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1 **Geochemical and textural investigations of the Eoarchean Ukaliq**
2 **Supracrustals, Northern Québec (Canada)**

3
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17

1 Structural, geochronological and geochemical studies of pre-3.75 Ga rocks of volcano-sedimentary
2 protoliths in the Inukjuak domain in the Superior Province in Québec have been mostly focused on the
3 Nuvvuagittuq Supracrustal Belt (NSB). The Porpoise Cove outcrops – at the southwestern limit of the
4 NSB – are the *de-facto* “type locality” for the supracrustals of the Inukjuak Complex. Yet, it remains
5 unclear whether the NSB rocks are geochemically distinct from, or are in fact common to, a host of other
6 supracrustal enclaves locked in the dominantly gneissic Inukjuak domain. Here, we report detailed
7 textural and geochemical studies for a suite of rocks from the Ukaliq Supracrustal Belt (USB), located
8 approximately 3 km northeast of the NSB. We find that the USB and NSB have a similar protracted
9 metamorphic history; both experienced amphibolite grade metamorphism and contain granitoid gneiss
10 sheets, quartz-magnetite rocks (banded iron-formation *s.l.*) and quartz-biotite schists within
11 amphibolitized rocks of basaltic affinity with local retrogressions. If the various Inukjuak supracrustal
12 belts were once a part of a larger coherent (now dismembered) terrane, they should show similar
13 emplacement ages and source chemistry. New zircon U-Pb geochronology from five gneissic units and
14 two quartz-biotite (metasedimentary) schists reveal the oldest emplacement ages across all units of each
15 individual rock type to be 3.68 ± 0.07 Ga (granitoid gneisses) and 3.65 ± 0.06 Ga (quartz-biotite). These
16 new ages are similar to those documented as likely minimum emplacement ages of the NSB determined
17 by U-Pb geochronology. Zircons from the quartz-biotite schist were also analyzed by ion microprobe U-
18 Pb geochronology and were found to yield statistically identical, albeit more precise, ages than those
19 obtained by LA-ICP-MS. Possible detrital zircons from the USB quartz-biotite schists were analyzed by
20 ion microprobe for their coupled $\delta^{30}\text{Si}_{\text{NBS28}}$ and $\delta^{18}\text{O}_{\text{VSMOW}}$ values with respective values between -0.75
21 and -0.07 ‰ ($\delta^{30}\text{Si}_{\text{NBS28}}$), and +5.61 and +6.59 ‰ ($\delta^{18}\text{O}_{\text{VSMOW}}$). The $\delta^{18}\text{O}_{\text{VSMOW}}$ values, which are on
22 average above mantle-derived zircon, indicate contribution of altered, non-mantle derived material into
23 the source of the rocks that weathered to form the quartz-biotite schists. Zircon mineral inclusions (quartz,
24 feldspar, apatite, biotite, muscovite and other unrecognized Fe/Al/Si rich phases), along with the major-
25 and trace element contents for the rocks were analyzed to substantiate this interpretation. Together with
26 $\delta^{30}\text{Si}_{\text{NBS28}}$, $\delta^{18}\text{O}_{\text{VSMOW}}$, our results suggest that lithologies like authigenic silica and serpentinized rocks re-
27 melted to form the parent melts that gave rise to zircons found in the USB quartz-biotite schists.
28 Additional Lu-Hf studies reported here on the same zircons also show similarities with NSB zircons. The
29 ϵ_{Hf} values showed a positive correlation with the measured U-Pb age from -22.7 ± 0.8 to $+1.9 \pm 1.1$. The
30 Lu-Hf system also reveals that the USB, extracted at ca. 3.8 Ga, carries isotopic signatures of an older
31 Hadean reservoir, having been formed from an Eoarchean mafic melt that incorporated them. Taken
32 together, this supports a co-genetic origin for the NSB and the USB.

1 **1. Introduction**

2 Although consensus is difficult to come by in any discussion of the Hadean (pre-3.85 Ga) Earth,
3 it is generally agreed that our planet underwent a major silicate differentiation event early in its history
4 (4.4 – 4.5 Ga) (e.g. Boyet et al., 2003, 2005; Caro et al, 2003; Harrison et al., 2008) subsequent to the
5 postulated Moon-forming “Giant Impact” (e.g. Tonks and Melosh, 1993; Nakajima and Stevenson, 2015)
6 and likely after the hypothesized “Late Veneer” at ca. 4.48 Ga (e.g. Mojzsis et al., 2019). It has also been
7 proposed that the magma ocean which cooled from the Giant Impact event created (ultra-)mafic Hadean
8 crust via fractional crystallization (e.g. Caro et al., 2005; Bourdon and Caro, 2007; Rizo et al., 2011).
9 Perhaps at odds with this interpretation, the relatively rapid crustal evolution towards granitoid crust is
10 revealed through study of pre-4.0 Ga (Hadean) zircons (e.g. Maas et al., 1992; Mojzsis et al., 2002; 2014;
11 Harrison et al., 2017; Trail et al., 2018, Turner et al., 2020). A sparse rock and mineral record from the
12 first few hundred million years on Earth, however, means that a better understanding of the nature of
13 early crustal recycling remains elusive. For instance, there is long-standing debate over the operative
14 mechanisms in primordial crustal recycling (Korenaga, 2018); many have advocated what is termed
15 “vertical tectonics” (e.g. Robin and Bailey, 2009) wherein the style of subduction is broadly similar to
16 reverse diapirism. It is noteworthy that this is a mechanism much slower compared to modern subduction
17 zones (Korenaga, 2006; Foley et al., 2014). While a number of workers propose some form of mobile-lid
18 processes in the Hadean (Polat et al., 2002; Furnes et al., 2007; Sleep et al., 2014; Foley and Rizo, 2017),
19 there also exists the view that, the Hadean Earth was under the so-called ‘stagnant lid’ plate regime
20 (O’Neill and Debaille, 2014; Griffin et al., 2014). The stagnant lid idea shares some dynamical
21 similarities to contemporary Venus wherein lateral lithospheric creep and mantle upwelling contribute to
22 a relatively serene rock cycle punctuated by catastrophic mantle/crust overturn events (e.g. Basilevsky
23 and Head, 2002 and references therein; cf. Rolf et al., 2018). The stagnant lid hypothesis may (or may
24 not) point towards crust foundering as the dominant recycling process on terrestrial-type planets with
25 irregular density contrasts (e.g. Jagoutz and Behn, 2013; Zandt et al., 2004; van Thienen et al., 2004).
26 Argument also prevails over whether plate tectonics as we know it could have initiated in sometime later
27 in the Archean (Shirey and Richardson, 2011; Bédard et al., 2003).

28 Nearly all that we directly know of the first few hundred million years of Earth history comes from the
29 study of Hadean zircons mostly from the northwestern Narryer Gneiss Complex in Western Australia
30 (Froude et al., 1983; Compston and Pidgeon, 1986). Although these are the oldest earthly materials for
31 which absolute ages are available, it is important to emphasize that the grains are orphaned from their host
32 rocks which either have long since been destroyed, or simply remain undiscovered. Extrapolations made
33 from the mineral geochemistry and inclusion mineralogy of these remarkable zircons has yielded useful
34 insight on the Hadean surface environment (Cavosie et al., 2005; Trail et al., 2016; Bell et al., 2017; Trail

35 et al., 2017; Harrison et al., 2017; Weiss et al., 2018). Considering this sampling handicap, however, any
36 new lading of Hadean/Eoarchean crust has the potential to greatly enhance what we know of this
37 formative time.

38 Few crustal domains are documented to be older than about 3.7 Ga (King, 2005; Condie, 2019).
39 The oldest generally agreed upon suite of rocks is the Acasta Gneiss in the Slave province, Canada at 3.96
40 to 4.03 Ga (Bowring and Williams, 1999; Mojzsis et al., 2014; Bauer et al., 2017; Reimink et al., 2018).
41 All known pre-3.65 Ga terranes have experienced protracted and complex metamorphic and deformation
42 histories that make unraveling their history and the chemistry of their parent melts a challenge. A
43 common thread in studies of Hadean/Eoarchean rocks is the challenge of pinning down the absolute
44 timing of their emplacement as well as their parent lithology (protolith). Perhaps the best known such
45 domain is the Isua Supracrustal Belt (Moorbath et al., 1973; Bridgwater and McGregor, 1974) which is a
46 part of the ca. 3000 km² Itsaq Gneiss Complex in southern West Greenland (Nutman et al., 1996); only
47 correlative petrographic and geochemical studies at the outcrop scale yield confident results that can be
48 used to assign protoliths (e.g. Cates and Mojzsis, 2006; Manning et al., 2006). More recently, other
49 Eoarchean (pre-3.7 Ga) supracrustal enclaves were discovered in the North-East Superior Province in
50 Québec, Canada in the Inukjuak domain (Simard, 2003) (**Fig. 1**). Like the Isua locality, the Inukjuak
51 rocks are metamorphosed to upper amphibolite facies that have experienced multiple generations of
52 deformation and have been intruded and surrounded by orthogneisses and mafic/ultramafic sills (Cates &
53 Mojzsis, 2009; Cates et al., 2013; O'Neil et al. 2018). The best studied of these is the Porpoise Cove
54 outcrops of the Nuvvuagittuq Supracrustal Belt (NSB). This locality is an archetype for supracrustal
55 enclaves that are present in the Inukjuak domain (Greer et al., 2020). The NSB is an arcuate sequence of
56 predominantly Ca-poor amphibolitic rocks along with felsic schists, chemical sediments, quartzites, a
57 metaconglomerate and mafic/ultramafic sills that have all been complexly deformed into an open,
58 southwardly plunging synform.

59 It was previously noted that the NSB is similar in age and in certain geochemical and lithological
60 aspects to parts of the Isua supracrustal belt (Cates & Mojzsis, 2007, 2009; O'Neil et al., 2016).
61 Considering these observations, and a similar evolutionary history laid out by David et al, 2009, it is
62 possible, that these two terranes are the disrupted remnants of what was once a coherent
63 Eoarchean/Hadean block. However, this claim needs further investigation since there are fundamental
64 dissimilarities in the $\mu^{142}\text{Nd}$ values between the two belts (Saji et al., 2018). Although the NSB has been
65 studied structurally and geochemically, there is no consensus, yet, on its geology, age and origin (O'Neil
66 et al., 2007, 2008, 2013; Dauphas et al., 2007; Darling et al., 2013; Cates & Mojzsis, 2009, 2013; David
67 et al., 2009). Multiple hydrothermal and metamorphic events have masked or erased lithological contacts,
68 altered primary igneous and sedimentary geochemical signatures and disturbed isotopic systems at the

69 local to regional scale (Greer et al., 2020). Some of the prominent debates surrounding the NSB are as
70 follows.

71

72 1. The most common rock type in the NSB is a cummingtonite (Ca-poor) amphibolite, with an age
73 of emplacement which remains under debate. Whereas a minimum age limit is established at 3.75
74 Ga from U-Pb zircon geochronology collected from crosscutting orthogneiss sheets in the
75 supracrustals at the Porpoise Cove locality (Cates and Mojzsis, 2007; Cates et al., 2013; cf.
76 Darling et al. 2013). Significantly, O'Neil et al. (2008, 2012) claim that the emplacement age is
77 Hadean with $^{146/147}\text{Sm} - ^{142/143}\text{Nd}$ model ages as old as 4.4 Ga and at least 4.115 Ga. This age is
78 based mostly on analyses on the cummingtonite amphibolite, as well as ultramafic/mafic sills
79 arguably structurally discordant to the amphibolite. O'Neil et al. (2012) have also noted a
80 depletion in the $\mu^{142}\text{Nd}$ and have argued that a partial melting event took place when the ^{146}Sm -
81 ^{142}Nd system was still active; i.e., within the first 300 Myr. Other authors have proposed that the
82 model Hadean ages (>4 Ga) instead reflect mixing lines between Archean melt and older Hadean
83 crustal fragments, or an enriched reservoir (Roth et al., 2013; Guitreau et al., 2013; Boehnke et
84 al., 2018). Hints of a Hadean component were shown by Augland and David, (2015) who
85 propose, using Lu-Hf model ages, that the NSB is at least Eoarchean and may be as old as
86 Hadean. Given the large range of reported Nd model ages for the cummingtonite schist from
87 various studies (3.9 – 4.4 Ga) (David et al., 2009; O'Neil et al., 2007; 2008; 2012), it may also be
88 that these reflect a mixture of provenance ages for the sedimentary protoliths (Cates et al., 2013).
89 What is not in dispute is that the Inukjuak rocks experienced thermal events and extensive
90 hydrothermal alteration that affected the Sm-Nd isotopic ratios and thus, possibly, the isochron
91 dates.

92 2. The NSB comprises a diverse suite of rocks (O'Neil et al., 2007; David et al., 2009; Mloszewska
93 et al., 2012, 2013; Cates et al., 2013; Darling et al. 2013). Although many of these rock units
94 were proposed to have volcano-sedimentary protolith and display extensive hydrothermal
95 alterations including in-place pervasive secondary jaspilite vein development (Greer et al., 2020;
96 cf. Dodd et al. 2017), they remain generally under-explored. Of specific interest are the relatively
97 high whole-rock $\delta^{18}\text{O}_{\text{VSMOW}}$ values (+6.7 to +11.7‰) for a fuchsite-bearing quartzite reported in
98 Cates et al. (2013) which would support a sedimentary protolith interpretation for these rocks.
99 Yet, this inference was challenged by Darling et al. (2013) with claims that the fuchsite-bearing
100 quartzite is a metasomatized orthogneiss, and that such high $\delta^{18}\text{O}$ values have been observed in
101 Archean fuchsitic metasomatic systems.

102 3. Quartz-magnetite rocks identified in the field as banded iron-formations (BIFs) have been
103 analyzed for their $\delta^{56}\text{Fe}$ (Dauphas et al., 2007) values to establish their chemical sedimentary
104 protolith. Follow-on analyses of mass-independent sulfur isotope fractionations ($\Delta^{33}\text{S}$) of the
105 same rocks bolster the claim that these are original marine sedimentary precipitates (Cates et al.,
106 2013; Thomassot et al., 2015).

107 Given the various studies thus far performed on the Nuvvuagittuq rocks and the ongoing debates
108 over the geology and age of the Inukjuak domain components cited above, it makes sense to compare
109 the NSB to other supracrustal enclaves in the same region. Such work would help to establish (or
110 refute) commonality with the NSB (cf. Caro et al., 2017). These other ubiquitous enclaves (Cates et
111 al., 2009; Greer et al., 2020) are practically unknown to the geological community despite the fact
112 that they could yield important new insights into the global Eoarchean Earth. At the very least, their
113 elucidation would help to put the Inukjuak domain into a clearer context viz. various
114 Eoarchean/Hadean provinces of the North American Craton.

115 One such enclave that has been noted (Caro et al. 2017) northeast of the NSB (**Fig. 1**) is the
116 informally-named Ukaliq Supracrustal Belt (USB). The USB is enveloped by the same granitoid
117 gneisses of the Boizard suite (2.7 Ga) as those surrounding the NSB (e.g. Saji et al., 2018; Greer et
118 al., 2020) and with a semi-coherent succession of (mafic) amphibolites, ultramafic rocks of probably
119 komatiite origin, quartz-biotite schists, granitoid (principally trondjemitic) gneisses and fuchsite-
120 bearing quartzites (**Fig. 1; Fig. S1**). The USB is an asymmetrical N-S trending elongate
121 metamorphosed volcano-sedimentary succession, and is one of among many of supracrustal
122 fragments and pods of varying size in the region that have been dubbed the Inukjuak Complex
123 (Simard, 2003). Thus far, $^{146,147}\text{Sm}$ - $^{142,143}\text{Nd}$ studies show deficits in $\mu^{142}\text{Nd}$ (Caro et al., 2017; Saji et
124 al., 2018) for some of the Ukaliq amphibolite units much like those of the Nuvvuagittuq area.

125 Here, we present petrographic thin section studies of select felsic lithological units present in the
126 USB and collected by us in the effort to explore low-T processes that might have affected them.
127 Complementary to this, we present geochemical analyses to constrain the age of the rocks using U-Pb
128 dating of zircons by LA-ICP-MS and secondary ion mass spectrometry (ion microprobe), from
129 targeted samples chosen for their potential to host zircon. Separated zircons from Ukaliq quartz-
130 biotite schists yield the oldest ages recorded, up to 3.79 ± 0.04 Ga. We find that the oldest zircons
131 from the USB are similar in age to detrital zircons documented from the NSB. The $^{207}\text{Pb}/^{206}\text{Pb}$ ages of
132 the zircons analyzed cluster at 3.3 Ga, 3.6 Ga, and 3.8 Ga much like the NSB zircons. To reveal
133 additional information about the origin of these grains, *in situ* zircon $\delta^{18}\text{O}$ and $\delta^{30}\text{Si}$ measurements
134 were also performed. A new technique that couples measurement of Si and O isotopes (Trail et al.,

135 2018) was used to help qualify the nature of weathering and alteration experienced by supracrustal
136 material subsequently assimilated into zircon parent melts. The $\delta^{18}\text{O}$ and $\delta^{30}\text{Si}$ of the zircons suggest
137 that source lithologies that were formed by low to moderate temperature processes (authigenesis,
138 serpentinization, etc.) contributed to the formation of parent melts that crystallized the zircons from
139 the analyzed lithologies. To constrain the emplacement age for these same crystals and the chemical
140 nature and source of their parent melts, Lu-Hf isotope data are also presented. In this way, the
141 emplacement age of the two belts as well as the tectonic setting might be broadly compared. Mineral
142 inclusions in the zircons and the major and trace element composition of the whole rock were studied
143 to explain the nature of the parent melts. Overall, considering the low-T water-rock interactions, our
144 textural and O and Si isotopic observations are consistent with a geologic environment that was
145 suitable for primordial biology to emerge around 3.8 Ga.

146 2. Sample description

147 In our work, further rock samples were collected from a variety of lithological units from an area
148 to the north of the study area of Caro et al. (2017) in the region of the Ukaliq belt (**Table 1**) to make
149 comparative textural observations along with geochemical analyses. The rock samples collected from the
150 USB were orthogneisses (I346, I349, I355, I365 and I373), a hornblende schist (I364), a cummingtonite
151 schist (I369) and quartz biotite schists (I370 and I374).

152 Some samples from the NSB were also collected from which several fuchsite-bearing quartzites
153 (I239 and I240) and are presented herein in order to compare USB and NSB textural observations cited
154 above. It is worth noting that although the USB and the NSB are a few kilometers apart in the Inukjuak
155 domain, they are separated by a shear zone. Superficially, the USB is like the NSB in that it is a sequence
156 of volcano-sedimentary rocks that have been intensely deformed, metamorphosed and intruded as well as
157 surrounded by gneisses; typical for Eoarchean supracrustal enclaves (e.g. Cates and Mojzsis, 2006).

158 The major lithologies in both supracrustal belts comprise medium to high grade metamorphic
159 rocks. **Table 1** lists the rock types discussed in this study. Particular attention was paid to a fuchsite (Cr-
160 rich mica) bearing quartzite found throughout the NSB. Thin section petrographic analysis was performed
161 on samples of this rock to compare with similar fuchsite-rich quartzites identified in the USB. A previous
162 textural analysis of a fuchsitic quartzite from the Porpoise Cove outcrops by Darling et al. (2013) was
163 used by them to conclude that this rock had an unspecified metasomatic origin from an igneous
164 (granitoid) precursor. All rock types examined herein from both supracrustal localities show both local-
165 (e.g. jaspilite box-veinings; Dodd et al. 2017; cf. Greer et al., 2020) and regional effects of hydrothermal
166 alteration either as alterations of the felsic or the mafic minerals. Different thin sections of the USB rocks

167 display evidence of having been affected differently by fluids even though they represent samples that
168 were collected close to each other or even from the same unit (within several meters). The NSB quartzites
169 share similar characteristics in that the micaceous content of the unit changes from location to location.
170 There are also effects of strain recovery amongst quartz grains of the USB.

171 3. Methods

172 3.1 Sample preparation and characterization

173 Fourteen thin sections were made of four distinct lithological units (Table 1) and investigated by
174 optical and electron microscopy. The latter was conducted with a Zeiss Auriga scanning electron
175 microscope (SEM) at the University of Rochester (USA). The instrument is also fitted with an EDAX X-
176 ray spectrometer with mapping capabilities. The units observed were: orthogneisses, amphibolites and
177 quartz-biotite schists. We also investigated two fuchsite-bearing quartzite samples (I239 and I240) from
178 the NSB equivalent to those reported in Cates et al. (2013) and Darling et al. (2013). The two quartzite
179 samples were collected along strike from the same unit about 10m apart.

180 Zircons were extracted from five orthogneisses and two quartz-biotite schists following standard
181 heavy liquid techniques. Briefly: hand samples were crushed in pre-contaminated ceramic mortars,
182 reduced to fine sand with pre-cleaned and pre-contaminated ceramic shatter-boxes, sieved, and heavy
183 minerals (including zircon) separated and concentrated with heavy liquids (MEI). The heavy mineral
184 fraction was further conditioned with a hand magnet and passed through a Franz magnetic separator to
185 create a zircon-rich concentrate of low magnetite susceptibility. Individual zircon grains were hand-
186 picked under a binocular microscope and mounted on double-sided tape and cast into 2.55 cm epoxy
187 rounds along with zircon U-Pb geochronology standard AS-3 (Paces and Miller, 1993). In select cases,
188 zircons were also mounted with zircon Si and O isotope standard KIM (Trail et al., 2018). After curing
189 the epoxy, mounts were polished to expose the grain centers using SiC lapping films with a 3 μm grit
190 size, and then finished to a fine polish with a 1 μm film. Zircons were documented by
191 cathodoluminescence and backscattered electron imaging using the University of Colorado JEOL JXA-
192 8230 electron microprobe, at Laboratoire Magmas et Volcans (Clermont-Ferrand, France) with a
193 Scanning Electron Microprobe (SEM), or at UCLA with the UCLA Tescan Vega-3 XMU SEM equipped
194 with a cathodoluminescence detector. Images were collected of these zircons to interpret internal textures
195 (e.g., discern metamorphic growth rims on zircon cores) and, in turn, determine the best location for *in-*
196 *situ* analysis. Representative BSE and CL zircon images from orthogneisses and quartz-biotite schist
197 samples examined herein are provided in **Fig. 2 & 3**.

198 3.2 LA-ICP-MS U-Pb geochronology

199 Zircons from five gneissic and two quartz-biotite schist units from the USB were analyzed on a
200 Photon Machines Analyte G2 193 nm laser ablation (LA) system attached to an Agilent 7900 Inductively
201 Coupled Plasma Mass Spectrometer (ICP-MS) at the University of Rochester. Epoxy mounts were
202 loaded into a HelEx 2 volume chamber that uses He as the carrier gas. The He flow in the HelEx chamber
203 was kept at 0.6L/min and in the HelEx arm at 0.2 L/min. The laser parameters for the analyses are listed
204 in **Table ST1a**. Some of the grains were large enough for multiple analyses; so, where possible, U-Pb
205 measurements were made on a core, intermediate region, and the rim area of the zircon. For zircons that
206 have multiple analyses, the results from the rim and the intermediate region are presented separately from
207 the core analyses. For each analysis, background counts were collected first (~20 sec), following which,
208 the zircons were ablated. Before entering the plasma, the material ablated along with He is mixed with Ar
209 which was introduced at a flow rate of 1.3 L/min. Post ablation, there was a washout period (~20 sec)
210 before moving on to the next analysis. Zircon ages were harmonized against the known AS-3
211 geochronology standard (Paces and Miller, 1993), and monitored with a secondary in-house standard,
212 Kuehl Lake, believed to be from the same locality as the international zircon standard 91500 (Trail et al.,
213 2018). For the geochronology, the isotopes analyzed were ^{202}Hg , ^{204}Pb , ^{206}Pb , ^{207}Pb , ^{208}Pb , ^{232}Th , ^{238}U , and
214 ^{29}Si . The integration time for ^{206}Pb and ^{207}Pb was 30ms and the for the rest of the isotopes, it was 10ms for
215 each cycle. Silicon-29 was used as an internal standard to determine U and Th concentrations, which were
216 obtained after reduction against NIST612 glass. All data reduction was done with the Iolite 3.32 software
217 package (Paton et al., 2011).

218

219 **3.3 Ion microprobe Si+O isotopes and U-Pb geochronology.**

220 Subsequent to confirmation by geochronology (Sec. 3.2) that some USB samples host pre-3.7 Ga
221 zircons (I370 and I374), a specific mount was made so that these zircons could be analyzed for their Si-
222 and O-isotopic compositions first. The zircons-of-interest were extracted from two quartz biotite schists
223 samples using the techniques described above (Sec. 3.1) and analyzed using the *ims1290* ion microprobe
224 at UCLA following the procedures described in Trail et al. (2018). Briefly, a 3-nA Cs⁺ primary beam,
225 rastering over 20 × 20 μm on the samples, yielded sufficient secondary ion signals (O^{18-} and $\text{Si}^{30-} \geq 6 \times$
226 10^6 and 3×10^6 counts per second, respectively) to be collected with Faraday cups (FCs) in dynamic
227 multicollection mode. This configuration allows for simultaneous measurement of $^{16}\text{O}^-$ and $^{18}\text{O}^-$ on the
228 L/2 and H/2 FCs, respectively, followed by that of $^{28}\text{Si}^-$ and $^{30}\text{Si}^-$ on C and H1 (all FCs) after only one
229 mass jump. The mass resolution (M/ΔM) was set at 2,400 (exit slit #1 on the multicollection trolley) to
230 separate molecular interferences from peaks of interest. One spot analysis is composed of 14 cycles, each
231 of which includes a counting time of 10s for oxygen isotopes, and of 15s for Si isotopes. The

232 backgrounds of FCs were determined during the 90s pre-sputtering before each analysis, and then were
233 corrected for in the data reduction.

234 After the completion of stable isotope measurements, the surfaces of the samples were gently re-
235 polished with special care taken to preserve the location of the Si + O isotope measurement as visible
236 analysis pits, and then re-coated with Au in preparation for ion microprobe geochronology. The U-Pb
237 ages for this subset of zircons were determined on the CAMECA ims1270 ion microprobe at UCLA using
238 a ca. 15 nA primary O⁻ beam generated from the duoplasmatron focused to a ca. 20 μm spot at a mass
239 resolving power of 5000 (e.g. Bell et al., 2017; Quidelleur et al., 1997). To increase Pb sensitivity the
240 analysis chamber was flooded with oxygen. Zircon U-Pb ages were standardized against AS-3, and most
241 likely source of common Pb contamination was from sample preparation. Thus, all data were corrected
242 for common Pb contamination using Pb isotope data reported for California surface waters (Sañudo-
243 Wilhelmy and Flegal 1994). Note that the ion microprobe geochronology was conducted *after* Si and O
244 isotope investigations to prevent contamination of exogenous O in zircon from the primary ion source.
245 For these samples, the ion microprobe was used in preference to the LA-ICP-MS to preserve enough of
246 the grain for Lu-Hf analyses.

247

248 **3.4 Lu-Hf**

249 The Lu-Hf isotope analyses were performed at LMV (Laboratoire Magmas et Volcans, Clermont-
250 Ferrand, France). Details regarding operating conditions are available in **Table ST3**. Zircons were
251 analyzed in batches of 10 unknowns (4 standards and 6 samples) bracketed by the synthetic MUN zircon
252 standard, which is doped in heavy REEs (MUN-0; Fisher et al., 2011). Standards analyzed as unknowns
253 (91500, MudTank, and two MUN zircons with low and high Yb/Hf ratios) served as quality control. For
254 zircons from Ukaliq samples reported herein, Lu-Hf spots were located on the trace pits of previous U-Pb
255 measurements. Both Yb and Hf instrumental mass biases were determined using ¹⁷³Yb/¹⁷¹Yb normalized
256 to the value of 1.129197 (Vervoort et al., 2004) and ¹⁷⁹Hf/¹⁷⁷Hf normalized to 0.7325 (Stevenson and
257 Patchett 1990). The Lu fractionation was assumed to follow that of Yb (Fisher et al., 2014), and values of
258 0.793045 for ¹⁷⁶Yb/¹⁷³Yb (Vervoort et al., 2004) and 0.02655 for ¹⁷⁶Lu/¹⁷⁵Lu (Fisher et al., 2014) were
259 used to remove isobaric interferences of ¹⁷⁶Yb and ¹⁷⁶Lu at mass 176. We carefully monitored each time-
260 resolved signal to minimize issues of domain mixing and grain heterogeneities. Owing to the fact that Hf
261 isotope standards tend to show a slight shift in ¹⁷⁶Hf/¹⁷⁷Hf after mass bias correction, we have used MUN-
262 0 to determine an offset factor and correct all LA-MC-ICP-MS analyses for this mismatch. Results
263 obtained on standards comport well with consensus values. The average ¹⁷⁶Hf/¹⁷⁷Hf measured in 91500

264 and MudTank were, respectively, 0.282307 ± 0.000041 (2 std; n = 7) and 0.282495 ± 0.000035 (2 std; n =
265 6), which agree well with consensus values of 0.282308 ± 0.000008 (Blichert-Toft, 2008) and $0.282507 \pm$
266 0.000006 (Woodhead and Hergt, 2005). The average $^{176}\text{Hf}/^{177}\text{Hf}$ measured in MUN_1-2b and MUN_4-2b
267 were 0.282133 ± 0.000037 (2 std; n = 7), and 0.282126 ± 0.000018 (2 std; n = 6) respectively, therefore
268 fully consistent with the value of 0.282135 ± 0.000007 reported in Fisher et al (2011). We have used the
269 ^{176}Lu decay constant value of Scherer et al. (2001) and Söderlund et al. (2004), which is $1.867 \times 10^{-11} \text{ yr}^{-1}$,
270 to calculate initial Hf isotopic compositions. Uncertainties associated with the radiogenic-ingrowth
271 correction were propagated using the algorithms of Ickert (2013). Finally, ϵ_{Hf} values were calculated
272 using CHUR values provided in Iizuka et al. (2015).

273

274 **3.5 Major and Trace elements**

275 Chips of two quartz-biotite schists (I370 and I374) were sent out to Hamilton Analytical
276 Laboratory, Clinton, NY (USA) to characterize major and trace element contents by X-Ray Fluorescence
277 (XRF) spectrometry. The chips were chosen for minimal alterations and were visibly as devoid of saw
278 marks as possible. Before processing the samples, saw marks were removed so that residual metal saw
279 particles would not interfere with the measurements. Chips were then milled to a fine powder using a
280 Rocklabs tungsten carbide (WC) or alumina ring mill and fused with a Li-tetraborate flux in a graphite
281 crucible at 1000 °C to form a pellet. These pellets are cleansed of residual carbon, reground and fused
282 again at 1000 °C. The resulting pellets were polished using diamond laps to create polished pellets 29mm
283 in diameter which were then analyzed using the Thermo ARL Perform'X XRF spectrometer.

284

285 **3.6 Mineral Inclusions**

286 Several zircons from two quartz-biotite schists (I370 and I374) were imaged using a Vega 3XMU
287 scanning electron microscope (SEM) at UCLA. Both EDS and WDS X-ray spectroscopy were performed
288 on the SEM and a JEOL 8600 Electron Microprobe respectively to identify mineral inclusions in the
289 zircons. The inclusions were classified in three separate groups. First, some inclusions were classified as
290 “isolated”; i.e., they were not in contact with any cracks and are thus considered primary. The second
291 group of inclusions is classified as is “possibly secondary minerals” as they were associated with cracks,
292 which are avenues of fluid intrusion. Third, those assigned as “crack-filling inclusions” are most certainly
293 secondary. The counting protocol employed, which characterizes the diversity of inclusions found, is that
294 established by Bell et al., (2015) and Bell, (2016) where the mere occurrence of a phase in a zircon is

295 regarded as a single occurrence regardless of the number of inclusions. For example, three crack-isolated
296 apatite inclusions in a single zircon contributes to a single count of crack-isolated apatite inclusions (Bell
297 et al., 2018). The comprehensive list of inclusions detected is summarized in **ST4**.

298 **4. Results**

299 **4.1 Thin section observations**

300 To summarize, the major lithologies are medium- to high-grade metamorphic rocks that range
301 from biotite schists to amphibolites all the way up to hydrous mineral bearing high-grade gneisses (**Fig. 4**
302 **a-c**). A biotite-fuchsite bearing quartzite unit was also sampled from the NSB as a point of comparison
303 with regards to hydrothermal alteration in the Inuakjuak domain (**Fig. 4d**). All these rock types show
304 effects of hydrothermal alteration either as alteration of the felsic or the mafic minerals or deposition of
305 secondary minerals or both.

306 **4.1.1 NSB Biotite-fuchsite bearing quartzite from the NSB (I239 and I240)**

307 This study samples two sections of a coarse-grained metamorphic rock from the NSB that is
308 predominantly quartz with minor plagioclase occurrences (I239 and I240) (**Fig. S2 d**). Quartz grains show
309 grain boundary migration and sub-grain formation. The plagioclase has been altered to sericite. Fuchsite
310 is present along the margins of the felsic grains but do not show distinct parallel foliation. One of the
311 sections (I239) contains a greater proportion of fuchsite and biotite even though they are from the same
312 rock unit. This terrane is non-uniform in terms of its pattern of hydrothermal alteration. Both these
313 sections contain as accessory phases, chromites, monazites and zircons. Consequently, we view the likely
314 protolith of this rock as a detrital sand rich in heavy minerals.

315 **4.1.2 USB Quartz-biotite schists (I370 and I374)**

316 These rocks are medium- to coarse-grained metamorphic rocks with continuous schistosity
317 defined by mafic and felsic layers (**Fig. S1a**). The mafic layers are defined by biotite that has been altered
318 by varying degrees to chlorite, Fe/Ti-oxides/sulfides and/or rutile (**Fig. S1b & c**). The layers show rough
319 foliation where the biotite crystals are disjoint grains. These micaceous minerals have inclusions of
320 apatite, zircon and monazite identified by radiation halos. The felsic domains comprise anhedral to
321 subhedral grains of plagioclase and quartz. In some sections, the plagioclase shows deformation twinning,
322 and the grain boundaries of both quartz and plagioclase are defined by a fine-grained matrix of mica and
323 similar phases as the felsic domains. This is probably the result of a combination of hydrothermal
324 alteration and recrystallization through pressure solution where grains having sharp edges with low
325 surface area are in contact thereby leading to high pressure regions. This causes material to be displaced

326 from these regions to surrounding low pressure regions. The quartz in all sections shows grain boundary
327 migration and sub-grain formation with abundant triple-junctions with pinning by accessory minerals
328 (**Fig. S1d & e**). Some of the felsic grains contain inclusions of mica which could be foliation from a
329 previous generation of deformation or relict grain boundaries that have been consumed by mineral growth
330 (**Fig. S1e**). There are also inclusions of quartz in many plagioclase crystals which may be myrmekitic or
331 an effect of re-melting and recrystallisation of feldspar. More detailed SEM investigations have revealed
332 the presence of monazite, native Pb, allanite, and titanite.

333 **4.1.3 USB Amphibolites (I364 and I369)**

334 Two kinds of amphibolites were identified: a hornblende-bearing (Hb-amphibolite; **Fig. 4e**)
335 (I364) and a cummingtonite-bearing (low Ca) type (Cum-amphibolite; **Fig. 4f**) (I369). Both show signs of
336 alteration with the hornblende being extensively altered to chlorite, biotite and Fe/Ti-phases and the
337 cummingtonite is sparingly altered to not altered at all. The amphiboles mentioned define the mafic layers
338 of all the schists. In the Hbl-amphibolite, biotite crystals flow around porphyroblasts (**Fig. S1f**) (no fine-
339 grained matrix surrounding the porphyroblasts) suggesting that they developed after the plagioclase
340 formed. The texture of the cummingtonite-amphibolites changes from one thin-section to another. The
341 grain size reduces, and the grains are more disjoint from one to another. The felsic domains comprise
342 plagioclase and quartz where the plagioclase crystals have been partially sericitized. Observations made
343 under an SEM show titanite, ilmenite, rutile, and zircon.

344 **4.1.4 USB Granitoid (ortho-)gneisses and granitic gneisses (I346, I349, I355, I365, and I373)**

345 The sampled granitoid gneisses are all mica-bearing tonalitic (I346 and I355) or granitic gneisses
346 (I349 and I365) (**Fig. 4a**). They are medium to coarse grained and the minerals are subhedral to anhedral.
347 There are both felsic and mafic minerals in all rocks sampled. The minerals are dominantly plagioclase,
348 quartz, alkali feldspar in some rocks and biotite. There are accessory phases such as apatite and zircon
349 present as inclusions in the biotite. The biotite has also been hydrothermally altered in multiple locations
350 to chlorite and Fe/Ti ore phases. In certain sections, the plagioclase has not been altered while the biotite
351 has (**Fig. S2a**). In some other sections the opposite has been observed (**Fig. S2b**). In the granitic gneisses
352 there are both alkali feldspar and plagioclase. In certain locations, myrmekite textures are visible on the
353 border of two feldspars. The tubular quartz might be considered evidence for metasomatism (**Fig. 4i**). In
354 all the gneisses, the quartz grains show recovery and recrystallisation by bulging, grain boundary
355 migration and sub-grain formation. In some locations, the quartz shows ribboning.

356 In one of the sections (I373), anhedral alkali feldspar grains in some locations are surrounded by
357 plagioclase (**Fig. S2c**). This may either be because alkali feldspar transformed to plagioclase or because

358 the alkali feldspar is secondary and crystallized from infiltrating fluids. SEM investigations reveal zircon,
359 ilmenite, monazite, and xenotime.

360 4.2 USB U-Th-Pb Geochronology (LA-ICP-MS and SIMS)

361 The $^{207}\text{Pb}/^{206}\text{Pb}$ ages of zircons and their Th/U ratios are summarized in **Table 2** (Complete
362 dataset in **Table S1b**). The latter can be used cautiously as an indication whether the analysis sampled a
363 metamorphic zone in the zircon (e.g., Mojzsis and Harrison, 2002; Rubatto, 2002). Metamorphic zircons
364 generally have widely variable Th/U; i.e., <0.001 or >10 (e.g., Yakymchuk et al., 2018), whereas igneous
365 zircons tend to have values around 0.8 (e.g., Kirkland et al., 2015). Thus, Th/U as a differentiating
366 parameter is better used in conjunction with CL and BSE images to consider whether zircon domains are
367 more consistent with igneous or metamorphic processes. Upon observing the zircons from the gneisses
368 and the schists, noticeable differences appear. Zircons from both rock types show oscillatory zoning in the
369 core but the zircons from the quartz-biotite schists have a more regularly occurring and prominent growth
370 rims (**Fig. 3**). The gneiss zircons either have very thin or non-existent rims. The Th/U ratio of the core of
371 the gneiss zircons have a greater spread (0.02 – 1.48) compared to the schist zircons (0.02 – 0.74) (**Fig. 5**)
372 but they have a similar Th/U ratio at an average of 0.4 compared to the schist zircons at 0.45. This is
373 dissimilar and opposite to the observations made on the NSB units (Cates and Mojzsis, 2007) both in
374 terms of spread and absolute values (**Fig. 6**). The zircons from the NSB gneisses, analyzed by Cates and
375 Mojzsis, 2007, have a smaller spread (0.01 – 0.85; average = 0.54) compared to the quartz-biotite schists
376 (0.03 – 2.19; average = 0.65). Thus, on average, based on the CL images and the Th/U ratios, zircon cores
377 in the USB orthogneisses or schists are consistent with a magmatic origin. The analyses considered when
378 assigning ages for each of the rock units, based on zircon U-Pb geochronology, (**Table 2**) were analyses
379 with U-Pb concordance between 90 and 100% and closest to the core of the grains. These grains have
380 been presented on a concordia diagram (**Fig. S3**). The age histograms, on the other hand, (**Fig. 5 & 6**)
381 present grains with % discordance

$$382 \left(1 - \frac{Pb^{206}}{U^{238}} / \frac{Pb^{207}}{U^{206}} \right) * 100 \text{ between } -10 \text{ and } 10.$$

383 4.2.1 USB orthogneissic units (I346, I349, I355, I365 and I373)

384 The $^{207}\text{Pb}/^{206}\text{Pb}$ ages for the gneissic zircons, when categorized based on their host rock and
385 averaged – except for I373 – range from 3.26 to 3.34 Ga. When the $\leq 10\%$ discordant zircons are plotted
386 on a concordia diagram (**Fig. S3a**), most of them cluster between 3.5 and 3.0 Ga implying that this time
387 period accounts for most of the sampled igneous non-disturbed zircons. The youngest age for a zircon is
388 from I355 at ca. 2.4 Ga. This analysis also has a very high Th/U ratio at 1.49 which might point to Pb

389 loss. The I373 zircons need to be mentioned separately because of all the analyses on zircons from this
390 unit, only two analyses of the intermediate region between the core and the rim is between 90 and 100%
391 concordant. The maximum I373 zircon age is 3.68 ± 0.07 Ga which may be considered as the
392 emplacement age of the orthogneisses.

393 4.2.2 USB quartz-biotite schists (I370 and I374)

394 The ages obtained from the zircons of the USB schists are between 2.73 – 3.77 Ga which is like
395 the quartz-biotite schists of the NSB (3.5 – 3.8 Ga) (David et al., 2009; Cates et al., 2007, 2009 and
396 2013). The maximum ages of zircons from the two quartz-biotite schists are 3.78 ± 0.08 and 3.79 ± 0.04 .
397 This implies that the protoliths of these schists incorporated material that is at least about 3.8 Ga. Zircons
398 from these schists were also measured using SIMS which gave much more precise maximum ages at 3.76
399 ± 0.01 and 3.74 ± 0.01 (2 s.e.) Ga. The quartz-biotite schist analyses show a distribution with a mode at
400 around 3.7 Ga with a few outliers around 2.7 Ga. Some of the outliers have a degree of concordance
401 below 90% implying Pb loss. Younger ages, defined here at 2.7 Ga, have corresponding low Th/U ratio
402 ($\sim 0.1 - 0.01$) when compared to older (< 3.0 Ga) ones. The concordant zircons that have been analyzed
403 with an LA-ICP-MS show a tight cluster at $\sim 3.6-3.8$ Ga (**Fig. S3b**).

404 4.2.3 Zircon zonal analyses

405 Zircons from some of the gneisses were analyzed zonally; i.e. the core, intermediate, and rim
406 sections of zircons from individual gneissic units. All the values from each of the zones were averaged to
407 investigate at the variation in the age and Th/U ratio within a single average zircon grain (**Fig. 7**). For
408 most of the units, the ages of the zones are similar (~ 3.3 Ga) except for I365 (Gneiss) which has rim and
409 intermediate ages (~ 3.45 Ga) greater than core ages (~ 3.1 Ga), but this may be because the I365 grains
410 have more cracks than grains from other units. The Th/U ratio of the rim measurements for some of the
411 rocks (I349 and I355) is also lower than the core values.

412 4.3 In-situ Si and O isotopes in USB zircon.

413 The zircons from the two quartz biotite schists (I370 and I374) are well scattered both in terms of
414 $\delta^{18}\text{O}$ and $\delta^{30}\text{Si}$, with latter values of both units having a larger range from -0.07 to -0.75 ‰. Most of the
415 zircons are indistinguishable from the $\delta^{30}\text{Si}$ mantle values of -0.38 ± 0.02 ‰ (**Fig. 8**). Zircons from both
416 samples show $\delta^{18}\text{O}$ (relative to VSMOW) to be distinct from the mantle value of 5.3 ± 0.3 ‰, with values
417 that range from 5.6 to 6.6 ‰ (**Table ST4**). All data collected are reported but of all the zircons measured
418 for their isotopic values (**Table ST4**), only the ones for which the analytical location of the ion
419 microprobe spot did not hit a crack, or an inclusion (determined using SEM images) are discussed in

420 detail. The rejected values are not anomalous from the rest of the dataset except for a couple of analyses
421 that show sub-mantle values for both isotopic systems. The samples that were rejected was done so
422 because we found cracks in the analysis location post-sputtering.

423 **4.4 Lu-Hf in USB zircon**

424 Zircon separated from quartz-biotite schists (I370 and I374) were analyzed for their Lu-Hf
425 isotope systematics and corresponding data are reported in **Table ST7**. These data have been acquired as
426 a means of comparison between the NSB and the USB, and to discuss the sources of zircon parental
427 melts. The Lu-Hf isotopic system in zircon is particularly effective in obtaining the Hf isotope
428 composition of the source to zircon parent melt since this mineral incorporates substantial amounts of Hf
429 (up to 1 wt%) whereas it only accommodates a minimal amount of Lu. This leads to minimal growth of
430 radiogenic ^{176}Hf over time; thus, the present-day $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of the zircon very closely resembles the
431 $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of the parent melt from which it crystallized. The $^{176}\text{Lu}/^{177}\text{Hf}$ ranges between $0.000428 \pm$
432 0.000018 and 0.001126 ± 0.000018 . The $^{176}\text{Hf}/^{177}\text{Hf}$ ratio varies between 0.280304 ± 0.000021 and
433 0.280413 ± 0.000031 . For both these ratios, the scatter is similar to that of previous studies performed on
434 the zircons from the NSB (**Fig. 10a & b**). Finally, the initial ϵ_{Hf} ($\epsilon_{\text{Hf}(t)}$) range from -22.7 ± 0.8 to $+1.9 \pm$
435 1.1 and data altogether define a positive correlation between $\epsilon_{\text{Hf}(t)}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages.

436 **4.5 Major and trace elements of I370 and I374 (USB).**

437 The major element data is plotted along with the NSB quartz-biotite schists (IN05004, IN05005,
438 IN05020, IN05037, IN08036) of Cates et al. (2013) (**Fig. 11**). This was done to look for any trends and
439 similarities when plotted as Harker diagrams. Any subsequent and inferred trends provide supporting
440 evidence that NSB and USB schists might be consanguineous. Also, major element compositions for
441 averaged Witswatersrand shales (WS) (Wronkiewicz and Condie, 1987), metasediments from the Isua
442 supracrustal belt (MISB) (Bolhar et al., 2005) and Archean platforms/shields (APS) (Rudnick and
443 Fountain, 1995) are presented alongside the schists from the USB. The WS is an Archean mafic shale,
444 thus comparisons were made to qualify the felsic nature of the USB schists while the Isua and the
445 Archean platforms data is presented to act as a point of comparison between Archean terranes. Trace
446 element spider diagrams, normalized to the NASC (North American Shale Composite) (Gromet et al.,
447 1984), are presented for schists from the USB along with the mafic averaged Witswatersrand Shales, Isua
448 metasediments and Archean platforms for reference (**Fig. 10**). The NASC is supposed to represent the
449 average upper continental crust trace element composition. The measurements show that the rocks are
450 high in SiO_2 (~70%) and low in mafic oxides MgO , Fe_2O_3 , MnO and TiO_2 (~2%, ~3.5%, ~0.035% and
451 ~0.35% respectively) compared to the apparently more mafic NSB rocks, the WS, MISB and the APS.

452 The CaO and K₂O contents of the NSB and USB schists are comparable while the USB Na₂O content is
453 higher. When the NSB and USB rocks are plotted together in Harker diagrams, MgO, Fe₂O₃, TiO₂, Al₂O₃
454 and K₂O show inverse correlations while only Na₂O arguably shows a positive correlation. The mafic
455 oxides (MgO, Fe₂O₃, MnO and TiO₂) are also plot against Al₂O₃ and except for TiO₂, which shows a
456 strong correlation, they show a weak to moderate positive correlation. The trace element measurements
457 have been presented as a comparison between the USB and the NSB while the other Archean lithologies
458 act as points of reference- The entire major and trace element data set is presented in **Table ST8**.

459 **5. Discussion**

460 Eoarchean supracrustal rocks are a rare commodity, possibly because during their long (>3.6 Ga)
461 crustal residence times they have been lost due to erosion combined with some form of crustal recycling.
462 Those that remain have – without known exception – undergone extensive metamorphism and
463 deformation. Such a history has all but erased primary textures and thus has made understanding the
464 surface processes active on Earth prior to 3.6 Ga, vague. Given the present-day close association of the
465 NSB and the USB, a reasonable question is whether and to what extent these Eoarchean supracrustal
466 sequences share any similarities, or on the other hand, if they have differences, what new information is
467 revealed about the Eoarchean Earth? Common themes among the two belts include hydrothermal
468 alteration, secondary mineral deposition, similar metamorphic grades and lithologies, similarities in U-Pb
469 ages and Lu-Hf isotopic ratios.

470

471 **5.1 General texture of the Inukjuak rocks**

472 The NSB fuchsite-bearing quartzite (I239 and I240) has a chaotic fuchsite pattern that is not
473 restricted to uniform and regular layers but rim the felsic grains which implies that the fuchsite might
474 have been a later hydrothermal deposit. Since this rock is a quartzite, its protolith is most probably a
475 sedimentary rock, yet the outcome of the work of Darling et al. (2013) claims an igneous protolith. Most
476 authors prefer Archean fuchsite-bearing quartzites to have an uncomplicated origin from sedimentary
477 protoliths (Raza et al., 2010; Maier et al., 2012; Nutman et al., 2014). Similar debates might arise when
478 the USB is studied in detail; such debates might likewise require comparative studies. Our observations
479 are consistent with the scenario that the Cr in the micas is likely derived from the partial dissolution of
480 chromites documented to present in the quartzite. Electron imaging has revealed chromites showing a
481 thatched texture surrounded by fuchsite (**Fig. S6**) which lends supports this interpretation (Cates et al.,
482 2013). Since chromite is generally not a primary phase in a granitoid, the protolith of these rocks are
483 more likely to be sedimentary and not a metasomatized orthogneiss as postulated by Darling et al. (2013).

484 Different thin sections like the orthogneisses (I346, I349, I355, I365 and I373) have evidence of
485 different degrees of hydrothermal alteration even though they represent samples that were collected close
486 to each other, or even from the same unit. In some sections, we observe in thin section that the plagioclase
487 is altered while the biotite is relatively pristine while in other samples, the opposite is observed. The NSB
488 quartzite shares similar characteristics with the USB quartz-biotite schists in that the micaceous nature of
489 the unit changes from location to location. There are also effects of strain recovery amongst quartz grains.
490 The quartz-biotite schists are unlike those of the NSB in that they do not contain any porphyroclasts of
491 quartz that have led authors to suggest that the NSB schists might be metaconglomerates (Cates and
492 Mojzsis, 2007). Apart from this dissimilarity they are similar in metamorphic grade and mineralogy. The
493 amphibolites and gneisses show moderate to extensive hydrothermal alteration of the mafic minerals
494 implying metasomatism, while the quartz grains show grain boundary migration, subgrain formation and
495 even ribboning in certain locations (**Fig. 6**), like the NSB. The ribboning suggests that the gneisses
496 experienced crystal-plastic deformation. The presence of cummingtonite amphibolites is also a different
497 observation from that made by Caro et al. (2017) on a part of the belt further south with difference
498 exposure, who observed that such rocks are rare in the USB. Some of the gneisses (I346; oldest zircon
499 age: 3.50 ± 0.05 Ga and I349; 3.49 ± 0.04 Ga) show myrmekite textures which indicate hydrothermal
500 alteration. Coupled observations made on mafic and felsic minerals in various gneissic sections imply
501 both a K-deficient fluid (chloritized biotite and relatively unaltered plagioclase) and a K-saturated fluid on
502 a different thin section (sericitization of plagioclase but unaltered biotite) (**Fig. S2c**). Based on these
503 observations and similarities with the NSB the protolith for the USB quartz-biotite schists are consistent
504 with a sedimentary origin (Greer et al., 2020). These observations pointing toward extensive and varied
505 metasomatism is corroborated by Caro et al. (2017) who documented a high degree of serpentinization in
506 southern section of the USB. The alterations are evidence that both supracrustal belts experienced
507 hydrothermal alteration.

508 **5.2 Evolution of the Inukjuak Domain: U-Pb geochronology, CL and BSE images and internal** 509 **zoning of USB zircons**

510 Several timelines for the evolution of the Inukjuak Domain have been proposed that use different
511 geochronological techniques applied to the NSB (Cates and Mojzsis, 2007, 2009; David et al., 2009;
512 O'Neil et al., 2008, 2012, 2016). While $^{142}\text{Nd}/^{144}\text{Nd}$ isochron ages of multiple units as well as negative
513 $\mu^{142}\text{Nd}$ values (up to -18 ppm for cummingtonite amphibolites; O'Neil et al., 2012) have been used to
514 argue that the NSB has a Hadean formation age (O'Neil et al., 2008, 2012 and 2016). This conclusion has
515 not been corroborated by other systems like the ^{147}Sm - ^{143}Nd and the ^{176}Lu - ^{177}Hf which report the
516 formation ages to be Eoarchean (Roth et al., 2013; Cates et al., 2013; Guitreau et al., 2013), which is in

517 turn supported by the in-situ U-Pb ages from Cates and Mojzsis, (2007) and David et al. (2009).
518 However, O'Neil et al. (2012) measured some mafic sills intruding into the cummingtonite amphibolites
519 to find their $^{143}\text{Nd}/^{144}\text{Nd}$ isochron ages are Hadean which according to them sets the minimum age of the
520 NSB at 4.1 ± 0.1 Ga. We emphasize, however, that the general de-coupling between these chronometers
521 was previously explained by Roth et al. (2013) to be a result of an Eoarchean thermal event that caused
522 the resetting of Nd isotopic ratios. As such, an accurate extraction age *cannot* be obtained using the Sm-
523 Nd chronometers. The “Hadean” model age and the negative $\mu^{142}\text{Nd}$ have been explained by authors to be
524 signatures from remnant Hadean crustal fragments or an enriched mantle source (Roth et al., 2013; Cates
525 et al., 2013; Boehnke et al., 2018). Guitreau et al. (2013) has also suggested the re-fertilization of the
526 mantle source that created the NSB.

527 Before making inferences about the ages measured on our zircons, they need to be identified and
528 confirmed as igneous (emplaced) rather than hydrothermal/metamorphic (neoforn). The zircons
529 discussed in this study all show oscillatory zoning in the grain core, consistent with an igneous origin.
530 Our age range from five different gneissic units is between 3.26 and 3.68 Ga. This implies that the USB
531 records multiple generations of magma formation or that the younger ages might possibly be recording a
532 later metamorphic event or be artifacts of metamorphism. The latter may not be the case because in
533 reporting the average or emplacement ages, highly concordant zircons were selected (<10% discordant).
534 The oldest ages from the gneisses range from 3.47 – 3.68 Ga which, if considered as an emplacement age
535 for the protolith, are younger than the oldest ages reported for the NSB. This could mean that the NSB
536 gneisses record an earlier generation of zircon crystallization. However, this would not be the case if the
537 USB zircon ages mentioned above are not igneous. The average Th/U of the gneissic zircons range
538 between 0.26 and 0.6 which is consistent with a zircon having an igneous origin. The Th/U values
539 however do not have a tight cluster (**Fig. 5**) which could be because the zircons retain a record of
540 metamorphism. The orthogneiss zircons analyzed by Cates and Mojzsis, (2007) also have similar Th/U
541 ratios between 0.51 and 0.74. In their depth profiling study, Cates and Mojzsis, 2009 obtained Th/U ratios
542 of 0.41 and 0.31 from the core of the two zircons that they analyzed from a TTG unit. This is close to the
543 values for our grains.

544 Two quartz-biotite schists were analyzed in this study (I370 and I374). The oldest grains have
545 ages of 3.78 ± 0.08 and 3.79 ± 0.04 Ga from the corresponding units. When compared to the zircons
546 analyzed by Cates and Mojzsis, 2007, our zircons seem to be slightly older than those from the NSB
547 quartz-biotite schists. Since these older rocks do not possess porphyroclasts as the quartz-biotite schists of
548 the NSB, these units might be a completely new unit discovered in the region and might have experienced
549 different environments from the lithologically similar rocks in the NSB. The difference is observed in the

550 Th/U values as well. The zircons analyzed by Cates and Mojzsis, 2007 from quartz-biotite schists
551 (IN05020 and IN05037) in the NSB show a greater range in Th/U which might mean a more
552 inhomogeneous parent melt or sampling of multiple parent melts. The average Th/U ratio of both the
553 USB units is 0.44 ± 0.06 and 0.52 ± 0.07 which seems to be in line with a magmatic origin. The Th/U
554 ratios of the USB zircons do not have a large spread (**Fig. 6**) which is consistent with a more homogenous
555 parent melt. The oldest U-Pb age thus far reported is 3.82 ± 0.02 Ga which is an upper intercept age on a
556 Wetherill Concordia diagram. This was determined by David et al. (2009) from a group of six zircons
557 from a quartz-biotite schist using the ID-TIMS method. This age is comparable to the minimum
558 emplacement ages presented for the gneisses from the NSB at 3.75 ± 0.01 Ga (Cates and Mojzsis, 2007)
559 which suggests that these two rock types might be coeval. This observation has also been made for rocks
560 from southern West Greenland (Nutman et al., 2004) where Archean orthogneisses and metasediments
561 contain zircons that range in age from 3.65 to 3.9 Ga. The ages reported here also overlap the ages for a
562 clast-free pelitic schist from the Isua Supracrustal Belt; which is 3.78 ± 0.01 Ga (Nutman et al., 1996).

563 Finally, we would like to mention that we consider the maximum depositional ages for both schist
564 samples to be around 3.65 Ga, and not that of the oldest zircon grains. This is because the Th/U and the %
565 concordance are 0.43 and 100. Also, the CL and BSE images of I370_3_7 reveal distinct oscillatory
566 zoning consistent with (but not proof of) an igneous origin (**Fig. S7**). The laser-ablation age for this grain
567 is 3.65 ± 0.06 Ga. Any zircons that are younger than this age, have been affected by Pb-loss and show a
568 100% concordance. There are older zircons showing magmatic features and greater concordance but if
569 they were to designate depositional ages, then I370_3_7 should be classified as metamorphic which is
570 hard to argue given the oscillatory zoning that it possesses.

571

572 **5.3 Lu-Hf isotopes**

573 Similar to the overall objective of the study, Lu-Hf systematics was used to establish similarities
574 or differences between the rocks of the NSB and the USB, as well as to deduce information about their
575 formation and evolution. Zircons in this study have been compared to NSB in terms of the chemical
576 nature of the host rock sources and the location of the source. Zircons and whole rocks from the NSB
577 have been analyzed for their isotopic ratios so there exists a substantial database against which
578 comparisons may be made. **Fig. 10** shows that data from this study have similar Hf isotope signatures, U-
579 Pb ages, and, hence, evolutionary pattern to NSB data from previous studies (Guitreau et al., 2012, 2013;
580 O'Neil et al., 2013; Augland and David, 2015). This observation strongly suggests a shared history
581 between USB and NSB.

582 The evolutionary pattern is similar to what would be observed if U-Pb ages were affected by
583 ancient disturbances as argued by the good match between the general correlation and the evolution lines
584 corresponding to measured $^{176}\text{Lu}/^{177}\text{Hf}$ in zircons (0.0004-0.002; **Fig. 10a and 10b**). Moreover, precise
585 ID-TIMS ages reported in Augland and David (2015) clearly indicate that non-zero-age disturbance
586 occurred in the NSB much like what is often seen in Archean zircons (e.g., Patchett, 1983; Gerdes and
587 Zeh, 2009; Guitreau et al., 2012, 2019). This results in U-Pb ages being younger than that of their actual
588 crystallization and associated ϵ_{Hf} values being artificially too low (e.g., Guitreau and Blichert-Toft, 2014).
589 Therefore, this consideration together with the fact that CL images display metamorphic recrystallization
590 textures would argue in favor of some of the datapoints representing this artifact. We estimate the
591 maximum depositional age of I370 and I374 being around 3.65 Ga and, hence, consider that zircons older
592 than this are magmatic and all those younger are artifacts of ancient U-Pb disturbance.

593 The zircon data presented by other authors, especially O'Neil et al. (2013), nevertheless, illustrate
594 that this seemingly artificial correlation (**Fig. 10b**) is similar to that exhibited by data for magmatic
595 zircons, and hence, represents the true evolutionary pattern of NSB and USB zircon-bearing rocks (**Fig.**
596 **10d**). This slope is equivalent to a $^{176}\text{Lu}/^{177}\text{Hf}$ typical of TTGs such as those present in the NSB. O'Neil et
597 al. (2013) argued that the slope is too steep to be accounted for by the NSB only being made of Archean
598 rocks, and used it as an evidence for the presence of Hadean rocks in the NSB. In fact, these authors
599 assumed that the slope should be equivalent to a $^{176}\text{Lu}/^{177}\text{Hf}$ typical of a basaltic rock because TTGs are
600 reworking products of metabasalts (e.g. Moyen and Martin, 2012). Although, this reasoning has some
601 petrological logic on a first order, the slope of such pattern cannot be interpreted at face value because, in
602 practice, seemingly simple evolutionary patterns can hide complex petrological processes that are
603 geochemically buffered by dominant enriched crustal lithologies. This was well illustrated recently by the
604 work of Laurent and Zeh (2015) who showed that felsic rocks of the Pietersburg block (South Africa)
605 define a simple evolutionary pattern in a ϵ_{Hf} versus age diagram despite these rocks being formed in
606 various geodynamic contexts and from several distinct sources. As a consequence, the combined NSB-
607 USB evolutionary pattern can merely be interpreted as being buffered by dominant ~3.7-3.8 Ga granitoids
608 of TTG affinity because the slope defined by data in **Fig. 10d** is equivalent to that of the felsic lithologies
609 in the NSB ($^{176}\text{Lu}/^{177}\text{Hf} = 0.002$; Guitreau et al., 2012, 2013) and fully consistent with TTGs worldwide
610 (Guitreau et al., 2012). Furthermore, the near-chondritic initial Hf isotope signature of the 3.7-3.8 Ga
611 zircons comply very well with that of global TTGs (e.g., Guitreau et al., 2012 and references therein).
612 Finally, the Hf isotope signatures of NSB amphibolites (**Fig. 10d**) are consistent with the 3.7-3.8 Ga
613 TTGs being generated by the reworking of these amphibolites, which was demonstrated experimentally
614 by Adam et al. (2012) who essentially reproduced earlier experiments on TTG petrogenesis (e.g., Rapp
615 and Watson, 1995). In a nutshell, available Hf isotope data do not require the involvement of a Hadean

616 reservoir to account for the evolutionary pattern of NSB and USB, and the following magmatic history
617 can be outlined:

- 618 1- Extraction of mafic rocks equivalent to amphibolites in the NSB and USB around 3.8 Ga;
- 619 2- Reworking of some of these mafic rocks to form the protolith of felsic gneisses (1st felsic
620 generation) between 3.7 and 3.8 Ga;
- 621 3- Deposition of the protolith of the quartz-biotite schists at 3.65 Ga and reworking of existing
622 lithologies (mafic and felsic) to generate the Voizel tonalites (2nd felsic generation; Greer et
623 al., 2020) between 3.45 and 3.65 Ga that have Hf isotope signatures buffered by gneisses of
624 the 1st felsic generation.
- 625 4- Further reworking and/or buffering during the Boizard granite formation 2.7-3.0 Ga (3rd felsic
626 generation) (Greer et al., 2020).

627 However, ¹⁴²Nd anomalies require the presence of a Hadean enriched reservoir, possibly crustal,
628 somewhere in NSB and USB history (O'Neil et al., 2008). Much like what is proposed in Guitreau et al.
629 2013, Boehnke et al., 2018 and Caro et al., 2017, we suggest that the involvement of the Hadean crust
630 dominantly, if not exclusively, took place during the mafic crust formation around 3.8 Ga. The signature
631 of this Hadean crust was transferred to granitoids during a second-stage process corresponding to mafic
632 crust reworking. The range of chemical and isotopic values exhibited by NSB/USB amphibolites can
633 account well for the evolutionary pattern and incorporated Hadean crustal signatures in multiple isotopic
634 systems (Sm-Nd, Lu-Hf; O'Neil et al., 2008, 2017; Guitreau et al., 2013; Caro et al., 2017), as well as in
635 its chemistry (Adam et al., 2012; Boehnke et al., 2018, Caro et al., 2017).

636

637 **5.4 Si+O isotopes in zircon: preservation of primary signatures and implications for source.**

638 The coupled measurement of Si and O isotopes of zircons is a new technique being applied to the
639 rocks of the Inukjuak domain, and we apply these measurements to zircons found in quartz-biotite schist
640 samples, which have the oldest reported ages. To our knowledge, only whole rock O isotope
641 measurements of fuchsitic quartzites from the NSB have been made (Cates et al., 2013). While rocks
642 from the USB preserve direct evidence of hydrothermal activity at 3.8 Ga or later, information can be
643 obtained from zircons about parent melts through stable isotope investigations. Coupled $\delta^{30}\text{Si}$ and $\delta^{18}\text{O}$
644 measurements can reveal the clues about protolith that hosted these zircons; specifically, the lithologies
645 that melted to produce the melt that formed the protolith. This is because both Si and O isotopes
646 fractionate away from mantle-derived samples during low temperature water-rock interactions depending
647 on the nature of the interaction (Robert and Chaussidon, 2006; Abraham et al, 2011; Wenner and Taylor,

648 1973; Muehlenbachs and Clayton, 1976). Trail et al. (2018) have shown that there could have existed
649 reservoirs that were out of equilibrium with the mantle though investigation of zircons from S- I- A-type
650 rocks, as well as detrital grains; in some cases, this inference was only reached because O isotopes and Si
651 measurements were made simultaneously on the same zircons.

652 Given the long existence and the metamorphic history of the zircons measured here, their isotopic
653 ratios might be disturbed but the measurements made here are considered valid because the O diffuses
654 slowly enough during the cooling of the zircon such that the primordial isotopic characteristics of the
655 crystal is preserved. While cooling, Watson and Cherniak (1997) suggests a closure temperature for O
656 isotopes in zircon of ~650 °C for a cooling rate of 100 °C/Ma. This T is very close to the “wet” granite
657 solidus and probably the T of formation for our zircons and thus the $^{18}\text{O}/^{16}\text{O}$ were probably locked-in soon
658 after the zircon crystallized. But our samples have also experienced at least mid-amphibolite grade
659 metamorphism which could have homogenized isotopic ratios. Cherniak and Watson, (2003), based on
660 experiments performed by Watson and Cherniak, (1997), showed that at mid crustal temperatures of ~500
661 °C the core of the zircon will retain the primary $^{18}\text{O}/^{16}\text{O}$ ratio for a maximum of 0.1 Ga for a 20 μm core.
662 This retention time will only increase for larger cores. More recent experimental studies have also posited
663 that zircon cores can preserve $^{18}\text{O}/^{16}\text{O}$ ratios up to much higher metamorphic T of >800 °C (Bindeman et
664 al., 2018). As for Si, the diffusivity of Si in zircon is at least five orders of magnitude slower compared to
665 O under “wet” conditions at ~1000 °C and even slower under colder scenarios (Cherniak, 2008) so Si is
666 likely preserved under most reasonable geologic conditions.

667 The zircons from the quartz-biotite schists (I370 and I374) show that the parent melt may have
668 formed by re-melting of rocks that underwent hydration of primary silicates and authigenic silica
669 precipitation. The datapoints that show a positive deviation in $\delta^{30}\text{Si}$ imply interaction with seawater (Trail
670 et al., 2018) and authigenic silica formation that was re-melted to form the parent melt. Some recent work
671 (André et al., 2019; Deng et al., 2019) shows that Archean granitoids could have evolved from melts that
672 had significant proportions of chert-rich, silicified basalt, the presence of which could be explained by
673 low-temperature water-rock interactions in the Hadean. The re-melting of such rock types suggests some
674 sort of tectonic activity that contributed to the reworking of the rock record or an active silica cycle in the
675 Eoarchean or earlier. Turner et al. (2014) propose a subduction like environment active during the time
676 when the NSB was developing and it might be that the USB could have experienced similar environments
677 but further geochemical analysis is required to prove this hypothesis.

678 In the modern age an active silica cycle is intricately linked to the long-term C cycle through
679 silicate weathering (Walker et al., 1981; Edson et al., 2012). Silicate weathering contributes to the
680 creation of silicate and carbonate-based microfossils. The absence of these organic Si sinks in the

681 Archean might have caused the precipitation of siliceous sediments as well as other inorganic
682 components, which if re-melted at subduction zones, create reservoirs that are chemically distinct from
683 the pristine mantle (Chen et al, 2020). Along with the inorganic siliceous components, the water-rock
684 interactions mentioned above might have caused some of the alkali-rich minerals in the primary rocks to
685 be chemically weathered to clay and mafic minerals to be serpentinized. This is reflected in the datapoints
686 that show a negative and no deviation respectively in $\delta^{30}\text{Si}$ from the mantle values. The important thing to
687 note here is that the $\delta^{30}\text{Si}$ values are, on average, in agreement with mantle-derived zircon, but the $\delta^{18}\text{O}$
688 values are mildly elevated from 'mantle' zircon values. The observations made here indicate two things,
689 firstly, the parent melt might not be pristine mantle material and secondly, the parental melt for the
690 zircons is consistent with having serpentinized lithologies and silicified basalts as significant components.

691 **5.5 Zircon mineral Inclusions**

692 The zircons from the quartz-biotite schists contain several primary inclusions that are isolated
693 from cracks. The inclusions suite has a low abundance of apatite ($13 \pm 19\%$ 2 s.d.) and is dominated by
694 quartz and feldspar indicating that the parent melt was probably felsic according to Bell et al. (2018).
695 There are far more inclusions in contact with cracks than isolated inclusions. The former type of
696 inclusions could imply that the zircons have experienced extensive volumes of hydrothermal fluids, since
697 the quartz/feldspar ratio $\left(\frac{\text{Quartz occurrences}}{\text{Feldspar occurrences}}\right)$ is higher in the on-crack category (1.4) than in the isolated
698 category (0.75). This implies either that the infiltrating fluid was silica-saturated (and thus quartz was
699 precipitated as secondary inclusions), or that primary inclusions were altered by the fluids, with feldspar
700 being more susceptible. The fluid itself seems to be enriched in Fe, Al and Si since the crack-filling
701 inclusions include biotite and other unidentified Fe-Al-Si rich phases. This is in-line with the observation
702 made for some of the thin sections in which some feldspar phases are altered but biotite is relatively
703 unscathed.

704

705 **5.6 Protoliths for the quartz-biotite schists**

706 When trying to establish the nature of the protolith that metamorphosed to form the quartz-biotite
707 schists, comparisons need to be made with similar schists from the NSB. There is a general agreement
708 that the NSB quartz-biotite schists are formed from detrital sediments (Cates and Mojzsis, 2007; Cates et
709 al., 2013; David et al., 2009; Greer et al., 2020) and the USB schists might have a similar protolith given
710 the textural, mineralogical and chemical similarities. The XRF analyses reveal that the parent melts for
711 the rocks, that either supplied sediments for the schists or were the protoliths themselves, were granitic in

712 terms of their chemistry and mineralogy. The Harker diagrams (**Fig. 11**) show that MgO, Fe₂O₃, Al₂O₃
713 and TiO₂ are negatively correlated with SiO₂. This implies that the NSB and the USB schists could have
714 originated from the same magma source.

715 The mafic oxides mentioned have a lower concentration than the WS, which is an Archean mafic
716 shale. The lower values imply that the USB schists had a more felsic progenitor compared to the WS.
717 Furthermore, we see enrichments in Sr, depletions in Rb and especially Ni and Cr compared to the WS
718 (**Fig. 12**). **Fig. 12** also shows that the USB quartz-biotite schists are similar to the NSB schists and the
719 MISB metasediments in terms of their chemistry. Cr/Th is a good indicator of the felsic nature of the melt
720 (Condie and Wronkiewicz, 1990). Condie and Wronkiewicz, 1990, for instance, show that Archean
721 granitoids have low Cr/Th (0.2 - 6.5) and Cr/Zr (0.04 - 0.3) values; the lower end being data from
722 granites while the higher end is from tonalites. The same ratios have much higher values for basalts; 500
723 for the Cr/Th and 5.5 as goes the Cr/Zr. These diagnostic values were further verified by Gao et al. (1999)
724 and Wu et al. (2012) when they measured gneisses from the Yangtse Craton, China. According to Gao et
725 al. (1999), Cr/Th values are between 0.31 - 14.83 where the lower values are for granitic gneisses while
726 the higher ones are for dioritic-trondjhemitic-tonalitic gneisses while the Cr/Zr values for the same
727 lithologies are between 0.01 - 1.08. Wu et al. (2012) only measured tonalitic gneisses and their Cr/Th
728 values are ~1.99 while the Cr/Zr values are ~0.1. These values are within range of the ones proposed by
729 Condie and Wronkiewicz (1990). If the USB schists are derived from eroded granitoids, they should
730 resemble them in terms of the two parameters just discussed. The USB schists analyzed here have Cr/Th
731 values of 7.05 (I370) and 11.4 (I374) and Cr/Zr values of 0.22 (I370) and 0.39 (I374) which are close to
732 the tonalitic values. Interestingly, the δ¹⁸O values for the quartz-biotite schist zircons analyzed here is also
733 similar to that of other Archean TTG zircons (Valley et al., 2005): ~5.53‰ (Barberton, South Africa) and
734 5.65 - 6.69‰ (Zircons from over 100,000 km² in the Superior Province).

735 For the NSB schists, there are large clasts of quartz and feldspar that might lead one to think that
736 it is a metaconglomerate. The USB schists we have collected so far have no such clasts, but the zircons
737 extracted from them are sub-rounded and not as euhedral (**Fig. S8**) as the gneissic zircons implying that
738 they may have experienced erosion. MgO, Fe₂O₃, TiO₂ and MnO were plotted against Al₂O₃ for both the
739 USB and the NSB schists. Except for TiO₂, which shows a strong correlation, all the other oxides show a
740 weak to moderate correlation. This implies that the oxides are probably associated with clays and micas,
741 the end-products of weathering and erosion, and not mafic minerals.

742 Finally, since most of the evidence indicate a sedimentary protolith, the rocks were compared on
743 the classification diagram for sands and shales proposed by Herron (1988) along with NSB schists (**Fig.**
744 **13**). I374 plots in a location like most of the NSB schists while I370 plots in the “wacke” field, which

745 might imply that I370 might be a bit more mature than the other schists and its protolith probably had a
746 greater amount of feldspar grains or clays than the former. The major objective of using the classification
747 diagram in Fig. 13 is to show how the major element measurements of similar lithologies from the NSB
748 and the USB compare to each other and make careful and conservative inferences about the protoliths of
749 the lithologies from the two belts. The classification diagram itself plays a secondary role and protolith
750 inferences have to be a bit speculative since hydrothermal alteration might severely affect mobile oxides
751 like K_2O but, most likely, both these belts could have undergone similar hydrothermal environments
752 given their proximity which makes their intercomparison a valid exercise.

753 **6. Conclusions**

754 Integrated studies work to deconvolute multiple alterations and tectonic events that Eoarchean
755 crustal enclaves have experienced, to finally distill the primary protolith of these enclaves. The main
756 emphasis of the present study is the Ukaliq Supracrustal Belt in northern Québec, a large metamorphosed
757 volcano-sedimentary enclave that may be structurally (and, petrogenetically) related to the neighboring
758 Nuvvuagittuq Supracrustal Belt. Few studies have explored these rocks (Caro et al., 2017; Saji et al.,
759 2018; Greer et al., 2020). Like the NSB, the USB is structurally complex and comprises medium to high
760 grade metamorphosed rocks that have been hydrothermally altered. Physical and geochemical tests were
761 done on these rocks to qualify the lithology of this belt, constrain out the ages of these rocks, understand
762 whether the zircons analyzed for U-Pb geochronology were igneous or metamorphic, try and uncover
763 what the protoliths of these rocks were and to establish similarities with the NSB. The major lithologies
764 studied were schists, amphibolites, gneisses from the USB. We also followed up with additional thin
765 section observation of a quartzite of controversial origin from the NSB. Thin section observations from all
766 these rock types reveal that these rocks were not metasomatized uniformly and that even samples from the
767 same rock unit display different alteration signatures. They also show that the USB rocks have
768 experienced crystal-plastic deformation like the NSB.

769 Before analyzing the zircons extracted from these rock units, CL and BSE images were taken
770 which show that all the zircons have igneous cores and some of them have rims of variable thicknesses.
771 Subsequent care was taken to avoid these rims which will not provide information about the parent melts.
772 Direct measurements have provided an Eoarchean age for the zircons extracted from the schists and the
773 gneisses and show a tight cluster in their Th/U ratios suggesting that the measurements do not include
774 metasomatized sections of the zircons. The ages of the gneisses and the quartz biotite schists (I370 and
775 I374) are comparable to the NSB units of the same lithology and even to the Isua Supracrustal Belt. We
776 propose that the gneisses have an emplacement age of 3.68 Ga and that the quartz-biotite schists have a
777 maximum depositional age of 3.65 Ga. The depositional age of the USB schists is similar to the oldest

778 ages of the zircons from the NSB schists. The melts from which they crystallized show that they possibly
779 formed by the re-melting of either clay minerals, authigenic silica or serpentinized basalt or a mixture of
780 some proportion containing some or all these materials although the last lithology seems to be most
781 probable and a significant constituent. This implies that crustal processes such as chemical weathering
782 that form clays and authigenesis as well as those that lead to re-melting of such lithologies were active
783 during the Eoarchean/Hadean.

784 The probable nature of the protolith was further explored through major and trace element
785 analysis of the quartz-biotite schists from the USB while being compared to similar lithologies from the
786 NSB. The data suggests that the schists were metamorphosed products of detrital sediments. Zircons from
787 the USB were also analyzed for their mineral inclusions and the Lu-Hf isotopic ratios. The mineral
788 inclusion occurrences suggest that the parent melts were probably felsic. The Lu-Hf systematics also
789 present a lot of chemical similarities between the USB and the NSB host rocks as well the constituent
790 zircons implying a chemically similar source or the same source for both belts. Finally, based on the $\epsilon_{\text{Hf}(t)}$
791 vs age plot for the USB and μNd^{142} anomalies reported for the USB and the NSB, the data are consistent
792 with a Hadean reservoir that imparted its isotopic characteristic upon the mafic parent melt that was
793 extracted at ~ 3.8 Ga. A significant observation we made is the similarity in U-Pb ages between the USB,
794 NSB and the Isua Supracrustal Belt as well as the chemical similarity (Lu-Hf isotopes and Trace
795 elements) between the quart-biotite schists of the USB and the NSB. These similarities might imply that
796 these supracrustal belts could have been a larger single terrane or that these belts experienced similar
797 geologic processes. This possibility was previously explored by Cates and Mojzsis (2007). Our study is
798 an initial foray into establishing the geochemistry and thus the evolution of the USB which will be a step
799 in the direction of studying the entire Inukjuak domain that might lend significant insight into the Hadean
800 environment.

801

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818

819 **FIGURE CAPTIONS**

820 **Figure 1:** Map showing the study area. The top left corner denotes the location of the town of Inukjuak.
821 The red star designates all the sampling locations. The map inset (left) shows the general location of the
822 field area which is on the eastern bank of the Hudson Bay. The inset maps show the location of the two
823 supracrustal belts. The dotted square is the study area of Caro et al., (2017) while the solid square is the
824 same for this study (Map modified from Caro et al., 2017). The vertical panel on the extreme right shows
825 the samples collected. The GPS co-ordinates of all locations are listed in Table 1. Scale bar on the left
826 inset map = 300 km and on the right inset map = 1 km.

827 **Figure. 2: a.** BSE (top four) and corresponding **b.** CL (bottom four) images of zircons from some of the
828 gneisses mentioned in this study. Clockwise from the top left, the samples are I349, I349, I355 and I365
829 (Scale bars for both sets of images are 100µm). These zircons are representative of the gneissic zircons.
830 Note the oscillatory zoning and the intact cores in all the zircons. The zircons contain cracks which were
831 avoided during U-Pb analysis as best as possible. The circles (not to scale) show Laser ablation locations
832 for U-Pb chronology and the corresponding ages are in Ga.

833 **Figure 3:** Zircons extracted from quartz-biotite schists of the USB. The images show BSE (Right panel)
834 and corresponding CL (Left panel) from I370 (top set) and I374 (bottom set). These grains are
835 representative of zircons from the schist. The grains show oscillatory zoning in the core; implying an
836 igneous origin at least for the core. The grains also show prominent metamorphic rims that do not show
837 any zoning and is uniform in its color. The circles show sputtering locations and the ages are ion
838 microprobe Pb²⁰⁷/Pb²⁰⁶ ages and Si and O stable isotope data. The scale bars in the left panel are 40 µm.

839 **Figure 4: (a-d):** Optical microscope pictures of the various rock types mentioned in this study. **(a).**
840 Quartz-biotite schist where the biotite is altered to chlorite. This is a sign of hydrothermal alteration in the
841 USB much like the NSB. **(b).** Cumingtonite schist. Note the alterations on the basal sections. **(c).** Biotite
842 bearing gneiss. **(d).** Fuchsite-bearing quartzite from the NSB. The fuchsite is the chaotic fibrolitic phase
843 that appear as alterations. All images show evidence for hydrothermal alterations. **(e-f): (e).** Hornblende
844 schist in the USB. This is one of the two kinds of amphibolites discussed in this study. The hornblende
845 has been hydrothermally altered to chlorite crystallizing Fe/Ti-phases as a byproduct. **(f).** Cumingtonite
846 schist in the USB. The other kind of amphibolite which is found in both the USB and the NSB. The rock
847 shows elongated and ribboned quartz grains implying that this sequence has experienced crystal-plastic
848 deformation and subsequent relaxation as evidenced by the sub-grain formation. **(g-i):** The two kinds of
849 gneissic samples discussed in the work; Granitic gneiss **(g)** which is alkali-feldspar rich and tonalitic
850 gneiss **(h)** which has significant plagioclase. Both gneisses are mica bearing and is commensurate with
851 the general observation that the region has experienced amphibolite facies metamorphism. **(i)** the tonalitic
852 gneiss also displays myrmekitic quartz (tubular quartz observed at the boundary between plagioclase and
853 alkali feldspar). This kind of texture is observed as alkali feldspar is hydrothermally altered to plagioclase
854 and the excess silica forms the observed tubular texture. There are also quartz grains that show ribboning
855 and sub-grain formation which implies deformation and subsequent relaxation. Scale bar = 500 μm .

856 **Figure 5: Clockwise from top left: I346, I349, I365, I355** Age histograms and corresponding Th/U of
857 the zircons (All zircons analyzed and reported in Table **ST1b**) from some of the gneisses. The histograms
858 all show a population at 3.4 Ga with some grains older at around 3.5 Ga. I349 has one grain at 3.71 ± 0.05
859 Ga but with a concordance of 109%. The large spread in Th/U values show that either there is a
860 metamorphic component to the measurements or that the parent melt was not homogeneous. Most of the
861 Th/U values of the youngest zircon analyses around 2.7 Ga are the lowest; this might be because of a
862 metamorphic event around this time.

863 **Figure 6: Clockwise from top left: I370, I374, IN05037, IN05020.** Comparison of Age distributions of
864 quartz-biotite schists zircons (Zircons reported in **Table ST1b**) from this study (two top figures;
865 measured using an LA-ICP-MS) and from the same lithology in the NSB (two bottom figures; measured
866 using a SIMS) (Cates and Mojzsis, 2007). The histograms show similar populations at 3.7 Ga and the
867 oldest age of all the zircons analyzed is 3.79 Ga for the USB rock units. Th/U ratios in the USB are
868 similar compared to the NSB Th/U values but have a smaller spread which may be because younger
869 zircons were not analyzed.

870 **Figure 7:** Average ages and Th/U of the Core and Rim zones of zircons from gneiss samples of the
871 USB. The ages of zones are similar for all samples except for I365. The Th/U ratio for I355 reduces from

872 the core to the rim indicating that the rim has a metamorphic component. Such trends are also observed
873 for I346 although it is not as pronounced. The diamonds without an error bar have errors lower than the
874 size of the diamond.

875 **Figure 8:** Si and O isotopic values of zircons from quartz-biotite schists of the USB (error bars are 2 s.e.).
876 The datapoints in solid grey and white with grey borders are rejects because the sputtering pits hit cracks
877 in the grain. The zircons show $\delta^{18}\text{O}$ values distinct from the mantle while the $\delta^{30}\text{Si}$ seems to be uniformly
878 distributed about the mantle values.

879 **Figure 9:** A comparison of the $\delta^{30}\text{Si}$ against the age of zircons from the USB and the Jack Hills (Trail et
880 al., 2018). In the case of multiple analyses of a sample, the average of all replicates has been plotted for
881 the Jack Hills zircons. The USB zircons have the same average isotopic values as the Jack Hills zircons.
882 One of the zircons having low $\delta^{30}\text{Si}$ has a $\delta^{18}\text{O}$ value quite close to mantle values implying that the zircon
883 might be out of equilibrium with the mantle. Such evidences are also seen among the Jack Hills sample
884 set.

885 **Figure 10:** Hf isotope signatures ($^{176}\text{Hf}/^{177}\text{Hf}$, ϵ_{Hf}) as a function of ages for zircons analyzed in this study
886 (a-b) as well as zircons, granitoids and amphibolites from previous studies (c-d). CHUR values are from
887 Iizuka et al., 2015). Grey dashed lines represent the evolution of geological reservoirs with $^{176}\text{Lu}/^{177}\text{Hf}$ of
888 = 0.025 (basalt/gabbro), 0.006-0.002 (TTG; e.g., Guitreau et al., 2012), 0.002-0.0004 (zircon; this study).
889 The good match between the datapoints and evolution lines for zircons (0.0004-0.002) in (a) and (b)
890 suggests possible disturbances of the U-Pb isotope system resulting in younger ages than that of actual
891 crystallization, no change in $^{176}\text{Hf}/^{177}\text{Hf}$, and spuriously low ϵ_{Hf} values. This evolution controlled by a
892 very low $^{176}\text{Lu}/^{177}\text{Hf}$ is, however, not so different from that exhibited by magmatic zircon data as
893 illustrated in (c) and (d). This general evolution line along a $^{176}\text{Lu}/^{177}\text{Hf}$ consistent with that of TTGs (e.g.,
894 Guitreau et al., 2012) is, hence, a true feature of NSB and USB, and echoes what can be seen in the
895 Pietersburg block (South Africa; Laurent and Zeh, 2015).

896 **Figure 11:** The left and the center panel are Harker diagrams where the NSB and USB schists are shown
897 together along with the Witwatersrand Shales (green triangles; Wronkiewicz and Condie, 1987),
898 metasediments from the Isua supracrustal belt (yellow stars; Bolhar et al., 2005) and Archean
899 platforms/shields (blue crosses; Rudnick and Fountain, 1995) . The mafic oxides, Al_2O_3 , and P_2O_5 show
900 negative correlations implying that the schists from the two belts formed by fractional crystallization from
901 the same source. The K_2O values might be incorrect since the rocks are heavily altered. The absence of
902 any trend in the alkali oxide plots however suggest that the rocks might be from different sources. The
903 third panel shows that the mafic oxides are moderately to strongly correlated with Al_2O_3 suggesting that

904 the oxides might be associated with clays or micas rather than mafic minerals. (All concentration values
905 are in percentages)

906 **Figure 12:** Trace element data for the USB Schists normalized to the NASC (North American Shale
907 Composite). The green triangles are the average value from various shales from Witswatersrand, the
908 yellow stars are the Isua metasediments and the blue crosses are Archean platforms/shields values. Note
909 the similarities between the USB schists and the Archean lithologies. The USB schists are prominently
910 depleted in Ni and Cr and enriched in Sr implying that the USB schist protoliths might be more felsic than
911 the mafic shales or that the USB has been metasomatized which will cause a depletion in Ni and Cr. The
912 other trace elements show broad similarities with the Archean shales implying that the USB schists could
913 be metasediments. The grey band is the maximum and the minimum concentrations for each element
914 across all the NSB quartz biotite schists analyzed by Cates et al. (2013).

915 **Figure 13:** The NSB (Cates et al., 2013 and Darling et al., 2013) and USB (This study) schists plotted on
916 the classification diagram for sands and shales from Herron (1988). The USB and NSB schists seem to
917 plot in similar locations. However, The K₂O values might be lower than from before the schists were
918 hydrothermally altered which would put the unaltered rocks in a much more immature field. Isua
919 metasediments (Bolhar et al., 2005) have also been shown for comparison.

920

921 **TABLE CAPTIONS**

922 **Table 1:** Details of samples collected for this study. GPS coordinates based on the WGS84 geodetic
923 reference coordinate system.

924 **Table 2:** ²⁰⁷Pb/²⁰⁶Pb ages reported in the manuscript. The ages considered here are derived for zircons that
925 are between 90 and 100% concordant. The errors are 2 standard deviation. For I373, the U-Pb date has
926 been reported from analyses made in the intermediate region between the core and the rim.

927

928 **SUPPLEMENTARY INFORMATION**

929 **SUPPLEMENTARY FIGURE CAPTIONS**

930 **Figure S1: a.** Continuous schistosity defined by the mafic and felsic minerals of the rock. Note the
931 extensive alteration of the biotite to chlorite. This is probably caused by a K and Fe deficient fluids. **b, c.**
932 Other biotite schists showing minimal alteration of the biotite (**b**) and moderate alteration (**c**). The dark
933 spots on the biotite crystals are radiation haloes caused by radioactive inclusions. The study area is

934 characterized by samples from the same rock unit showing signs of having been affected differently by
935 hydrothermal fluids. Scale bar = 500 μm . **d.** Quartz grains from gneisses showing grain boundary
936 migration and sub grain formation. Note the triple junction between quartz grains implying growth or an
937 alteration of the grain boundary during a metamorphic event. Scale bar = 200 μm . **e.** Possible matrix
938 inclusion in quartz. These finer grains might have been captured during growth of the quartz grains during
939 a metamorphic event. The quartz grains also show a triangular grain boundary which is another sign that
940 there was growth during a thermal event. Scale bar = 100 μm . **f.** Hornblende schist from the USB. This
941 section shows quartz porphyroblasts (not surrounded by fine-grained matrix) with biotite crystals flowing
942 around the grains (encircled). This observation is unique to this thin section. Such textures might mean
943 that these quartz grains grew during a deformation event rather than being a primordial igneous grain.
944 Scale bar = 500 μm . (Images in Plane polarized light using an Optical Microscope). **g-i.** Field
945 photographs of some of the gneisses sampled. **j-l.** The quartz-biotite schists discussed in the text. The red
946 circle in photo **l** shows the coin we used for scale. Note the alterations and weathering on all the rocks.
947 Abbreviations: QBS – Quartz-biotite Schists, Gn – Gneiss. Hammer and coin used for scale.

948 **Figure S2:** **a.** Relatively clean plagioclase and altered biotite: probably K-deficient fluid; since biotite
949 altered to chlorite involves K^+ being removed from the biotite crystal structure and the Fe being re-
950 arranged to form chlorite grains. Al can also be added during chloritization. **b.** Altered plagioclase
951 (smudged colorless phase) with relatively unaltered biotite: K and Fe-saturated fluid; since plagioclase
952 being altered to K-rich clay is an indicator of a K-rich fluid. Also, since it is observed that biotite is not as
953 altered as the section on the left, the same conclusion about fluid chemistry may be reached. Scale bars =
954 500 μm . (Plane polarized light). **c.** Plagioclase surrounding relict K-feldspar grain in the purported
955 Tonalitic gneiss. This rock may have been granitic to begin with and the alkali feldspar might have been
956 altered to plagioclase. Or the K-feldspar may be secondary and formed after the plagioclase was affected
957 by a K-rich fluid from which the K-feldspar crystallized. **d.** Minor plagioclase occurrences in the fuchsite
958 bearing quartzite. Scale bar = 500 μm . (Cross-Polarized light (XPL)).

959 **Figure S3:** The zircon grains from **(a)** orthogneisses and **(b)** quartz-biotite schists that have been
960 considered when presenting ages for the units. Note that the I373 grains are analyses from an intermediate
961 region of the zircon. The green triangles are age markers. All ratios reported here were measured using an
962 LA-ICP-MS.

963 **Figure S4:** Comparison of $\text{Pb}^{207}/\text{Pb}^{206}$ average ages measured on zircons from the quartz-biotite schist of
964 the USB. The SIMS measurements are more precise than the LA-ICP-MS measurements in this study. All
965 measurements used to calculate this average age has % concordance between 90 and 100. The data has
966 been listed in Table 2.

967 **Figure S5:** Age distributions and Th/U ratios of unconsolidated sediments collected from stream beds.
968 All distributions show a strong peak at ~2.7 Ga which is mostly because the surrounding igneous suite
969 dominates the regional lithology. The younger zircons have a large range of Th/U ratios implying that
970 igneous activity was coeval with a metamorphic event.

971 **Figure S6:** Secondary electron image of a chromite grain surrounded by fuchsite from the NSB. The
972 texture of the grain suggests dissolution of the spinel to form the Cr-rich mica. The mica might therefore
973 be a secondary deposit. (Image taken using a Zeiss Auriga SEM/FIB at the University of Rochester).

974 **Figure S7:** CL image of zircon grain I370_3_7 whose laser ablation age was selected as the maximum
975 depositional age for the quartz-biotite-schists. The prominent oscillatory zoning shows that the core of the
976 zircon is magmatic in origin.

977 **Figure S8:** BSE (Top) and CL (Bottom) images of representative zircons from the USB Quartz-biotite
978 schist. These zircons are sub-rounded and not as euhedral as the gneissic zircons implying that these
979 zircons underwent transport. Scale bars = 10 μm .

980

981 **SUPPLEMENTARY TABLE CAPTIONS**

982 **Table ST1: a.** Instrumental Parameters for U-Pb geochronology using an LA-ICP-MS. **b.** Results of the
983 U-Pb geochronology on detrital zircon grains using an LA-ICP-MS. **c.** Results of the U-Pb
984 geochronology on detrital zircon grains using a SIMS.

985 **Table ST2:** Results of the U-Pb geochronology on detrital zircon grains from the unconsolidated
986 sediments using an LA-ICP-MS.

987 **Table ST3:** Parameters for the Lu-Hf measurements.

988 **Table ST4:** Results of the Si and O isotope measurements on the detrital zircons.

989 **Table ST5:** Results of the mineral inclusion study of the detrital zircons.

990 **Table ST6:** Raw data for the Si and O measurements

991 **Table ST7:** Results for the Lu-Hf measurements.

992 **Table ST8:** Major and trace element values of the whole rock measurements performed on the quartz-
993 biotite schists.

994

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1318

Figure 1

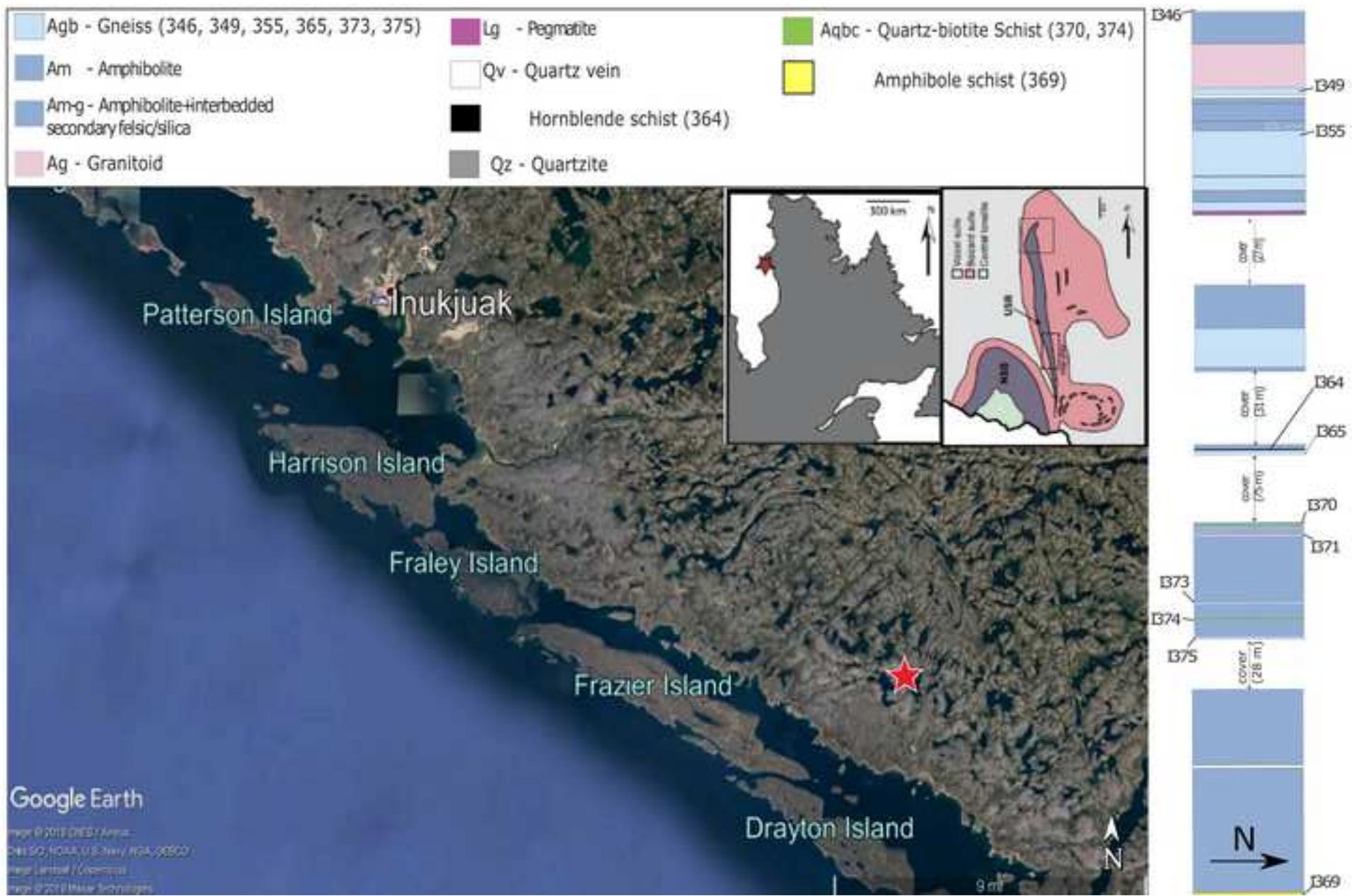


Figure 2

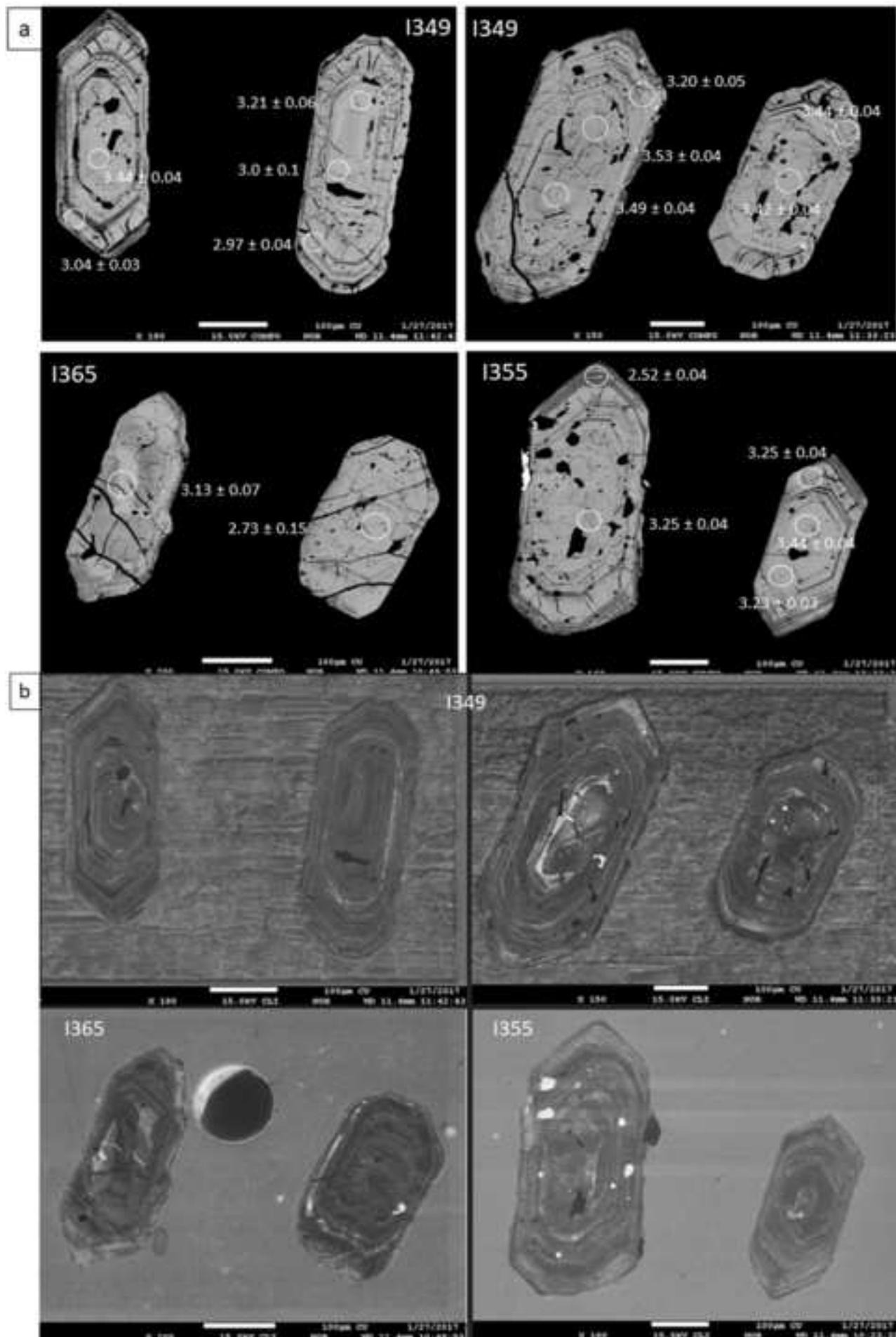


Figure 3

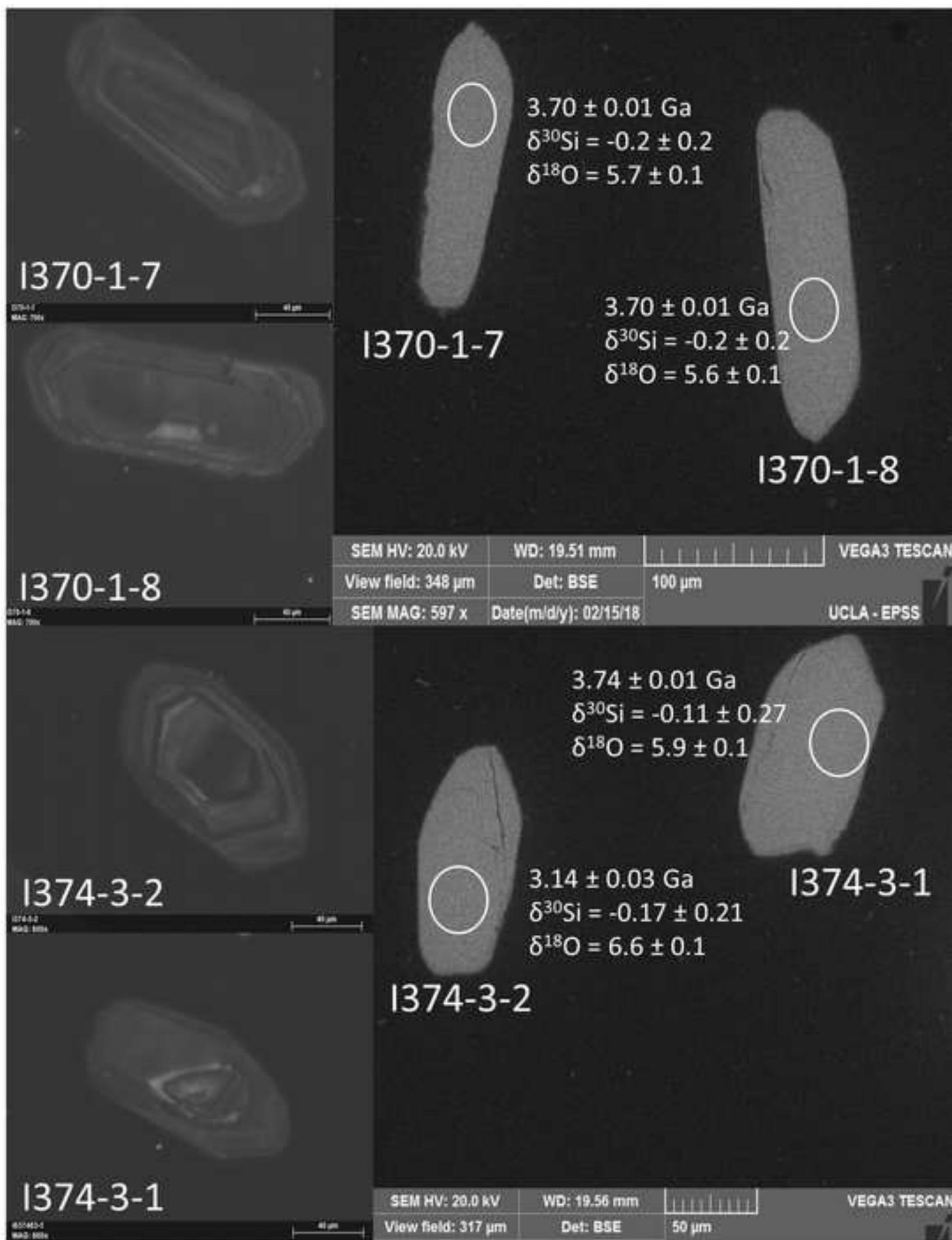


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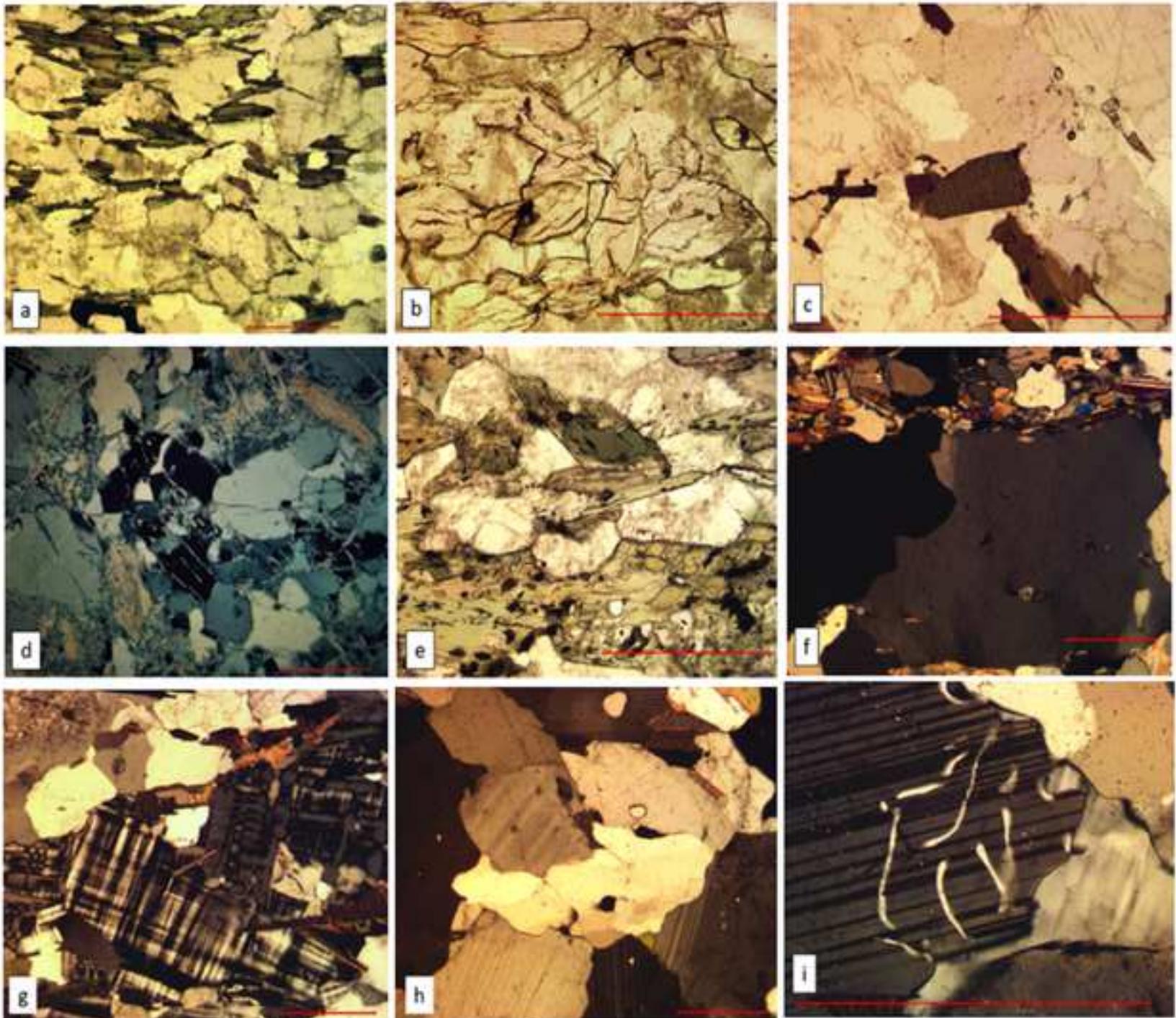


Figure 5

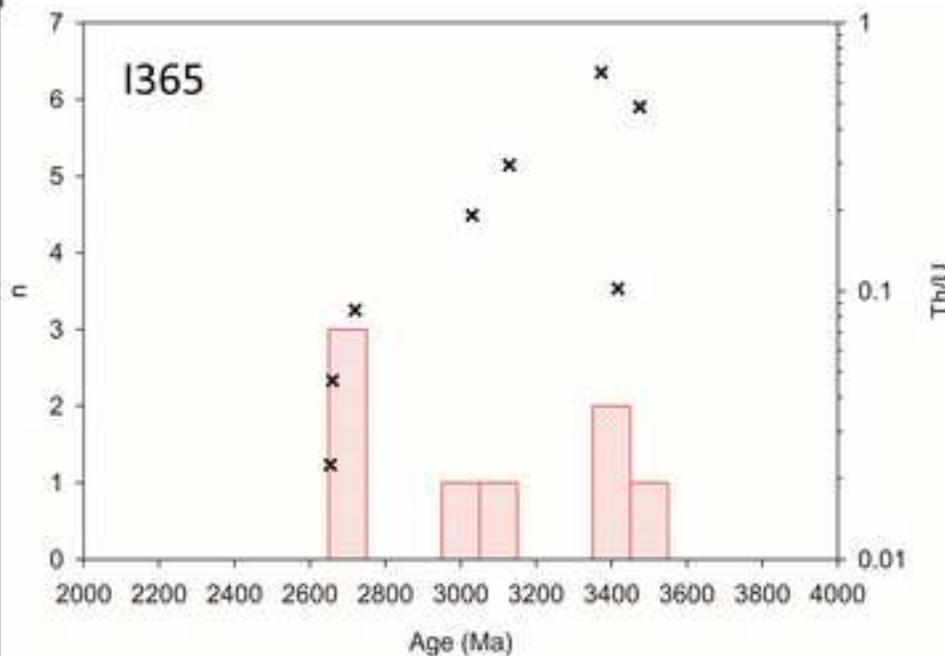
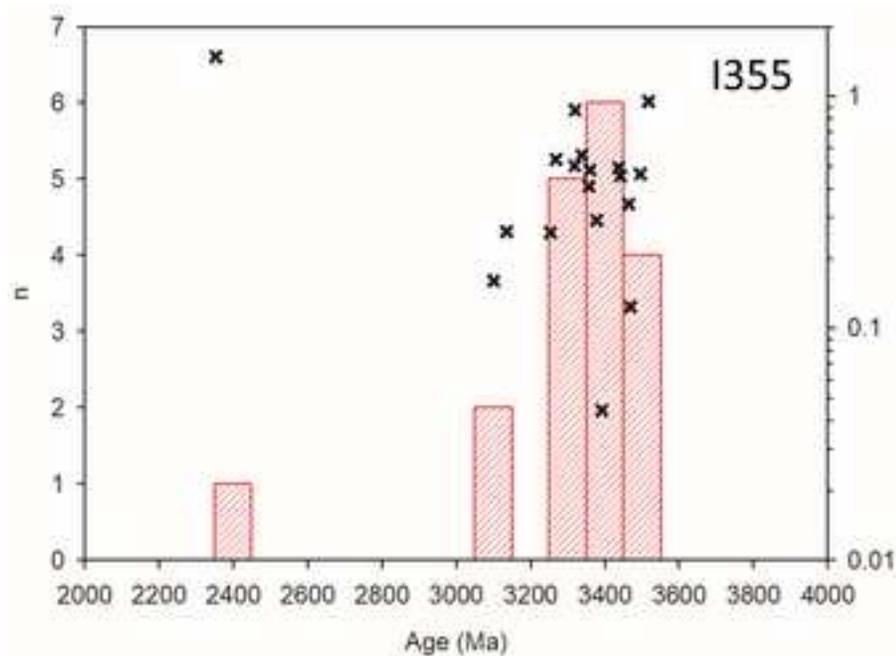
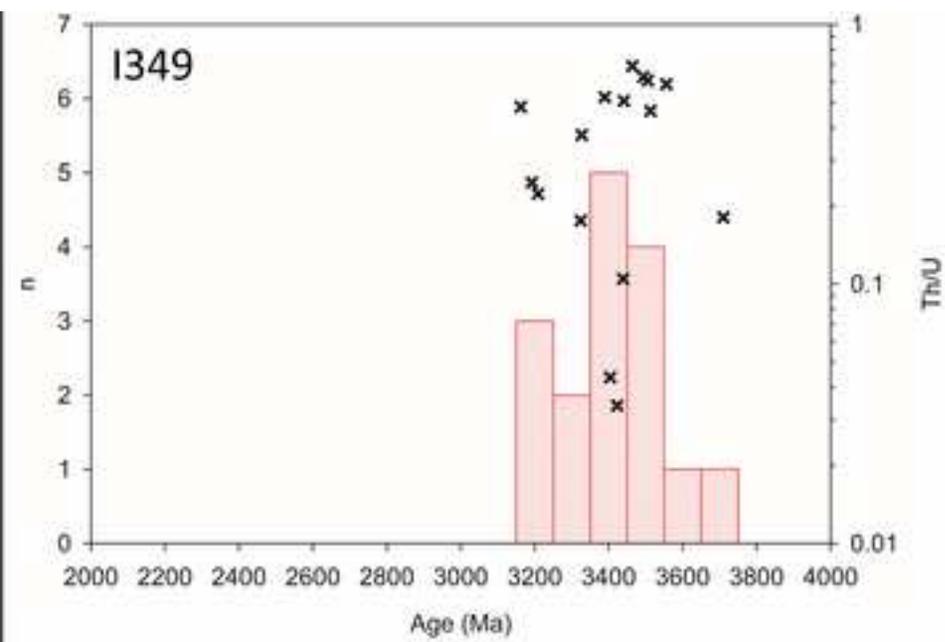
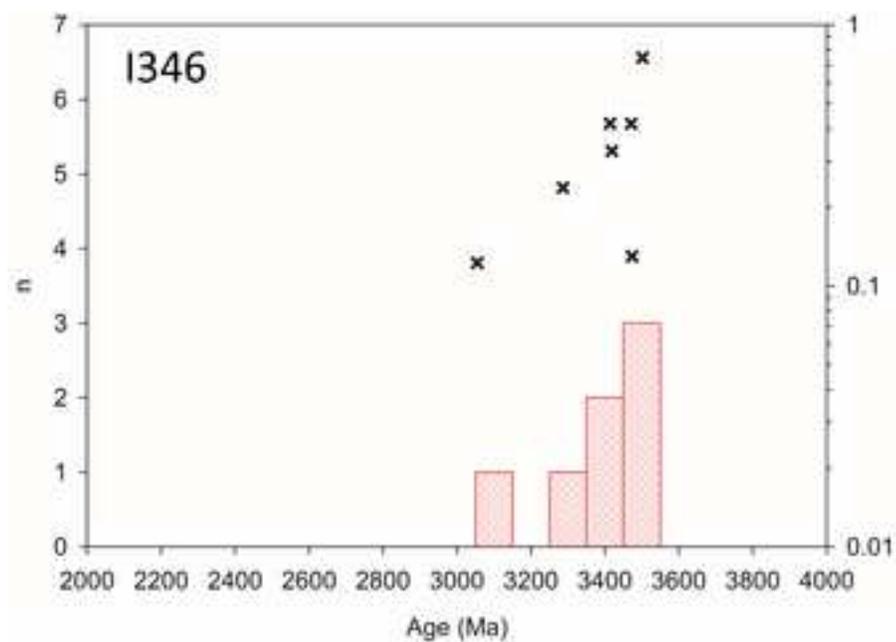


Figure 6

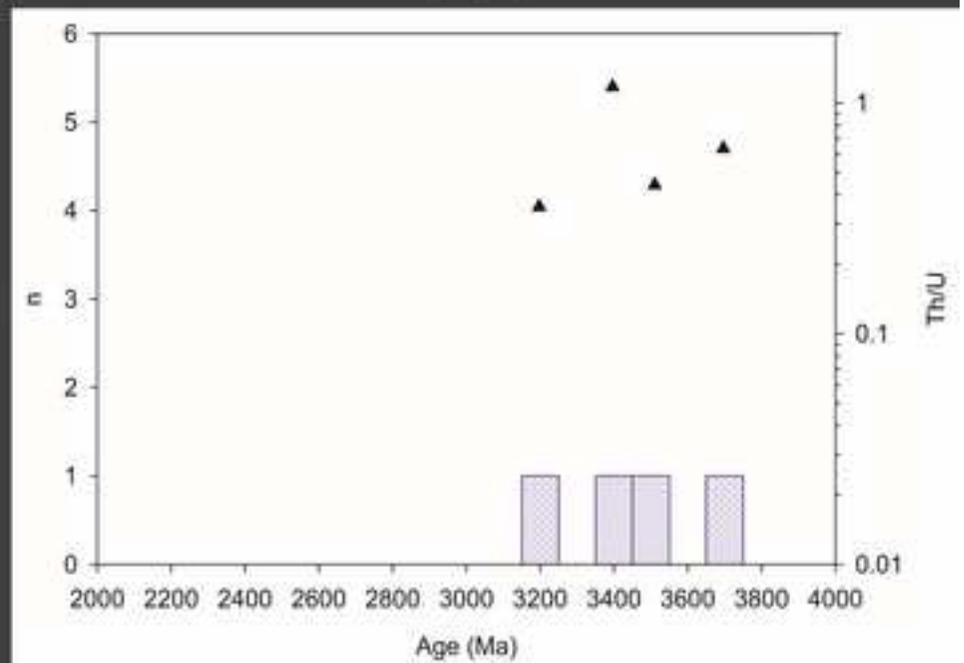
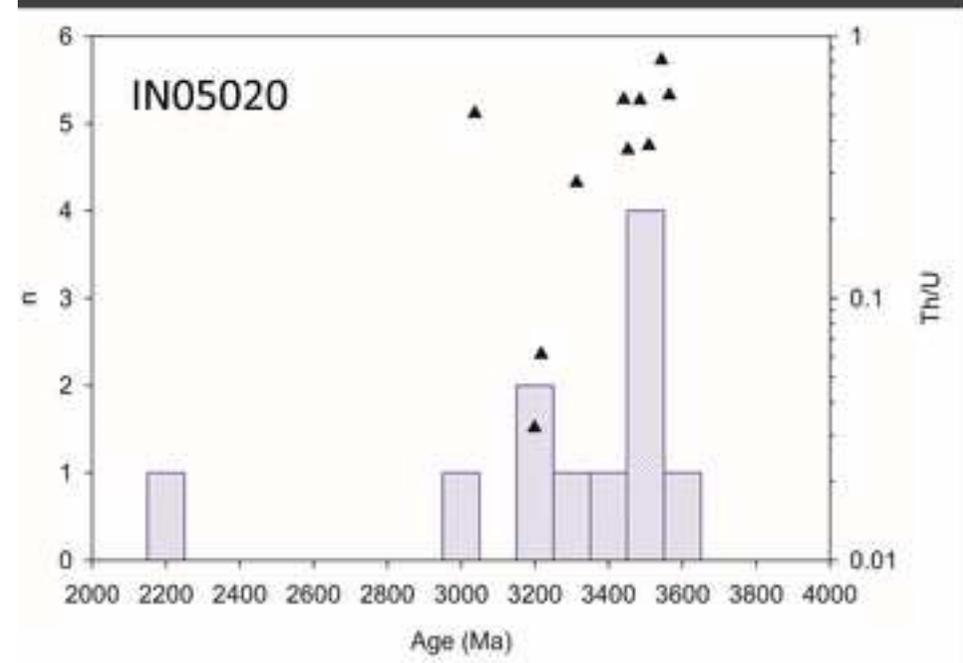
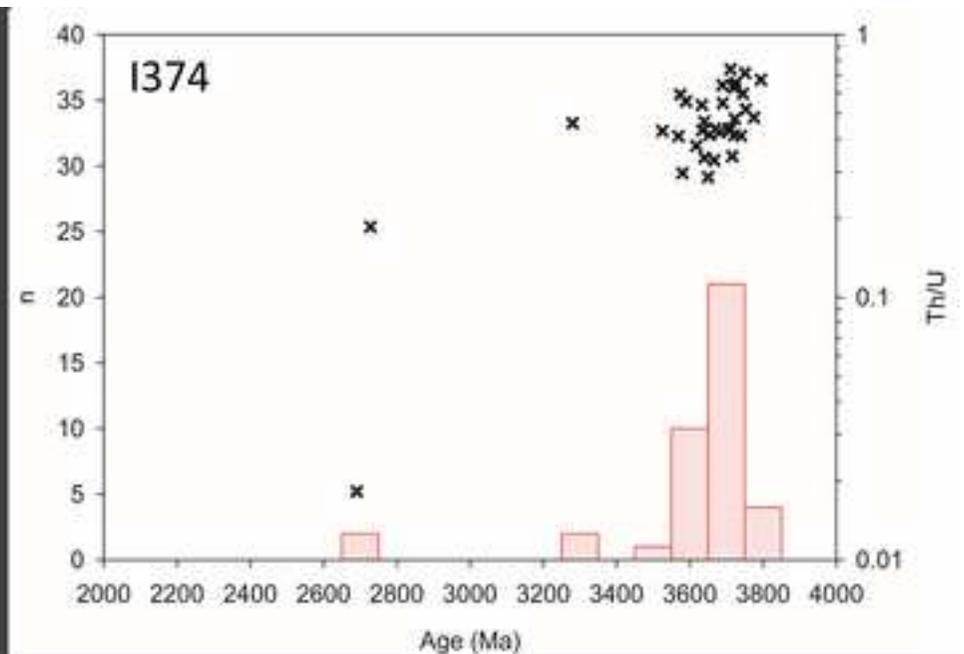
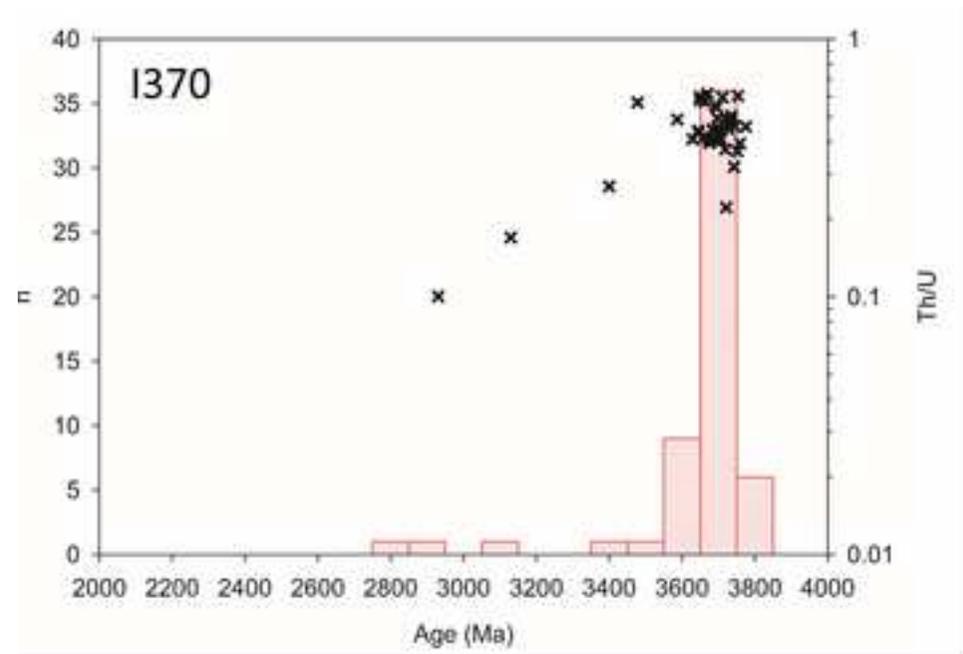


Figure 7

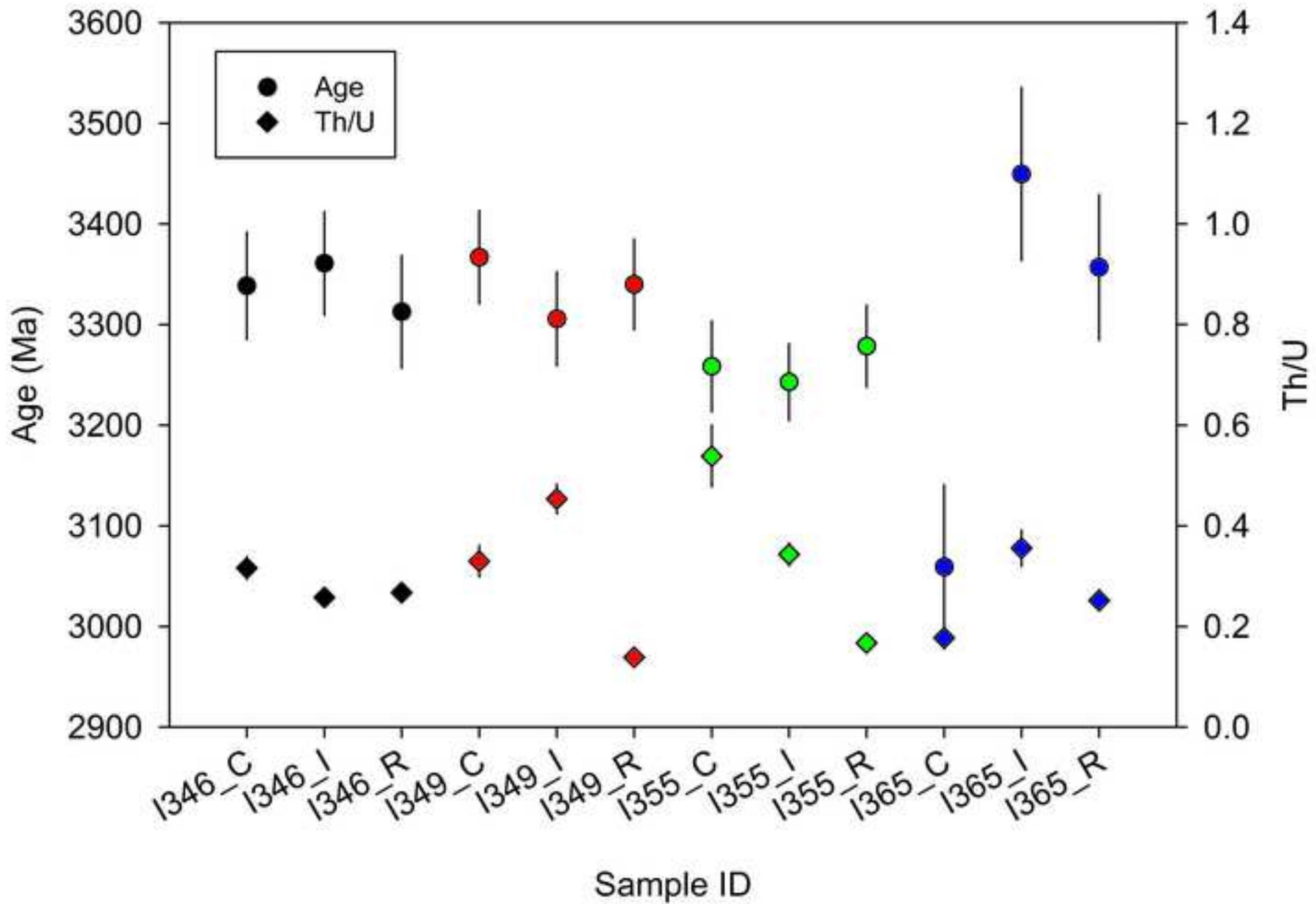


Figure 9

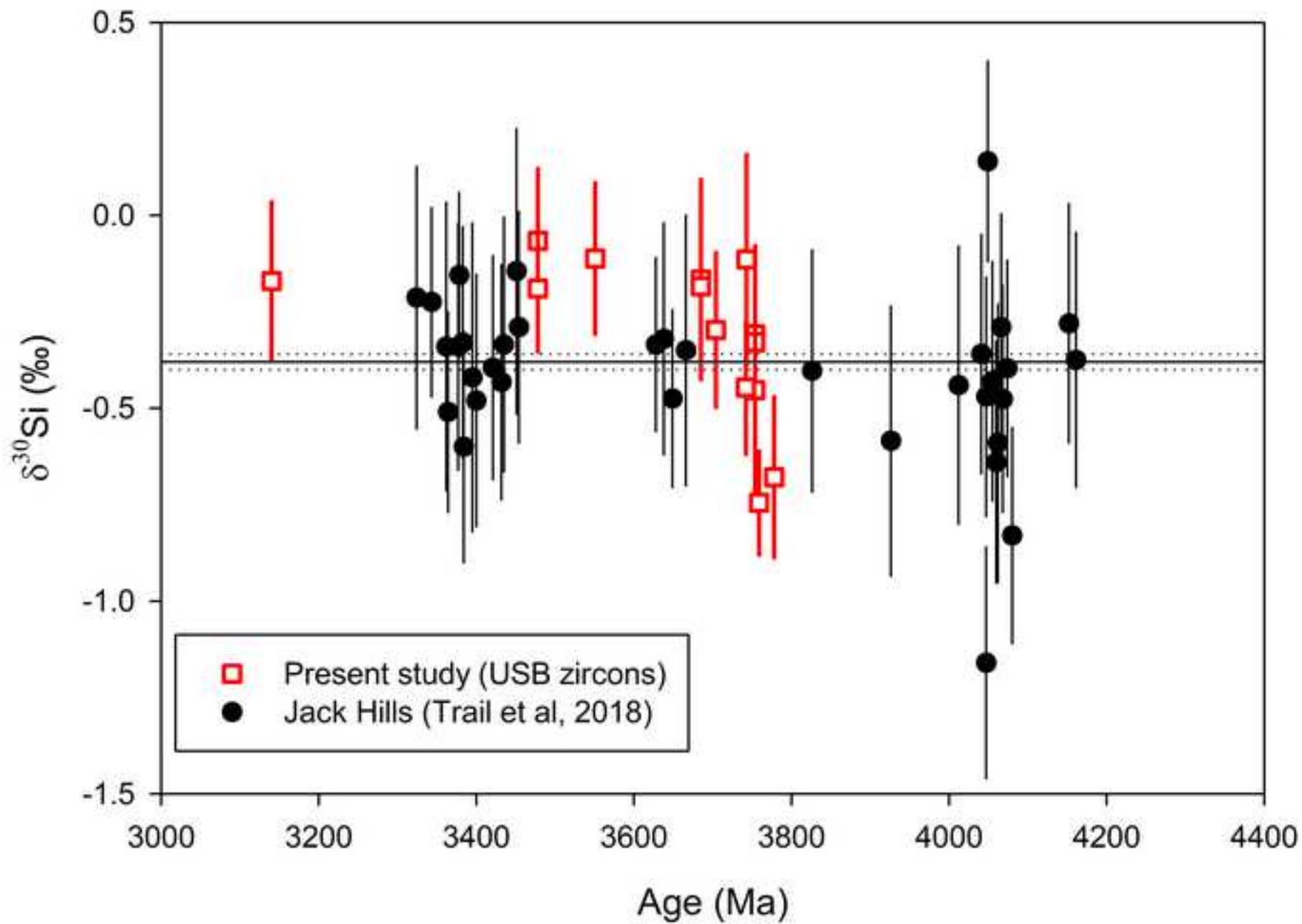


Figure 10

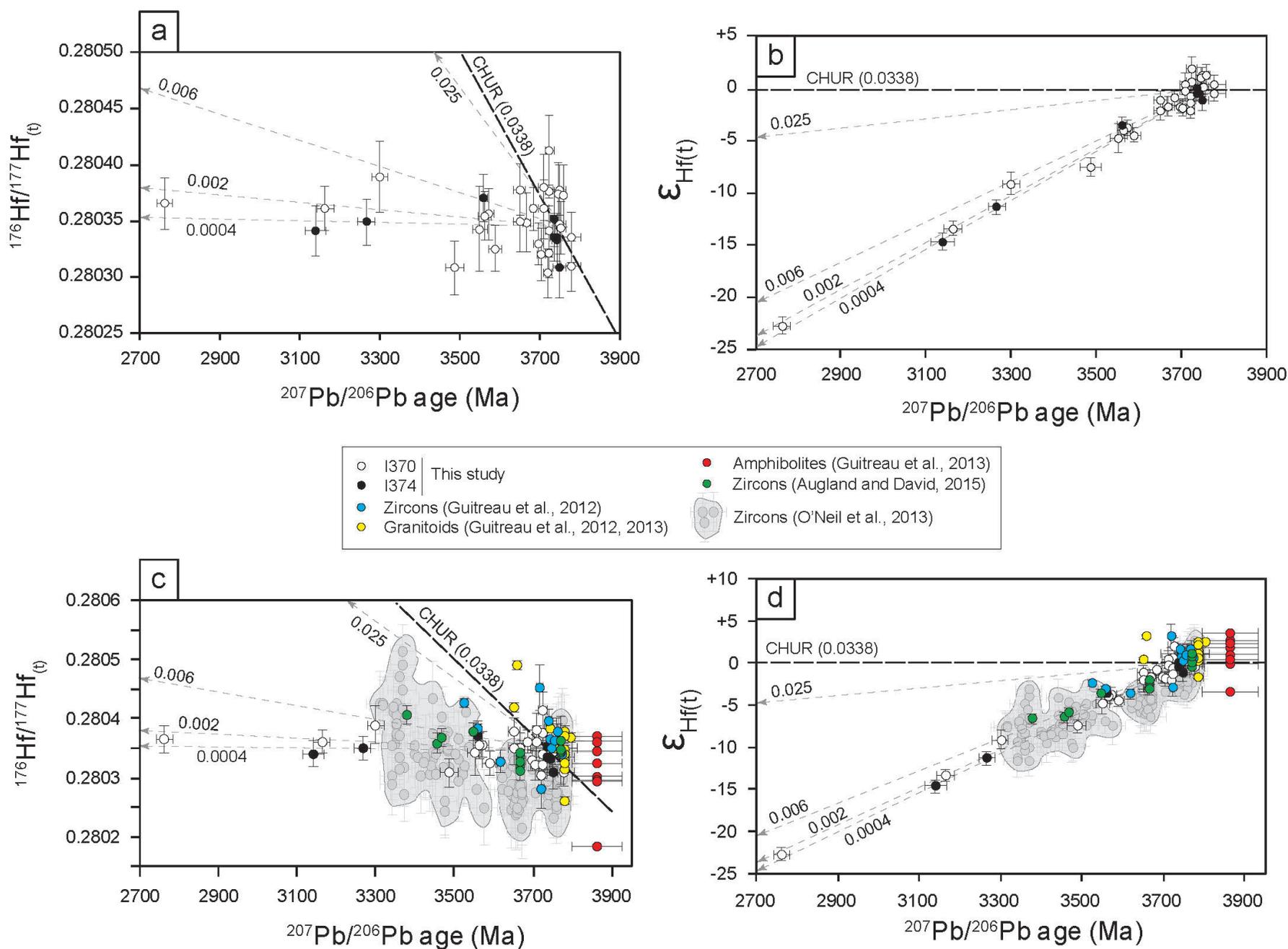
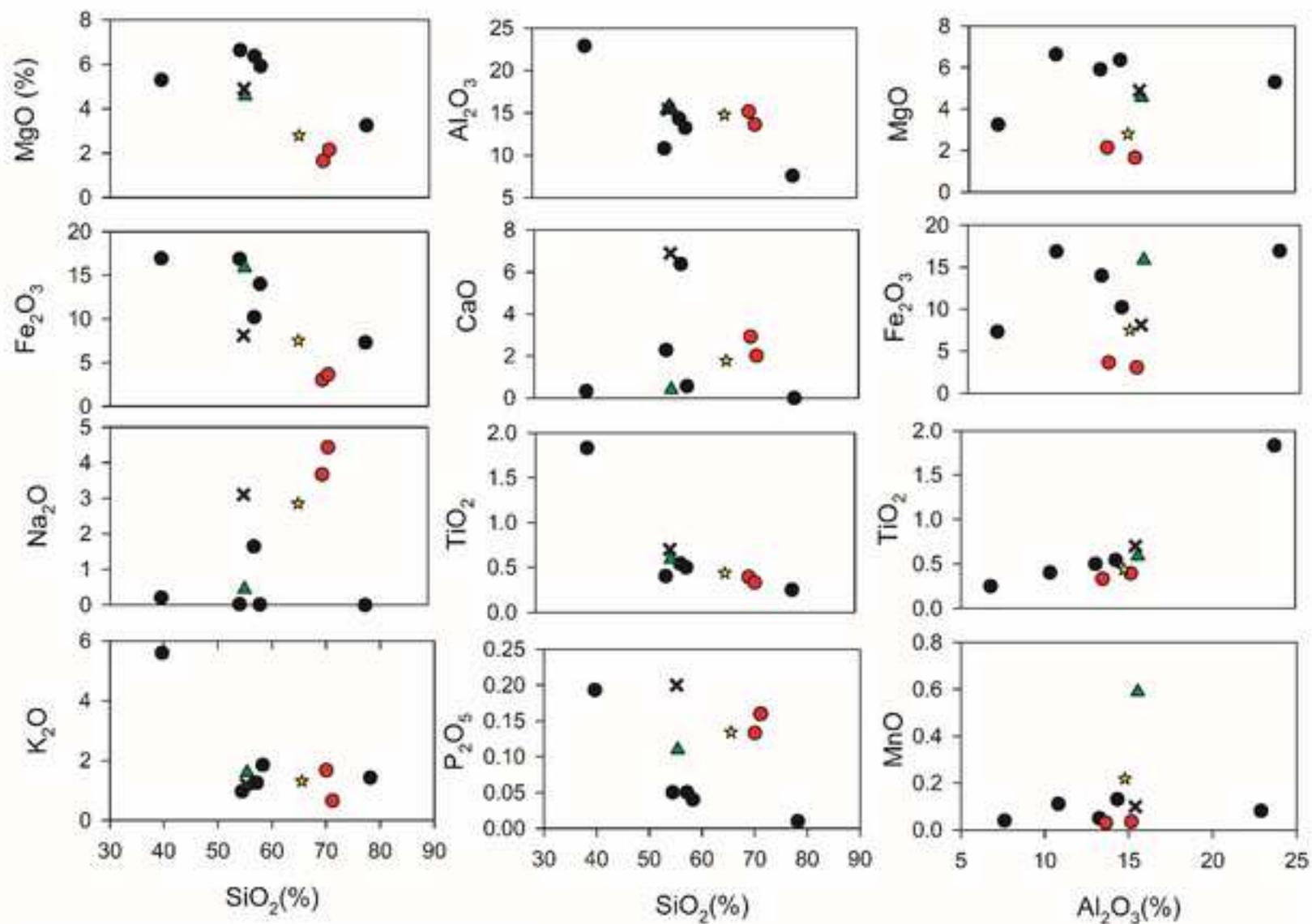


Figure 11



- NSB Quartz-Biotite Schists
- USB Quartz-Biotite Schists
- ▲ Averaged Witwatersrand shales
- ★ Isua felsic schists
- ✕ Archean continental crust

Figure 12

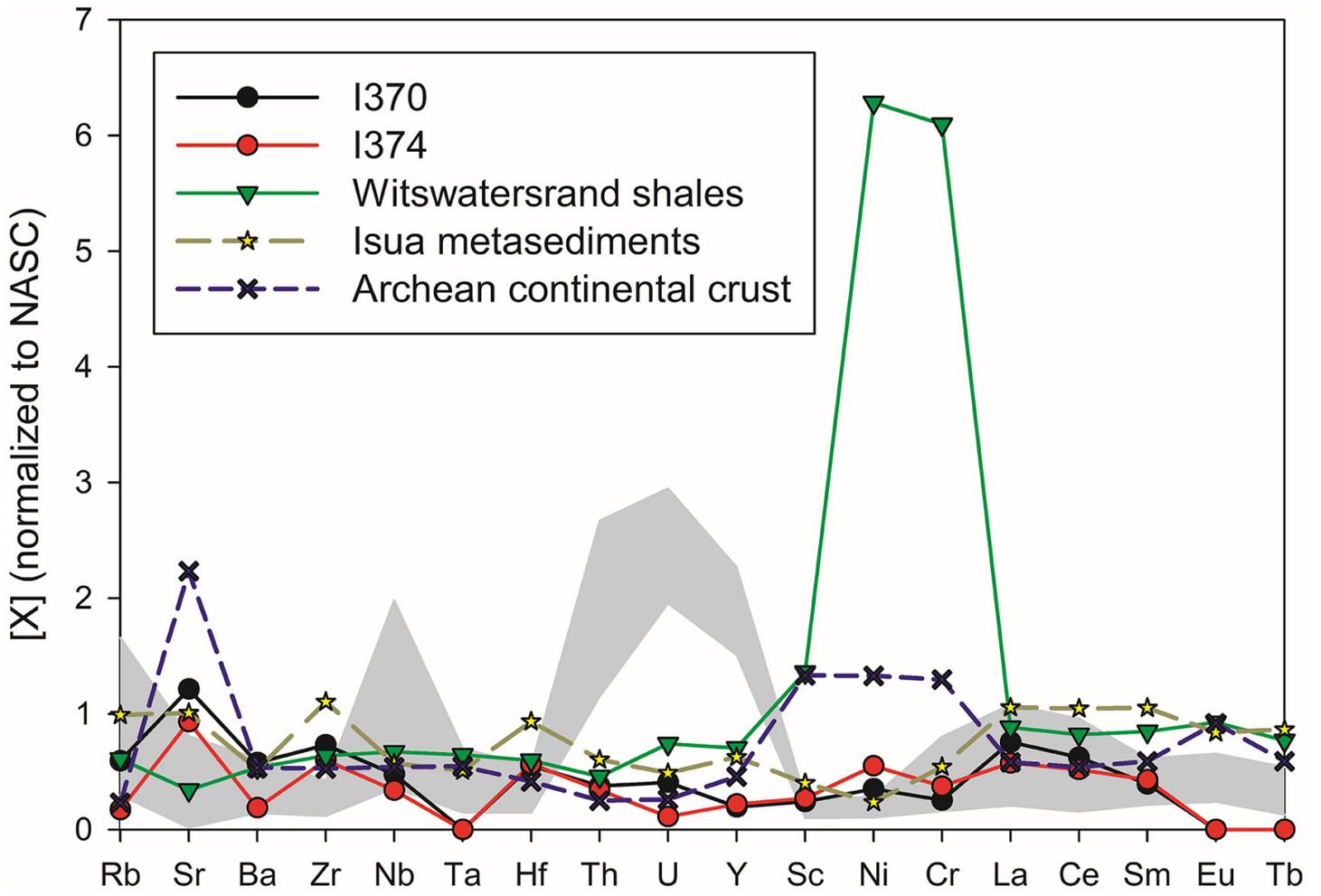


Figure 13

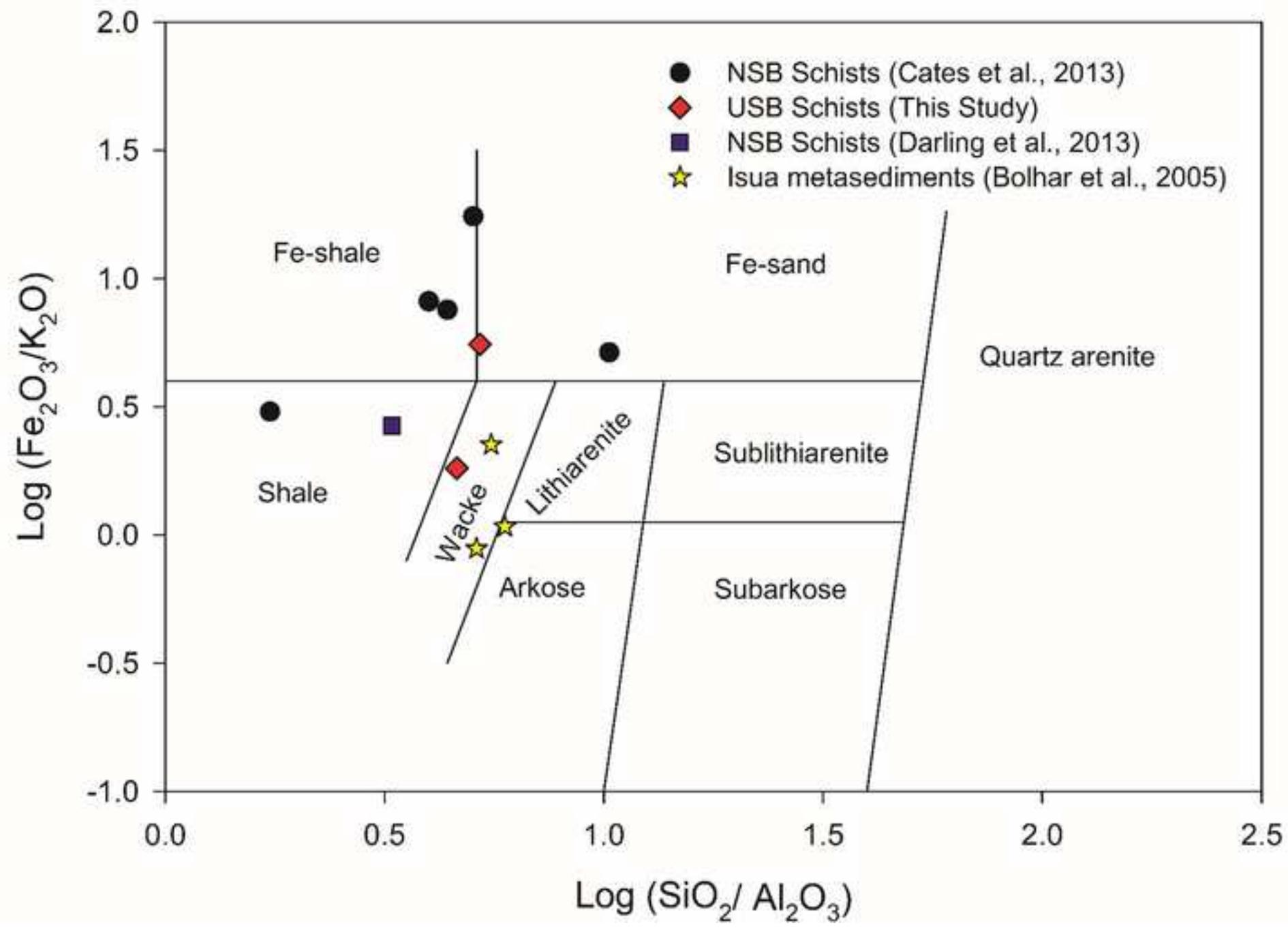


Table 1

Sample ID	Sample type	GPS Location	
		Latitude	Longitude
I239	Fuchsite-bearing Quartzite	58.27907°	-77.73080°
I240	Fuchsite-bearing Quartzite	58.27897°	-77.73091°
I346	Tonalitic orthogneiss	58.318044°	-77.692306°
I349	Granitic Orthogneiss	58.317987°	-77.692536°
I355	Tonalitic Gneiss	58.318018°	-77.692669°
I364	Hornblende Schist	58.317633°	-77.694142°
I365	Granitic Gneiss	58.317739°	-77.693695°
I369	Cumingtonite Schist	58.317037°	-77.693334°
I370	Quartz-Biotite Schist	58.317055°	-77.693349°
I373	Tonalitic Gneiss	58.316982°	-77.693538°
I374	Quartz-Biotite Schist	58.316985°	-77.693595°
I226	Unconsolidated Sediment	58.285324°	-77.697742°
I235	Unconsolidated Sediment	58.275978°	-77.684671°
I247	Unconsolidated Sediment	58.282120°	-77.728605°
I300	Unconsolidated Sediment	58.300750°	-77.688830°

Table 2

Sample ID	U-Pb Age(Ga)				Th/U
	LA-ICPMS (Average)	LA-ICPMS (Oldest)	SIMS (Average)	SIMS (Oldest)	
I239	-		-	-	-
I240	-		-	-	-
I295	-		-	-	-
I346	3.34 ± 0.05	3.50 ± 0.05	-	-	0.32 ± 0.02
I349	3.37 ± 0.05	3.49 ± 0.04	-	-	0.33 ± 0.03
I355	3.26 ± 0.05	3.52 ± 0.04	-	-	0.54 ± 0.06
I364	-		-	-	-
I365	3.31 ± 0.09	3.47 ± 0.08	-	-	0.26 ± 0.03
I369	-		-	-	-
I370	3.66 ± 0.06	3.78 ± 0.08	3.68 ± 0.01	3.76 ± 0.01	0.44 ± 0.06
I373	3.66 ± 0.06	3.68 ± 0.07	-	-	0.60 ± 0.05
I374	3.68 ± 0.07	3.79 ± 0.04	3.67 ± 0.01	3.75 ± 0.01	0.52 ± 0.07
I375	-		-	-	-
I226	2.68 ± 0.12	3.49 ± 0.12	-	-	0.69 ± 0.13
I235	2.77 ± 0.08	3.56 ± 0.06	-	-	0.75 ± 0.12
I247	2.57 ± 0.08	3.21 ± 0.08	-	-	0.26 ± 0.04
I300	2.74 ± 0.08	3.45 ± 0.05	-	-	0.64 ± 0.09