have been unavailable. Alternatively, the whole-rock Sm–Nd ages for the Acasta Gneiss:

5.1.6. Whole rock Sm–Nd ages for the Acasta Gneiss:

Overall, our results are consistent with previous Sm–Nd studies of the AGC in that they confirm the lack of a preserved 4000 Myr-old $^{143}$Nd$^{144}$Nd chronological record for this system. While the linearity of the Sm–Nd data presented by Moorbath et al. (1997) was interpreted by them as the age of final closure of the Sm–Nd system in the Acasta gneisses, it is an age that is actually rarely recorded in the Acasta zircons. Bleeker and Stern (1997) reported a granitic gneiss of this age, Bowring and Housh (1995) reported some zircon overgrowths at about 3400 Ma. Stern and Bleeker (1998) noted a 3356 Ma age on ca. 4000 Ma grains, and we found a single zircon (3 analyses) with an age of 3355 ± 49 Ma. Zircon growth requires that the necessary elements for growth be released at the time of metamorphism; as these rocks had previously undergone two amphibolite grade phases of metamorphism (ca. 3700 Ma and ca. 3600 Ma), which included partial anatexis (Bowring and Housh, 1995; Iizuka et al., 2007), the constituent elements for zircon growth may have been unavailable.

Alternatively, the whole-rock Sm–Nd isochron age of 3371 Ma is a spurious estimate resulting from pooling lithologies with different reset ages. Our analysis found that the age of 3502 ± 320 Ma for Lithologies 1 and 2 is consistent with global resetting of the Sm–Nd system during anatexis at 3660–3590 Ma (Bowring and Housh, 1995; Iizuka et al., 2006, 2007). Component analysis of sample BNB99-151B also reveals domains characterized by a late Archean metamorphic age (ca. 2700 Ma) in components 3 and 5. We speculate that since Lithologies 3 through 5 are derived from intrusive rocks they may have represented preferential pathways for metamorphic fluids, thereby favoring isotopic re-equilibration within these layers as the ages correspond to the time of known plutonism in the area.

 Ages obtained using whole rock Sm–Nd isochrons on complicated polymetamorphic rocks should not be interpreted simply at face value. If applied to samples with
characteristics similar to BNB99-151B, the whole rock method merely randomly samples domains with different equilibration ages to produce "intermediate" dates with no geological significance. Such bias would be difficult to identify on the basis of whole-rock analyses but is readily identifiable using the component analysis presented here. While the existence of a 3370 Ma event completely resetting the Sm–Nd system on the whole-rock scale cannot be ruled out, it is also plausible that this age is an artifact produced by including lithologies with different Sm–Nd closure ages. For this reason, it is more likely that accurate information on mantle heterogeneity and/or early crustal growth will be achieved by analyzing those rocks that preserve the oldest igneous abundant concordant zircons for their Lu–Hf isotope compositions (Guitreau et al., in revisions).

5.1.7. Heterogeneous redox of the Hadean crust from $f_{O_2}$ estimates of Acasta magmas

Magma oxygen fugacity ($f_{O_2}$) influences mineral saturation, melt polymerization, and the speciation of volatile gases that exsolve during a volcanic eruption; these processes can be recorded in zircon (Ballard et al., 2002; Burnham and Berry, 2012). To explore these processes in more detail and how they may be recorded in ancient zircon crystal chemistry, Trail et al. (2011, 2012) performed laboratory experiments in which zircons were synthesized in Ce- and Eu-doped granitic melts; the oxygen fugacity was controlled (i.e. buffered) from about iron–wüstit (IW) to magnetite–hematite (MH). Analysis revealed that zircon Ce anomalies were sensitive to both $f_{O_2}$ and temperature, while Eu anomalies were sensitive only to the $f_{O_2}$ of the magma. In natural systems, however, zircon Eu anomalies are influenced by other factors such as feldspar depletion of Eu relative to neighboring REEs Sm and Gd in the competing mineral effect. Cerium does not suffer from this complication because no major rock-forming phase preferentially sequesters La and Pr relative to Ce. In this way Ce anomalies form the basis of an $f_{O_2}$ sensor for zircon-bearing rocks, as long as the crystallization temperatures are known. The data presented here permit such a calculation to be made. Zircons plotted in Fig. 9a (Population 1) are considered in this model because these patterns generally

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Fig. 16. Compatible vs. incompatible major element plots showing major trends in Lithologies 1 and 2 towards a parental rock of composition similar to Oto-1 (Tatsumi et al., 2006). Biotite and plagioclase compositions taken from electron microprobe analyses from Lithologies 1 and 2. Trends of possible contamination and fractional crystallization indicated by arrows. (a) SiO$_2$ vs. CaO + Na$_2$O; (b) SiO$_2$ vs. FeO$_t$ + MgO; (c) K$_2$O vs. CaO + Na$_2$O; (d) K$_2$O vs. FeO$_t$ + MgO.
display typical monotonic increases for trivalent REE-normalized patterns from La to Lu (Trail et al., 2012). If the two types of REE patterns are considered separately (Table 2), then magma oxygen fugacities of ~IW (“no Eu anomaly”) and ~IW to <FMQ (“Eu anomaly”) for Population 1 zircons are predicted.

The simplest interpretation is that these two zircon types crystallized from a reduced and a more oxidized suite of magmas, respectively. That said, diminished Ce anomaly from small inclusions or subsequent metamorphic episodes – or other effects related to melt composition – cannot be excluded (Trail et al., 2012). Nevertheless, the magmas considered to have given rise to the Acasta gneisses are, on average, more reduced than those of the source melts of detrital Hadean (Jack Hills, Western Australia) zircons documented thus far. This would seem to indicate that heterogeneous magma redox states were present in the Hadean lithosphere, which is not surprising. The fact that negative Eu anomalies and positive Ce anomalies correlate (using either CN or WR normalization) is an apparent contradiction which seems to indicate reducing and oxidizing conditions simultaneously. When crystallization temperature is considered from Ti\(_{\text{Sm}}\) (Section 4.3.2), however, this apparent contradiction vanishes. Cerium anomalies correlate with 1/T, whereas in a simple zircon-melt system, Eu anomalies are not predicted to correlate with temperature in agreement with Trail et al. (2011, 2012). In natural systems the modal abundance of feldspar increases along the liquid line of descent so that progressively greater negative melt anomalies will prevail. This explains the larger negative Eu anomalies in Acasta zircon, which correlate with lower Ti-in-zircon crystallization temperature.

### 5.2. Implications for the Hadean surface state: Late heavy bombardment and its plausible effects on the Acasta crust

Although its age overlaps with estimates for the time of the LHB from both dynamical arguments (Morbidelli et al., 2012) and detailed analysis of lunar rocks (Norman and Nemchin, 2014 and references therein), we deem it unlikely that the ca. 3920 Ma Population 2 zircons represent a record of impacts to the Hadean crust. This is mainly due to the preservation of the inherited and concordant ca. 4020 Ma Population 1 zircons and the geochemical compositions of the host rocks, despite the thermal effects of a global LHB modeled for that time (Abramov and Mojzsis, 2009; Abramov et al., 2013). As shown by Wielicki et al. (2012), zircons collected from central melt sheets of impact craters generally show a uniform age distribution and the absence of inherited grains due to high initial temperatures and the resulting resorption of all relict grains. Central uplifts, which underlie melt sheets of large craters, can reach temperatures of ~1000 °C and take over 10^6 years to cool (Ivanov, 2004; Abramov and Kring, 2007). The expectations are that these conditions ought to lead to effective age-resetting in zircons.

As reviewed in Abramov et al. (2013), the distribution of temperatures associated with impacts – with very hot central regions and temperatures decreasing exponentially from the center of an impact – leads to an “all-or-none” condition for age-resetting in zircons (e.g. Moser et al., 2011). Exceptions to this condition exist in only relatively narrow zones within an impact structure large enough to generate melt (Fig. 17). Population 1 zircons display a high degree of concordance, which suggests that their Pb isotope systematics were not disrupted by a highly energetic later event such as by impact. Zircon saturation temperatures typically underestimate crystallization temperature for differentiated impact melts by ~50–100 °C (Wielicki et al., 2012), but for Population 2 zircons we find that the two are equivalent (~840 °C). In general, although the ca. 3920 Ma age assigned to the Acasta Gneiss in this work is unlikely to be representative of an impact-generated melt, an increased bombardment flux at ca. 3900 Ma (e.g. Tera et al., 1974; Ryder, 1990; Kring and Cohen, 2002) may have caused significant fracturing of the crust, enabling or facilitating magmatic incursions at that time.

An impact origin for the oldest, ca. 4000 Ma Population 1 zircons found in this study cannot be ruled out. The REE abundances for most Population 1 zircons show positive Ce and negative Eu anomalies, and are actually a good match for REE distributions found in zircons derived from impact melt sheets (Wielicki et al., 2012). This distribution is not diagnostic as it is also commonly found in igneous zircons of various ages (Peck et al., 2001; Hoskin and Schaltegger, 2003; Grimes et al., 2007). Moreover, the average zircon crystallization temperature of 803 °C for Population 1 zircons is close to the 773 ± 87 °C average reported for zircons produced in impact melt sheets as reported by Wielicki et al. (2012). Population 1 zircons further display tight
age clustering (Fig. 8; Table 5) that suggests (but does not prove) that they were produced in a single, relatively short-lived event. The inferred mafic-to-intermediate composition of the Population 1 zircon protolith also happens to be consistent with the expectation for crustal melts generated by large impact events: lower (mafic granulite) crust and even mantle material may be included in impact melt sheets, and all other things being equal, the larger the impact, the deeper the extent of the melt zone and the more mafic the composition of the melt sheet. For example, the ~180-km Chicxulub crater (Yucatan, Mexico) melt zone extended down to ~28 km beneath the original pre-impact surface and likely reached the lower crust (Kring, 2005). Impact-generated uplift is another plausible mechanism for delivery of mafic material. The amount of uplift scales as 0.086D^{1.03}, where D is the final rim-to-rim crater diameter in kilometers, based on empirical data from a total of 24 terrestrial impact structures (Grieve and Pilkington, 1996). This implies ~20 km of structural uplift for a ~180-km crater, which can potentially deliver mafic rocks from the lower crust to the near-surface.

The latest dynamical and observational estimates for the start of the LHB on the Moon put it at approximately 4100–4200 Ma (Bottke et al., 2012; Morbidelli et al., 2012; Norman and Nemchin, 2014), adding further plausibility to the impact origin of Population 1 zircons and to the oldest ca. 4200 Ma inherited zircons discovered by Iizuka et al. (2006). Regardless, the likelihood that a particular parcel of the crust was age-reset during the LHB is only ~10% (Abramov and Mojzsis, 2009) at the most (Abramov et al., 2013), thus rendering this scenario less than likely from a probabilistic point of view. Following the results of Abbott et al. (2012), and Bell and Harrison (2013), it would be worth undertaking a thorough analysis of the oldest zircons of the Acasta Gneiss Complex in the search for direct evidence of the LHB on Earth. More data from the few, rare oldest (ca. 4200 Ma) zircons thus far discovered in the AGC by Iizuka et al. (2006) need to be interrogated to constrain plausible formation scenarios and to determine whether they, too, may be of impact origin or from normal crustal processes.

6. CONCLUSIONS

Although the oldest U–Pb assigned ages for the Acasta Gneiss Complex go back as far as about 4050 Ma and yet older ca. 4200 Ma zircons have also been found, debate over the actual emplacement age(s) for different generations of Acasta rocks has long clouded the potential for these rocks to reveal more about the Hadean crust. We find that the trace-element chemistry of all AGC zircons older than about 4000 Ma is incompatible with the rock that hosts them and conclude that they do not record the emplacement of these rocks but instead represent inheritance. There is no doubt that the AGC preserves a legacy of Hadean crust, and that this crust was not destroyed in the bombardment epoch. We do find, however, that this rock was migmatized before ca. 3920 Ma in a process which involved incomplete assimilation of older material. What was the nature of that pre-existing crust? Analysis of U, Yb and Y zircon trace element discrimination diagrams and partition modeling of the oldest AGC zircons with natural rock samples shows that a good candidate for the parent of the migmatized rocks, and host crust zircons older that about 4000 Ma, is a mafic to intermediate rock with only moderate LREE/HREE enrichment and no positive Eu anomaly. A dominant population of ca. 3920 Ma concordant zircons (what we have termed Population 2) has the highest degree of compositional correspondence via zircon/rock partitioning, Ti0.66 zircon saturation temperatures and other trace element criteria is with the host rock of these zircons. We interpret this 3920 Ma age as the age of the AGC. Melts which formed the oldest indigenous lithological components of the AGC have distinctive trace element compositions that include strong positive Eu anomalies as was also noted in the only other published whole rock trace element geochemical analysis from the area (Bowring et al., 1989). Based on field relationships, a group of plagioclase-hornblende rocks loosely termed “leucogabbros” are the oldest coherent rocks of the complex, but as yet their absolute age remains unknown.

New 143Nd–144Nd geochronology data for a subdivided ~60 cm2 slab of Acasta banded gneiss comprising five separate lithologic components were compared to other variably deformed granitoid- and plagioclase-hornblende AGC gneisses. Despite the fact that different gneissic components carry distinct [Th/U]arc vs. Ti0.66 and [REE]arc signatures correlatively with different zircon age populations and WR compositions, they do not agree with 143Nd–144Nd isotope systematics.

Partition modeling shows that the magmas that responsible for the inherited ca. 4020 Ma AGC zircon population formed at ~IW (iron–wüstite) to <FMQ (fayalite–magnetite–quartz) oxygen fugacities. These values are lower than those invoked to explain the Hadean detrital zircon record from Western Australia (Narryer Gneiss Complex) and demonstrate that the Hadean crust was heterogeneous with respect to oxidation state.

The events that emplaced the protolith of the migmatized Acasta Gneiss Complex at ca. 4020 Ma overlap in time with the proposed LHB of the inner solar system. Owing to this coincidence, it makes sense to extend observations in the AGC to markers of catastrophic impacts. We cannot rule out the possibility that the oldest AGC zircons, including the rare ca. 4200 Ma population, could have formed in impact melt sheets. Nor can we rule out the role of impact hearing in the migmatization of the Acasta protolith. Further work is needed to explore these possibilities.

Better documentation of the state of preservation of the oldest rocks of the AGC, their age spectra and details of compositional variations, paves the way for new searches for remnants of Hadean crust. The Acasta rocks are dominantly felsic TTG-gneisses with characteristics that can be understood in the same way later Archean-Proterozoic rocks have been interpreted. The availability samples of such antiquity means that they are well poised for new studies that seek to document the decay products of short-lived isotopes such as 143Nd, 185W, and 53Cr, coupled to complementary long-lived radiogenic isotope systems.
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APPENDIX A. SUPPLEMENTARY DATA

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.gca.2014.02.019.

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