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1 Over one billion years of Archean crust 2 evolution revealed by zircon U-Pb and Hf 3 isotopes from the Saglek-Hebron Complex

4 1. Introduction

5 Although the extent of the early felsic crust is debated, Archean cratons are mostly composed of silica-
6 rich rocks from the tonalite-trondhjemite-granodiorite (TTG) suites, which abundance appears to
7 significantly decrease after 2500 Ma (Laurent et al., 2014; Moyen and Laurent, 2018; Moyen and
8 Martin, 2012). The geochemical composition of TTG differs from that of the modern continental crust,
9 suggesting they may have formed through distinct processes (e.g. Moyen and Martin, 2012). The study
10 of ancient terrain containing TTG, therefore, is crucial to our understanding of how continents formed
11 and stabilized. The Saglek-Hebron Complex (SHC), located in Northern Labrador, Canada, is a
12 polymetamorphic terrain dominated by TTG as old as 3900 Ma, and encompassing more than a billion
13 years of the geological record (e.g. Komiya et al., 2017). With such protracted geological history, the
14 SHC is one of the best candidates to decipher ancient crustal processes. Previous studies highlighted
15 the existence of multiple generations of granitoids within the SHC that exhibit a large variation of
16 chemical compositions and tonalite ages (Komiya et al., 2017; Krogh and Kamo, 2006; Sałacińska et
17 al., 2018; Schiøtte et al., 1989a; Shimojo et al., 2016; Vezinet et al., 2018). Most of these studies,
18 however, largely focused on the oldest Paleo to Eoarchean history of the SHC. This contribution
19 investigates the evolution of the SHC over the whole Archean Eon and combines detailed whole-rock
20 geochemistry of the wide compositional array of SHC granitoids, along with in-situ U-Pb and Lu-Hf
21 isotopic compositions of their zircon. The zircon U-Pb geochronology is used here to refine the
22 complex felsic magmatic and metamorphic history of the SHC over more than one billion years, while

23 the in-situ zircon Hf isotopic compositions, combined with the whole-rock geochemical composition
24 of the host rocks, are used to unravel the evolution of the crustal sources involved. Together, these
25 tools contribute to better understand the timing, the formation, and the long-term evolution of the SHC
26 and the overall Archean sialic crust.

27 **2. Geological context**

28 The Saglek-Hebron Complex, located on the east coast of Northern Labrador in Canada, is dominated
29 by orthogneisses from the TTG series along with granitic rocks and includes meter to kilometer scale
30 supracrustal enclaves (Fig. 1). Supracrustal rocks mainly include mantle-derived rocks (mafic
31 metavolcanic and ultramafic rocks), and chemical or clastic metasediments (Baadsgaard et al., 1979;
32 Bridgwater et al., 1975; Komiya et al., 2015; Nutman and Collerson, 1991). The metavolcanic rocks
33 are divided into two distinct units, the Upernavik assemblage interpreted to be Mesoarchean and the
34 Nulliak assemblage interpreted to be Eoarchean (Bridgwater and Schiøtte, 1991; Morino et al., 2018,
35 2017; Nutman et al., 1989; Schiøtte et al., 1992). Both units, however, are compositionally similar and
36 interpreted as a series of tholeiitic basaltic flows displaying some extent of differentiation (Wasilewski
37 et al., 2019). The SHC orthogneisses are divided into four main units including the Iqaluk grey gneiss
38 (sometimes referred to as the Nanok gneiss), the Uivak gneiss, the Lister gneiss, and late granitic
39 intrusions. The oldest TTG unit in the SHC is the Iqaluk grey gneiss dated at ~3900 Ma (Collerson,
40 1983a; Komiya et al., 2017; Regelous and Collerson, 1996; Shimojo et al., 2016). The Uivak gneiss
41 are the predominant lithology in the SHC, and were originally subdivided into two units based on their
42 age and mineralogy, with the older tonalitic Uivak I dated between 3863 Ma and 3732 Ma (Sałacińska
43 et al., 2018; Schiøtte et al., 1989b; Vezinet et al., 2018) and the younger granodioritic Uivak II dated at
44 ~3600 Ma (Hurst et al., 1975; Komiya et al., 2017; Sałacińska et al., 2018). However, a recent study

45 argued that the Uivak gneiss rather includes five units produced throughout the Eoarchean, from 3890
46 to 3610 Ma, and thus form almost continuous protracted magmatism occurring over more than 250
47 million years (Komiya et al., 2017). The younger granitoids include Paleoproterozoic TTG called the
48 Lister gneiss consisting of granodioritic intrusions emplaced between 3240 Ma and 3350 Ma (Komiya
49 et al., 2017; Schiøtte et al., 1989b) and Neoproterozoic granitic rocks intruding the SHC with a magmatic
50 peak occurring at ~2766 Ma (Collerson, 1983a, 1983b; Schiøtte et al., 1989b).

51 The SHC records at least two major protracted thermal episodes leading to some extent of crustal
52 reworking around ~3620 Ma and high-grade metamorphism up to granulite facies around ~2760-
53 2600 Ma (Bridgwater and Collerson, 1976; Hurst et al., 1975; Kusiak et al., 2018; Nutman and
54 Collerson, 1991; Sałacińska et al., 2018; Schiøtte et al., 1992, 1986; Van Kranendonk, 1990). Based on
55 detrital zircon, it has also been suggested that the Neoproterozoic thermal event could have been caused by
56 a massive collision and terrane accretion after the emplacement of the ~3300 Ma Upernavik
57 metasedimentary assemblage (Schiøtte et al., 1992). A recent geochronology investigation on
58 monazites and apatites suggests two thermal closure ages at 2500 and 2200 Ma, interpreted either as
59 cooling ages, or as two separate successive thermal events reaching, respectively, upper amphibolite
60 and greenschist facies metamorphic conditions (Kusiak et al., 2018).

61 **3. Methods**

62 Forty-seven orthogneiss samples have been collected for this study. All samples have been analysed
63 for whole-rock geochemistry. A subset of eighteen samples have been selected for coupled to U-Pb
64 and Hf in-situ analysis in zircon. These samples have been selected to cover the full compositional
65 range of SHC granitoids and to include all felsic lithologies previously described in the literature. This

66 includes type localities for the Iqaluk gneiss described by [Shimojo et al. \(2016\)](#) in the Kangidluarsuk
67 Inlet (St John's Harbour), the Uivak II gneiss described by [Bridgwater et al.\(1975\)](#) on the opposite
68 coast of Nulliak Island and “White Point” and the Lister gneiss found on Lister Island. The outcrop
69 studied by Shimojo et al. (2016) on which they obtained an age of 3920 ± 49 Ma and interpreted as the
70 oldest rock of the complex was resampled because of its significance and given that its exact age was
71 questioned due the relatively large imprecision ([Whitehouse et al., 2019](#)). Specific locations
72 previously studied by [Bridgwater et al., \(1975\)](#) were also targeted given the paucity of
73 geochronological data on the Uivak II and Lister gneiss from these localities. The detailed
74 geochronology on multiple generations of orthogneisses provided the framework to the Hf isotope
75 work and helped targeting the best-preserved zircon grains to analyse for Hf isotopic compositions
76 which are used to better understand the crustal evolution of the SCH over the whole Archean Eon.
77 [Figure 1](#) shows the main sample locations and GPS coordinates for all samples are given in [Table 1](#).
78 Selected outcrop photographs are available in the supplementary material [Figure S1](#)). The full methods
79 regarding whole-rock major and trace element geochemistry can be found in the supplementary
80 material ([Supplementary Table S1](#)).

81 Eighteen samples from our set of 47 granitoids were analyzed for both in-situ U-Pb geochronology and
82 Hf isotopes in zircon. Rock samples were crushed using a steel jaw crusher and disk mill and then
83 sieved to collect grain sizes between 250-106 μm . Heavy minerals, such as zircon and metallic oxides,
84 were separated using a water shaking table and methylene iodide heavy liquids. The heavy mineral
85 fractions were then passed through a Frantz magnetic separator to remove the magnetic minerals.
86 Between 100 and 120 zircon grains were then handpicked, mounted on an epoxy resin and polished.
87 Cathodoluminescence (CL) images of the polished zircon grains were taken using the JEOL 6610LV
88 scanning electron microscope (SEM) at the University of Ottawa, to identify the different zircon zones

89 and guide the laser ablation work. Both U-Th-Pb and Hf isotope analyses were undertaken at the
90 Laboratoire Magmas et Volcans (LMV) (Clermont-Ferrand, France). U-Th-Pb isotope analyses were
91 performed on a Thermo Scientific Element XR-ICP-MS, and Hf isotope analyses were conducted on a
92 Thermo Scientific Neptune Plus multicollector ICP-MS, both instruments were coupled with a
93 Resonetics M50E 193 nm excimer laser ablation (LA) system. The analytical method for isotope
94 dating by LA-ICP-MS described in [Hurai et al.\(2010\)](#) and [Paquette et al.\(2014\)](#) were followed and
95 more details on analytic conditions are available in the supplementary material Table S1. Primary (GJ-
96 1) and secondary (91500) zircon reference materials have been measured to ensure the quality of the
97 data over the different analytic sessions and data are found in the supplementary material Table S1 and
98 Table S2. Common Pb was not corrected owing to the large isobaric interference of ^{204}Hg . The ablated
99 zircon zones were carefully chosen in order to avoid any mixed zones and Hf isotopic measurements
100 were performed on the same spots than those chosen for the U-Pb analyses according to analytical
101 techniques described in [Moyen et al. \(2017\)](#) and [Paquette et al. \(2017\)](#). The full Lu-Hf isotope data are
102 available in the supplementary material Table S3. Initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios were calculated using the
103 $\lambda^{176}\text{Lu}$ decay constant of $1.867 \times 10^{-11} \text{ yr}^{-1}$ of [Söderlund et al. \(2004\)](#) and the CHUR parameters of
104 $^{176}\text{Hf}/^{177}\text{Hf} = 0.282785$ and $^{176}\text{Lu}/^{177}\text{Hf} = 0.0336$ ([Bouvier et al., 2008](#)) were used for calculation of
105 ϵHf values.

106 **4 Results**

107 **4.1 Petrography & Geochemistry**

108 Although felsic rocks from the SHC have been metamorphosed and are commonly gneisses, we use
109 the term “granitoid” *sensu lato* in this study to refer to the felsic plutonic rocks. Similarly, the terms
110 granite, tonalite, trondhjemite, granodiorite are use to refer to a mineralogical and geochemical

111 composition regardless of metamorphic fabrics. Based on observed mineralogy and CIPW normative
112 compositions, the SHC granitoids can be divided into four distinct groups, including trondhjemite,
113 tonalite, granodiorite (grouped as the TTG, *s.l.*) and granite (Fig. 2). From the TTG studied here, only
114 a few samples of granodiorite have been analyzed, but they are compositionally similar to other SHC
115 granodiorite previously described by Schiøtte et al. (1993) (Fig. 2). The TTG are commonly medium
116 grained and exhibit variable degrees of deformation (Fig. S1 and Fig. S2) with common evidence of
117 migmatization (Fig. S1). Pegmatitic and porphyritic textures can also occasionally be observed (Fig.
118 S1 d. e. f.). Trondhjemite samples (Fig. S1 b. g. i. and Fig S2 e. f. g. h. i.) consist of typical grey gneiss
119 composed of quartz + oligoclase + biotite + titanite ± zircon ± apatite. Compared to the trondhjemite,
120 rocks that exhibit tonalitic to granodioritic compositions can be described as melanocratic grey gneiss
121 containing higher amounts ferromagnesian minerals such as clinopyroxene and biotite (Fig. S1 a. c. e.
122 f. j. k. and S2 k. l. m. n.). The granodiorite locally exhibits augen textures at the outcrop scale (*e.g.*
123 Ilulik & White Point; Fig S1.e), which is not observed in trondhjemite. The granite is composed of
124 quartz + orthoclase + oligoclase ± biotite ± zircon ± apatite ± garnet and typically defines meter to tens
125 of meter scale leucocratic units (Fig. S1 d. and S2 a. b. c. d. j.). Granitic rocks commonly occur as
126 migmatite veins but also as larger plutonic bodies. Granite usually shows little fabric and minor
127 deformation compared to TTG.

128 The TTG display a wide range of silica content (56.6 wt. % to 76.8 wt. %SiO₂), with most samples
129 having below 72 wt. %SiO₂. The tonalite exhibits the lowest SiO₂ contents among the TTG and
130 display high Mg concentrations relative to the other SHC granitoids (Fig. 3a), and therefore are here
131 referred to as Mg-rich tonalite. Except for one sample, the granite shows high and uniform silica
132 contents ranging from 73.0 to 77.5wt. %SiO₂ (Fig. 3a). Both the Mg-rich tonalite and the granodiorite
133 display relatively high FeO_t and TiO₂ contents compared to the granite and the trondhjemite (Fig. 3b).

134 Compared to all TTG, the granite exhibits higher concentrations in K_2O (from 3 to 6 wt.%; Fig. 3c)
135 and Rb, and lower CaO concentrations (<2 wt. %; Fig. 3e). The tonalite and trondhjemite exhibit
136 higher Na_2O , CaO, and Al_2O_3 but lower K_2O compared to the granite and granodiorite (Fig. 3c-d-e).
137 The Mg-rich tonalite, however, exhibits lower A/CNK ratios relative to the trondhjemite, with the
138 granite showing the highest A/CNK ratios (Fig. 3f).

139 All granitoids exhibit pronounced negative Nb-Ta anomalies (Fig. 4a) along with strong depletion in
140 heavy rare earth elements (HREE) compared to light rare earth elements (LREE, Fig. 4b). Mg-rich
141 tonalite and granodiorite are generally characterized by lower $(La/Yb)_N$ ratios than most trondhjemite
142 and granite samples (Fig. 5). The trondhjemite samples exhibit higher $(La/Yb)_N$ ratios (up to 150) and
143 most samples display positive Eu anomalies (Fig. 4b; Fig. 5). The trondhjemite samples collected on
144 White Point, however, display higher HREE contents and thus lower $(La/Yb)_N$ ratios compared to
145 other trondhjemite samples. The trondhjemite sample SG-019 collected on Ukkalek Island exhibits
146 extremely low trace element and REE concentrations, a prominent positive Eu anomaly and mostly
147 plots outside of the compositional field comprising the other trondhjemite samples (Fig. 4). The Lister
148 gneiss (SG-265) is compositionally similar to the Mg-rich tonalite but exhibits the lowest REE
149 concentrations (Fig. 4b). Granitic samples exhibit high Rb and Pb with lower Sr concentrations,
150 relative to all other TTG. In general, granite samples are more enriched in LREE compared to the
151 trondhjemitic rocks with large variability in $(La/Yb)_N$ ratios ranging from 0-300 (Fig. 5). The granite
152 samples commonly display small positive or negative Eu anomalies while the leucogranitic samples
153 SG-007 and SG-017 exhibit very low REE concentrations with pronounced positive Eu anomalies
154 ($Eu_N/Eu^* \sim 60$) (Fig. 4b).

4.2 U-Pb Geochronology

Zircon grains from 18 granitoids have been analyzed by LA-ICP-MS for U-Pb geochronology. Samples were selected to comprise representative samples of each rock types with an effort to include the different generations of granitoids known in the SHC. This selection was based on whole-rock geochemistry, overall quality of sample (no veins, limited weathering, or alteration) and field relationship within the SHC. For each sample, a set of 100 to 120 zircon grains have been mounted for analysis. Data for all individual zircon analyses are available in supplementary material Table S3 and Figure S3. Analyses that were considered as of lesser quality (e.g. presence of ^{204}Pb) were discarded. Therefore, some sample yielded a smaller number of analyzed zircon grains, especially those including a larger amount of metamict zircon grains (e.g. SG-017, SG-019 and SG-080). Figure 6 shows cathodoluminescence (CL) zircon images for 6 samples (all CL images are included in Figure S3) and the Concordia plots for each sample are displayed in Figure 7. Zircon grains exhibiting inconsistent texture in CL imaging with numerous inclusions and a high uranium content have been characterized as metamict (e.g. B11, B12 from the SG-143 granite; Fig. 6-f.). Recrystallized zircon grains are mostly rounded and exhibits black color in CL imaging. Recrystallized rims are commonly following the pre-existent zonation or irregularly crosscut the zoned core in not fully recrystallized zircon grains. They often exhibit a high uranium content and a younger age than the core. Table 2 presents a summary of the crystallization ages, inherited ages and metamorphic ages obtained for each sample analyzed. Reported ages for each sample in Table 2 were determined from either the Concordia age, for samples with populations of clustered concordant zircon analyses, the weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age, for populations of less concordant zircon analyses, or the upper intercept age, for populations with mostly discordant zircon analyses. For samples with populations showing concordant results over a wide

177 range of ages, indicative of ancient Pb-loss, the oldest concordant zircon analyses were considered as
178 the most representative of the crystallization age.

179 *a. Trondhjemite*

180 Four trondhjemite samples: SG-019, SG-024, SG-025, and SG-122, have been analyzed for U-Pb on
181 zircon.

182 Zircon grains from sample SG-019 are mostly metamict and rounded. A total of 33 analyses were
183 performed (Fig. 7a). The main population of zircon yields a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of
184 $2732 \pm 25 \text{ Ma}$ (MSWD=2.2; $n = 12$). These zircon grains consist of homogeneous black core to
185 oscillatory zoned grains, which exhibit an average U concentration of $\sim 470 \text{ ppm}$ (All $< 700 \text{ ppm}$).
186 “Black cores” or “black grains” refer to the black CL imaging rendered in response to a relatively high
187 U content. A resolvable second population mostly consists of metamict zircon grains (U concentrations
188 from 600 to 2860 ppm; average $\sim 1600 \text{ ppm}$) and yields a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of
189 $2574 \pm 14 \text{ Ma}$ (MSWD= 0.46; $n = 14$). Two older zircon cores yield Concordia ages at $3477 \pm 18 \text{ Ma}$
190 and $3553 \pm 15 \text{ Ma}$. Given the large number of metamict zircon grains in this sample, and the relatively
191 small number of analyses compared to other trondhjemite samples, the crystallization age of SG-019 is
192 ambiguous. If the 2732 Ma population is taken as representing the crystallization age, the 2
193 Paleoproterozoic zircon grains would likely be inherited. Contrastingly, the Neoproterozoic ages could
194 represent reset ages or secondary recrystallization, in which case the older zircon grains could be more
195 representative of the crystallization age. Nevertheless, the geochronological data from this sample
196 must be taken with caution.

197 A total of 128 U-Pb analyses have been conducted on sample SG-024 (Fig. 7b). Zircon grains are
198 dominated by two main populations that exhibit U concentrations $< 1000 \text{ ppm}$, associated with a wide

199 range of Th/U ratios (0.01-2.83). A number of zircon grains from sample SG-024 are subeuhedral to
200 euhedral and exhibit well defined oscillatory zoning (e.g. zircon #A17: Fig. 6a). The few concordant
201 analyses plot between 3800 and 3870 Ma. However, most grains from this population are discordant
202 and spread along a poorly defined discordia line (Fig. 7b). The two oldest concordant cores yield a
203 Concordia age of 3869 ± 10 Ma (MSWD_(Concordance+Equivalence) = 1.5; $n = 2$) considered as the
204 crystallization age for this sample. A population of Neoproterozoic zircon is dominated by subeuhedral to
205 rounded zircon grains that show clear sector zoning and exhibits a higher average Th/U ratio of 1.64
206 (e.g. zircon #A05: Fig. 6a), which altogether yield a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2721 ± 8 Ma
207 (MSWD = 0.89; $n = 42$). A few Neoproterozoic oscillatory zoned rims and cores with lower Th/U ratios
208 (<0.2) are also present. However, these Neoproterozoic zircon grains exhibit much more scattered ages
209 and yield a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2769 ± 19 Ma (MSWD = 4; $n = 28$).

210 Zircon grains from sample SG-025 are dominated by oscillatory and sector zoned textures to
211 homogeneous black in CL images. A total of 122 U-Pb analyses have been done on sample SG-025.
212 All analyses of oscillatory zoned zircon spread along the Concordia curve from 3838 to 3000 Ma (Fig.
213 7c). The older zircon grains from this population are characterized by higher Th/U ratios, up to 0.81,
214 relative to the average Th/U ratio of oscillatory zoned zircon of 0.16, whereas the zircon grains
215 displaying younger measured $^{207}\text{Pb}/^{206}\text{Pb}$ ages show more evidence of ancient Pb-loss. Therefore, the
216 crystallization age for this sample is best represented by the oldest concordant zircon grain with a
217 Concordia age of 3838 ± 10 Ma. Sample SG-025 also includes a population of sector zoned and black
218 zircon with high Th/U (>0.2) and yielding Neoproterozoic ages. These grains cluster around a weighted
219 mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2747 ± 5 Ma (MSWD = 0.54; $n = 91$).

220 A total of 105 U-Pb analyses were conducted on sample SG-122 (Fig. 7d). Zircon grains from this
221 trondhjemite sample are dominated by rounded grains with oscillatory zoned textures and

222 homogeneous black grains in CL images. The population of oscillatory zoned zircon scatters along a
223 discordia line with an upper intercept age at 3752 ± 33 Ma (MSWD=13; $n = 62$). Moreover, a number
224 of concordant zircon analyses from this population spread along the Concordia curve between ~ 3790
225 and ~ 3700 Ma, suggestive of ancient Pb-loss. The oldest concordant zircon from this scattered
226 population gives a Concordia age of 3781 ± 12 Ma, representing the crystallization age for this sample.
227 This euhedral zircon grain shows a clear oscillatory zoning and a relatively high Th/U ratio (~ 0.2). A
228 subordinate population of U-rich rims yields a younger weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2804 ± 9 Ma
229 (MSWD= 0.81; $n = 28$).

230 *b. Mg-rich tonalite*

231 Four Mg-rich tonalite samples have been analyzed for U-Pb on zircon, including sample SG-026, SG-
232 027, SG-210c and SG-265.

233 Sample SG-026 is dominated by oscillatory zoned to sector zoned subeuhedral zircon grains (e.g.
234 zircon #B16 & #E12; Fig. 6b) that exhibit variable U concentrations (20 to 1000 ppm) and relatively
235 high Th/U ratios (>0.6) compared to zircon grains from other samples. A total of 115 U-Pb analyses
236 were performed on sample SG-026, which includes a single zircon population defining a discordia line
237 with an upper intercept at 3820 ± 20 Ma, interpreted as the crystallization age, and a lower intercept at
238 2514 ± 71 Ma (MSWD=1.12; $n = 103$; Fig. 7e). Most zircon grains exhibit recrystallized black and
239 white rims, that systematically yield younger and more discordant ages compared to the zoned cores.

240 A total of 138 analyses were done for the sample SG-027, which includes three main zircon
241 populations (Fig. 7f). Zircon grains from the oldest population exhibit oscillatory zoning and igneous
242 Th/U ratios from 0.2 to 0.5. Several zircon grains from this population are concordant, spreading on the
243 Concordia curve between ~ 3600 and ~ 3500 Ma, which may reflect ancient Pb-loss. Therefore, the

244 average age of the two oldest concordant zircon analyses of 3632 ± 9 Ma (MSWD_(C+E) = 1.3; n =2) is
245 taken as the crystallization age. The other zircon populations are Neoproterozoic, which include two
246 groups characterized by metamict textures and black CL images, commonly occurring as rims around
247 older cores. One Neoproterozoic population exhibits lower Th/U ratio and yields a weighted mean
248 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2785 ± 7 Ma (MSWD= 1.3; n =46), while a secondary smaller population with
249 higher Th/U ratios (>0.4) yields a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2595 ± 20 Ma (MSWD= 0.93; n
250 =6).

251 Sample SG-210c was collected from the same outcrop and lithology as the sample previously dated at
252 3920 ± 49 Ma (Shimojo et al., 2016), the oldest reported U-Pb age in the SHC. A total of 120 U-Pb
253 analyses were obtained from this sample. A number of zircon analyses scatter along the Concordia
254 curve from ~3900 to ~3500 Ma defining two populations (Fig. 7g). The oldest concordant zircon
255 analyses display a relatively homogenous population exhibiting oscillatory zoning (e.g. zircon #A01;
256 Fig. 6c), with high Th/U ratios (> 0.3) and a Concordia age of 3869 ± 6 Ma (MSWD_(C+E) = 1.11; n =34)
257 interpreted as the crystallization age. The younger zircon analyses show a fair amount of scattering and
258 a progressive decrease of Th/U ratios (to <0.1) as ages get younger. This population consists of U-rich
259 recrystallized zircon, black on CL images (e.g. zircon #A02; Fig. 6c) and displays a weighted mean
260 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3558 ± 11 Ma (MSWD=0.49; n =19). A single zircon yields a Concordia age of
261 2720 ± 16 Ma (F02, black rounded grain).

262 A total of 86 U-Pb analyses were performed on sample SG-265 that was collected on Lister Island. All
263 zircon grains from this sample show clear and well-defined oscillatory zoning as well as relatively
264 high Th/U ratios, between 0.3 and 0.5. Zircon analyses from a single population define a discordia line
265 with an upper intercept age of 3229 ± 8 Ma (MSWD= 1.3; n =70; Fig. 7h). The most concordant zircon

266 analyses yield a Concordia age of 3224 ± 7 Ma (MSWD_(C+E) = 0.13; $n = 13$) and considered as the
267 crystallization age.

268 *c. Granodiorite*

269 One sample of granodiorite SG-203 has been analyzed for U-Pb with a total of 121 in-situ analyses
270 (Fig. 7i). The main population of zircon exhibits oscillatory zoning and high Th/U ratios ranging
271 between 0.3 and 0.5. They define an upper intercept age of 3330 ± 15 Ma (MSWD=1.2; $n =103$)
272 interpreted to be the crystallization age. Two additional minor populations of zircon with lower Th/U
273 ratios (<0.1) yield younger weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2975 ± 17 Ma (MSWD= 1.02; $n =9$) and
274 2703 ± 30 Ma (MSWD=0.69; $n =3$).

275 *d. Granite*

276 Nine granite samples have been analyzed for U-Pb on zircon, including sample SG-007, SG-017, SG-
277 037, SG-080, SG-084, SG-087, SG-127, SG-143, and SG-208. All granitic samples were collected
278 from large granite exposures, except for leucogranite sample SG-017 collected from meter scale veins
279 between tonalite and ultramafic rocks.

280 A total of 98 U-Pb analyses have been conducted on sample SG-007 (Fig. 7j). Zircon grains are
281 dominantly elongated subeuhedral grains exhibiting oscillatory zoning or metamict textures. Most
282 zircon grains with oscillatory zoned cores show thin recrystallized rims. This sample displays zircon
283 grains with variable ages and multiple heterogeneous populations. Concordant zircon analyses display
284 a wide range of ages from 3883 to 3300 Ma which correlates with a decrease of Th/U ratios (from 0.7
285 to <0.1). The three oldest concordant grains consist of oscillatory zoned low U cores with ages
286 spreading on the Concordia from 3883 ± 10 Ma (Concordia age from zircon #C5;Fig. 6d) to
287 3853 ± 12 Ma (Concordia age from zircon #E10-c;Fig. 6d), with a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean age =

288 3869 ± 26 Ma MSWD=0.41; $n=3$). The younger concordant ages mostly correspond to rims around
289 oscillatory zoned older cores. On a Kernel density estimation (KDE) diagram (Fig. 7j), two high-
290 density peaks of apparent ages yield weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages at 3521 ± 19 Ma (MSWD = 5.4;
291 $n=35$) and 3815 ± 22 Ma (MSWD= 3.1; $n=15$) indicative of ancient Pb-loss around 3800 and
292 3500 Ma. A younger concordant population consisting of recrystallized rims to completely metamict
293 grains (e.g. zircon #C6; Fig. 6d) with very high U concentrations (up to 2599 ppm) and a low Th/U
294 ratio (<0.1), yields a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2769 ± 22 Ma. A third population of U-rich
295 metamict zircon rims (up to 5790 ppm) yields a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2649 ± 17 Ma
296 (MSWD =1.7; $n=9$).

297 A total of 22 U-Pb analyses were conducted on sample SG-017 (Fig. 7k). It is dominated by round to
298 elongated metamict zircon grains. Two zircon populations with low Th/U ratios (<0.1) yield weighted
299 mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2802 ± 13 Ma (MSWD= 1.2; $n=13$) and 2712 ± 13 Ma (MSWD= 1.2; $n=4$).
300 A single zircon grain, perhaps inherited, yielded a Concordia age of 3576 ± 14 Ma.

301 A total of 99 U-Pb analyses were obtained on sample SG-037 (Fig. 7l), collected within a granitic unit
302 in contact with mafic metavolcanic rocks. Most zircon grains from this granite sample are black on CL
303 images and display metamict cores or rare zoned cores with high Th/U ratios (>0.2). A consistent
304 zircon population (Fig. 7l) yields a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2744 ± 8 Ma (MSWD=0.6; n
305 =44). Younger grains spread along the Concordia curve from 2744 to 2200 Ma and mostly consist of
306 metamict grains with high U concentrations (700-3800 ppm) and relatively low Th/U ratios (<0.3).
307 Two older discordant inherited zircon grains are also present, with the oldest one yielding a $^{207}\text{Pb}/^{206}\text{Pb}$
308 age of 3303 ± 48 Ma.

309 A total of 38 U-Pb analyses were conducted on sample SG-080 (Fig. 7m). The dominant zircon
310 population is characterized by high Th/U ratios and oscillatory zoned grains yielding a weighted mean
311 $^{207}\text{Pb}/^{206}\text{Pb}$ age of $3612 \pm 9 \text{ Ma}$ (MSWD=0.71; $n =26$), interpreted as the crystallization age. Two
312 inherited zircon cores appear to exhibit older ages with one concordant analysis (e.g. zircon #C15; Fig.
313 6e) which yielded a Concordia age of $3805 \pm 15 \text{ Ma}$. Two low Th/U (<0.3) homogeneous black zircon
314 rim and core, yield Concordia ages of $2752 \pm 9 \text{ Ma}$ (MSWD_(C+E)=1.4; $n =2$).

315 A total of 18 analyses were conducted on the sample SG-084 (Fig. 7n). This granitic sample is
316 intruded by pegmatites and mostly includes metamict zircon grains commonly containing many
317 inclusions, explaining the limited number of data obtained from this sample. Only 10 oscillatory zoned
318 cores yield a poorly defined population with a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of $3710 \pm 24 \text{ Ma}$
319 (MSWD= 1.9; $n =10$)

320 A total of 54 analyses were obtained on sample SG-087 (Fig. 7o). Zircon grains from this granite
321 sample are dominated by sector and oscillatory zoned grains that exhibit variable Th/U ratios. Sector
322 zoned grains are grouped in three major clusters. Two younger populations yielded weighted mean
323 $^{207}\text{Pb}/^{206}\text{Pb}$ ages of $2755 \pm 13 \text{ Ma}$ (MSWD=1.2; $n =15$) and $2584 \pm 14 \text{ Ma}$ (MSWD=1.5; $n =14$)
324 whereas the oldest population defines a weak discordia line, but with the two oldest zircon analyses
325 yielding a Concordia age of $3758 \pm 10 \text{ Ma}$ (MSWD_(C+E)=1.3; $n =2$).

326 A total of 45 U-Pb analyses were conducted on sample SG-127 (Fig. 7p). Zircon grains with high
327 Th/U ratios (>0.2) showing oscillatory zoned cores with few metamict U-rich grains, yield a weighted
328 mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of $2789 \pm 12 \text{ Ma}$ (MSWD=0.83; $n =18$) interpreted as the crystallization age. A
329 younger population with lower Th/U ratios yields a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of $2662 \pm 18 \text{ Ma}$
330 (MSWD=0.68; $n =8$). One inherited zircon grain with clear oscillatory zoning yields a discordant

331 (~4%) $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3661 ± 48 Ma. All other analyzed zircon grains consist in low Th/U ratio
332 metamict grains with $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging from 2600 to 2200 Ma.

333 A total of 53 U-Pb analyses were done on sample SG-143 (Fig. 7q). This granite is dominated by
334 metamict zircon grains with subordinate euhedral oscillatory zoned textures (e.g. zircon #B11
335 and #B12; Fig. 6f). These zircon grains exhibit a wide range of Th/U ratios (from 0.02 to 0.7) with the
336 main population yielding a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2780 ± 8 Ma (MSWD=0.78; $n=42$) that
337 corresponds to the crystallization age. A smaller population of U-rich metamict grains yields a slightly
338 younger weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2706 ± 19 Ma (MSWD=1.7; $n=7$). Three inherited cores
339 yield older $^{207}\text{Pb}/^{206}\text{Pb}$ ages, with a single concordant grain exhibiting a Concordia age of
340 3655 ± 12 Ma.

341 A total of 77 U-Pb analyses were obtained on sample SG-208 (Fig. 7r). This granitic sample mostly
342 includes oscillatory zoned zircon grains with low U concentrations (31-400 ppm) and igneous Th/U
343 ratios ranging between 0.2 and 0.6. Concordant zircon analyses yield two different populations. The
344 first population gives a Concordia age of 3330 ± 7 Ma (MSWD_(C+E) = 1.4; $n=18$) and is considered as
345 the crystallization age. The second slightly younger population yields a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age
346 of 3241 ± 11 Ma (MSWD=0.68; $n=23$), possibly reflecting age resetting. Four zircon grains also
347 exhibit a younger weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age around 2946 ± 8 Ma (MSWD= 0.33; $n=4$).

348 **4.3 Zircon Lu-Hf isotopic compositions**

349 For each rock sample, a subset of 10 to 30 representative zircon grains has been analyzed for in-situ
350 Hf isotopes (supplementary material Table S4) and selected based on concordance ($\pm 2\%$) and
351 $^{207}\text{Pb}/^{206}\text{Pb}$ age (Table 2). Measurements were acquired on the same spot used for the U-Pb analyses
352 (supplementary material Figure S3). Figure 8 shows the initial ϵ_{Hf} values for each zircon grains,

353 calculated by using their corresponding $^{207}\text{Pb}/^{206}\text{Pb}$ age. However, given that the studied samples are
354 granitoids with a single crystallization age, the initial Hf isotopic composition for all of the zircon
355 grains from a single sample should be at the same crystallization age. Therefore, initial Hf isotope
356 compositions of each zircon analyses from the same sample were also calculated at the age of
357 crystallization reported in Table 2 and the average initial ϵHf values ($\pm 2\text{SE}$) for each sample are also
358 shown on Figure 8. This is performed in order to avoid calculating an initial Hf composition at an over
359 or underestimated $^{207}\text{Pb}/^{206}\text{Pb}$ age for zircon grains that would have suffered ancient Pb-loss but
360 remained on the Concordia curve (Fisher et al., 2014a, 2014b; Kemp et al., 2010) as illustrated in
361 Figure 9. Figure 8 shows that Eoarchean samples (3883 to 3781 Ma) display initial Hf isotope
362 compositions that are suprachondritic, with average initial ϵHf values ranging between +3.7 to +1.7.
363 The Mg-rich tonalite sample SG-026 contains zircon grains exhibiting the most radiogenic Hf isotopic
364 composition with an average initial ϵHf value of $+3.7 \pm 0.2$ at 3820 Ma, falling on the depleted mantle
365 evolution line. Most samples with positive ϵHf initial values are tonalitic or trondhjemitic in
366 composition, except for granitic sample SG-007 showing a complex zircon population (Fig. 7j) and
367 with initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios between 0.28027 and 0.28042 for all zircon grains analyzed and an
368 average ϵHf initial value of +2.0 at 3883 Ma. However, given the complex and heterogeneous age
369 populations of this sample, this initial ϵHf must be taken with cautious as its Eoarchean age is based on
370 the three oldest zircon grains, that may be inherited or recrystallized. If the crystallization age of most
371 zircon grains from this sample is rather Neoproterozoic, recalculation of initial ϵHf values at 3883 Ma for
372 all zircon analyses would not be representative of its source. Five granitoids with Paleoproterozoic ages
373 from 3632 to 3330 Ma display subchondritic initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios corresponding to average ϵHf
374 initial values from -1.0 to -6.3. The low ϵHf initial value of -4.6 at 3576 Ma for the oldest zircon grain
375 from the sample SG-017, however, is from an inherited grain and the main Neoproterozoic zircon

376 population for this sample shows an average ϵ_{Hf} of -12.7 ± 0.3 at 2802 Ma. Sample SG-203 is the only
377 granodiorite sample analyzed but displays the same 3330 Ma age and Hf isotopic composition as the
378 spatially associated granitic sample SG-208, with initial ϵ_{Hf} values of around -6. The later ~ 3200 Ma
379 Paleoproterozoic Mg-rich tonalite from Lister Island, sample SG-265, is characterized by a radiogenic
380 initial Hf isotopic composition and does not appear to follow the general ϵ_{Hf} trend vs. time that most
381 other granitoids display on Figure 8. This sample has a slightly suprachondritic initial ϵ_{Hf} value of
382 $+1.0 \pm 0.3$ at 3224 Ma. The late Mesoproterozoic granitic sample SG-087 is characterized by an initial ϵ_{Hf}
383 of -2.6 ± 0.2 at 2996 Ma. However, this sample has multiple and complex zircon generations (Fig. 7o)
384 and geochronology results should be taken with caution as it also includes few inherited older zircon
385 cores with initial $\epsilon_{\text{Hf}} = +1.5 \pm 0.4$ at 3745 Ma.

386 Three Neoproterozoic granite samples dated between 2789 and 2744 Ma have been analyzed for zircon Hf
387 isotopic composition. They all display low initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios corresponding to average ϵ_{Hf}
388 initial values of -11.2 to -14.3. These Neoproterozoic granite samples include a few inherited older grains
389 which ϵ_{Hf} values fall within the main time vs. initial ϵ_{Hf} array with two ~ 3660 Ma cores with initial
390 $\epsilon_{\text{Hf}} = -0.2$ to -1.7 and one 3303 Ma core with initial $\epsilon_{\text{Hf}} = -7.8$. The Neoproterozoic zircon population of
391 the trondhjemitic sample SG-019 yielded an average ϵ_{Hf} initial value of -11.3 if the 2732 Ma age is
392 considered as the crystallization age. This sample also includes two zircon cores with older ages of
393 3477 and 3553 Ma that respectively yielded initial ϵ_{Hf} values of -3.3 and +7.0. However, given the
394 ambiguity of the crystallization age of this sample, these initial ϵ_{Hf} values need to be considered with
395 prudence. A few Eoproterozoic to Paleoproterozoic granitoids display Neoproterozoic zircon rims (sample SG-
396 024, SG-025, and SG-027) with variable initial ϵ_{Hf} values ranging from ~ -15 to ~ -22 . These zircon
397 grains, however, show variable and often low Th/U ratios, which could be indicative of metamorphic

398 recrystallization. Therefore, the initial ϵ_{Hf} values calculated at these ages may not be reflective of their
399 crustal source.

400 **5 Discussion**

401 **5.1 Composition and petrogenesis of the SHC felsic crust**

402 The geochemical composition of Archean felsic rocks has been widely used to constrain the nature of
403 their sources and the various processes leading to the formation of Archean crust (e.g. Hoffmann et al.,
404 2019; Laurent et al., 2014; Moyen et al., 2001; Moyen and Martin, 2012; Whalen et al., 2002). Rocks
405 from the TTG series make up a major part of the Archean cratonic nucleus forming our stable
406 continents. Various petrogenetic models have been proposed to explain the origin of TTG, and several
407 lines of evidence point to an incompatible element enriched mafic source as their precursor (e.g.
408 Hoffmann et al., 2019, 2011; Jayananda et al., 2015; Moyen, 2011; Moyen and Martin, 2012; Smithies
409 et al., 2009). Melting of this mafic source at medium to high-pressures, with the involvement of
410 residual garnet, is the mechanism dominantly proposed to account for the TTG's typical HREE
411 depletion, correlated with high Sr/Y ratios (Moyen, 2011; Nagel et al., 2012). In comparison, granite
412 requires a K-rich source and are generally thought to derive from the melting of felsic lithologies (e.g.
413 Laurent et al., 2014; Moyen, 2011; Moyen and Laurent, 2018).

414 Given the high-grade metamorphic conditions of the SHC, our interpretation is focused toward the
415 least mobile elements, although the SHC granitoids appear to have relatively well-preserved whole-
416 rock geochemical compositions, despite metamorphism reaching up to granulite facies. Granitoids
417 from the SHC display variable geochemical compositions, but the samples analysed here are consistent
418 with geochemical compositions of SHC felsic rocks from the literature (Bridgwater and Collerson,

419 1976; Schiøtte et al., 1993, 1989a) and can be divided into four main rock types including
420 trondhjemite, Mg-rich tonalite, granodiorite and granite. Table 3 and Figure 10 present the main
421 compositional characteristics of each granitoid groups. While there is no systematic relationship
422 between the age of the SCH granitoids and their whole-rock chemical compositions, there appears to
423 be a general temporal evolution of the SHC felsic rocks composition. The TTG are more commonly
424 found in the Eoarchean and the Paleoproterozoic, whereas most granite samples are often Neoproterozoic
425 (Table 2). This broad compositional secular evolution is observed in most Archean cratons (Laurent et
426 al., 2014).

427 In general, the SHC granitoids exhibit a wide range of HREE depletion with most rocks exhibiting
428 high $(La/Yb)_N$ ratios typical of Archean TTG ($La/Yb_N > 15$; Fig. 4 and 5). The HREE depletion in TTG
429 is generally attributed to the presence of residual garnet in the source, which seems to be mainly
430 controlled by the pressure of melting (e.g. Moyen, 2011; Nagel et al., 2012). Most SHC trondhjemite
431 samples exhibit a more pronounced HREE depletion compared to the Mg-rich tonalite and
432 granodiorite (Fig. 4 and 5), which would suggest that they were produced from melting at higher
433 pressures. This is also supported by the higher Sr/Y ratios of the trondhjemite compared to the Mg-rich
434 tonalite consistent with formation at lower pressures (Fig. S5).

435 Several trondhjemite samples exhibit pronounced positive Eu anomalies that appear to correlate with
436 La/Yb and Zr/Nd ratios (Fig. 11). Different petrogenetic processes have been proposed to explain
437 positive Eu anomalies in Archean granitoids such as accumulation of feldspars, fractionation of small
438 amounts of allanite or residual rocks from partial melting (Condie et al., 1985; Martin, 1987; Rudnick,
439 1992). Given the complex reworking history displayed by some SHC granitoids, supported by the high
440 discordance of some Eoarchean zircon (Fig. 7b-c-d & supplementary material Table S3) and the
441 relative abundance of secondary sector zoned zircon in the trondhjemite, it is possible that the

442 trondhjemite which exhibit positive Eu anomalies represent residual rocks after some extent of partial
443 melting. All trondhjemitic samples analyzed here for U-Pb include at least a small proportion of
444 Neoproterozoic sector zoned igneous zircon, despite their Eoarchean crystallization ages. The main zircon
445 population of sample SG-019, which displays the largest Eu anomaly, is Neoproterozoic with only 2 older
446 zircon cores preserved. Field observations support the relatively high migmatitisation of the Saglek-
447 Hebron Complex crust, which exhibits important, if not systematic, leucocratic veins of granitic melt
448 (see supplementary material Fig. S1). Contrastingly, the Mg-rich tonalite and granodiorite do not
449 exhibit such positive Eu anomalies and show higher REE contents, suggesting that their whole-rock
450 geochemical composition may not have been affected by remelting event(s) to the same extent as the
451 trondhjemitic rocks. The SHC granodiorite have a REE composition similar to modern granite as
452 opposed to the high La/Yb ratios typically displayed by TTG (Fig. 5).

453 Interestingly, the SHC granitic rocks show a wide range of HREE compositions (Fig. 4) and La/Yb
454 ratios similar to, or much higher than the TTG, as opposed to typical modern granitic rocks (Fig. 5).
455 This would be expected if they were mostly produced from re-melting of a precursor with a wide
456 compositional range and high La/Yb ratios, such as the older SHC trondhjemite and Mg-rich tonalite.
457 Most granitic samples exhibit negative to slightly positive Eu anomalies (Fig. 4b and 11) consistent
458 with this petrogenetic interpretation. The leucogranite samples SG-007 and SG-017 are the two
459 exceptions that exhibit pronounced positive Eu anomalies with very low REE concentrations (Fig. 4b
460 and 11) and most likely represent restites from the melting of older crust. This is also consistent with
461 their zircon populations. SG-017 includes mainly low Th/U Neoproterozoic zircon with a single inherited
462 Paleoproterozoic zircon (Fig. 7k), whereas sample SG-007 displays a complex zircon population with
463 several Eoarchean zircon that show evidence of important Pb-loss between 3800 and 3500 Ma and

464 abundant Neoproterozoic re-crystallized zircon (Fig. 7j). This is indicative of a complex protracted crustal
465 reworking history consistent with the restitic compositions of the leucogranite samples.

466 Figure 12 can be used to determine the potential source for felsic melts and therefore, the precursors of
467 Archean granitoids. The SHC TTG appear to result from the melting of a mafic source characterized
468 by variable K contents. The SHC granitic rocks, on the other hand, are more consistent with a felsic
469 crustal source. The older Eoarchean and Paleoproterozoic tonalite and trondhjemite are the likely crustal
470 precursor sources to the Neoproterozoic granitic rocks, which would explain the wide range of REE
471 compositions of the granite with La/Yb ratios similar to both the Mg-rich tonalite and the trondhjemite.
472 Several granite samples exhibit negative Eu anomalies which could be complementary with the
473 positive anomalies found in the trondhjemitic (perhaps restitic) rocks. The ternary diagram shown in
474 figure 12b was proposed by Laurent et al. (2014) to highlight the relative end-member petrogenetic
475 processes involved in the formation of Archean granitoids. A high $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratio indicates the
476 melting of mafic rocks producing K-poor granitoids, a high A/CNK ratio is consistent with an Al-rich
477 source such as metasediments, and the FMSB [FSBM = $(\text{FeO}_t + \text{MgO}) \times (\text{Sr} + \text{Ba})$ wt. %] value is
478 indicative of the interaction with a metasomatized mantle. Applied to the SHC granitoids, this diagram
479 further supports the hypothesis that the likely source of the tonalite and trondhjemite is a mafic crust,
480 whereas the granite is more consistent with derivation from the melting of Al-rich felsic crust. Some
481 Mg-rich tonalite and granodiorite samples plot towards the FMSB end member, suggesting that they
482 may have interacted with, or included, a mantle component. Similar geochemical characteristics are
483 observed in Archean sanukitoid (Laurent et al., 2014) originally defined as diorite to granodiorite with
484 high Mg# (>0.6), Ni (>100 ppm) and Cr contents (200-500 ppm), with variable TiO_2 contents and
485 relatively high K, Sr, Zr and Nb concentrations (Shirey and Hanson, 1984). Despite the geochemical
486 signature that could be supportive of interaction with the mantle and the relatively high abundance of

487 Cr-Ni displayed by the SHC Mg-rich tonalite, it still does not reach the typical enrichment in Ni-Cr-Sr
488 found in typical sanukitoid (Heilimo et al., 2010; Martin et al., 2009; Martin and Moyen, 2005). The
489 SHC tonalite, nevertheless, exhibits high Mg, Fe, Cr, Ti and V concentrations relative to the
490 trondhjemite, but perhaps not quite comparable to typical Neoproterozoic sanukitoid.

491 **5.2 Geochronology of the SHC granitoids**

492 Early work on the SHC granitoids associated the main episodes of felsic magmatism with defined rock
493 compositions. For example, the Uivak I gneiss was described as ~3700-3800 Ma tonalite, whereas the
494 Uivak II gneiss was defined as ~3600 Ma granodiorite (Baadsgaard et al., 1979; Bridgwater and
495 Schiøtte, 1991; Nutman and Collerson, 1991). It has become clear with the more recent work,
496 however, that specific geochemical compositions are not associated with specific ages. This is also
497 evident from the new dataset we present in this study. While Mesoproterozoic and Neoproterozoic felsic rocks
498 are mostly composed of granite (with perhaps subordinate trondhjemite), the Paleo- to Eoproterozoic
499 granitoids in the SHC include trondhjemitic, tonalitic, granodioritic and granitic rocks, (Table 2).
500 Therefore, we propose here that the terminology for the different SHC units (e.g. "Iqaluk" or "Uivak
501 I", etc...) refers to distinct temporal magmatic events, regardless of geochemical compositions, we will
502 use names such as: "Uivak I Mg-rich tonalite", "Uivak I trondhjemite" or "Iqaluk Mg-rich tonalite", in
503 order to refer to both the age and respective geochemical composition.

504 To get a statistical overview of the age distribution of the felsic magmatism in the SHC, we have used
505 Kernel density estimation diagrams (KDE; Fig. 13a-b) to highlight the probability of a zircon to be
506 either metamorphic or magmatic, based on the textural and chemical characteristics of each zircon, at a
507 given time. Probabilities from KDE, provide an overview of maximum magmatic production, thermal
508 recrystallization and relative timing between different maximums, as well as show whether magmatic
509 production is punctual (sharp Gaussian peak) or diffused in time (asymmetrical peak or large

510 wavelength Gaussian distribution). It should also be noted that sample bias or missing data can be a
511 caveat for interpretation of these KDE.

512 **5.2.1 Magmatic history**

513 The SHC is one of the rare geological terrains on Earth preserving Eoarchean rocks. Precise and
514 accurate age determination of Earth's oldest rocks has important implications, as it brings timing
515 constraints on various geological processes to understand the early Earth (e.g. Whitehouse et al.,
516 2019). Shimojo et al. (2016) proposed age of 3920 ± 49 Ma for a banded grey gneiss sample
517 (LAA995) they interpreted as from the Iqaluk gneiss. This would represent the second oldest
518 occurrence of felsic crust on Earth, after the Acasta gneiss (Bowring and Williams, 1999; Reimink et
519 al., 2016). However, over 300 spots from 231 zircon were analyzed by LA-ICP-MS for this sample
520 and only the 6 oldest zircon were used to define this age, of which 5 yield over-concordant ages.
521 Whitehouse et al. (2019) recently questioned the exactitude of this age and pointed out that a larger
522 subset of analyses from this granitoids yields a statistically significant population with a mean
523 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3865 ± 4 Ma. Sample SG-210c (this study) is the same banded grey gneiss, collected
524 on the same outcrop as sample LAA995 (supplementary material Fig. S4). The Concordia age we
525 obtained for the 34 oldest concordant zircon from sample SG-210c is 3869 ± 6 Ma (Fig. 7g). When we
526 apply the same data filtering method used for SG-210c to select a subset of analyses from sample
527 LAA995 ($\pm 2\%$ concordance; Th/U > 0.3 ; $^{207}\text{Pb}/^{206}\text{Pb}$ age older than 3800 Ma), it defines a consistent
528 concordant population with a well-defined Gaussian distribution around a Concordia age of
529 3869 ± 5 Ma (MSWD=1.1 $n = 181$). Although it still represents one of the oldest rocks on Earth, the
530 oldest Iqaluk granitoids unit is thus more consistent with a < 3900 Ma age, which was displayed by
531 other granitoids samples from the SHC, such as 3860 ± 10 Ma (Vezeinet et al., 2018); 3851 ± 46 Ma,

532 3829 ± 27 Ma, 3869 ± 31 Ma, 3895 ± 33 Ma and 3897 ± 33 Ma (Komiya et al., 2017); 3849 ± 260 Ma
533 (Collerson, 1979); 3863 ± 12 Ma (Schiøtte et al., 1989a).

534 Figure 13a displays the probability density for concordant zircon ($\pm 2\%$) that exhibit Th/U ratios >0.3
535 and igneous textures (sector/oscillatory zoned), interpreted to represent the probability of
536 crystallization ages. The limit of 0.3 for Th/U ratios was set based on the correlation between the rock
537 crystallization age and the zircon's Th/U ratio. Within our dataset, the zircon grains that are closer in
538 age to the interpreted crystallization age of their host rocks, exhibit higher Th/U ratios that are
539 generally above 0.3. The Concordia age deconvolution analysis shows that three distinct probability
540 peaks of zircon production are recorded during the Eoarchean and early Paleoproterozoic, at 3857 Ma,
541 3744 Ma and 3575 Ma (relative misfit of 0.131). In contrast to the proposed protracted and continuous
542 magmatic activity proposed by Komiya et al. (2017), we suggest that the magmatic activity is most
543 likely represented by three discrete magmatic pulses during the Eoarchean, which would correspond to
544 the units previously defined as the Iqaluk gneiss, the Uivak I gneiss and the Uivak II gneiss.

545 Between 3400 and 3200 Ma, two distinct generations of granitoids intruded the SHC, supported by the
546 Concordia age of 3224 ± 7 Ma for the Lister gneiss sample SG-265 and the ages of 3330 ± 15 Ma and
547 3330 ± 7 Ma, obtained respectively from samples SG-203 and SG-208. The 3224 and 3330 Ma
548 samples have distinct initial ϵ_{Hf} values of +1 and -6, and we thus interpret these felsic magmas to be
549 derived from distinct sources produced from two separate magmatic events. We therefore define the
550 Lister gneiss unit as being emplaced at ~ 3220 Ma and propose a distinct ~ 3330 Ma magmatic event,
551 referred to as the "Iluilik", the local Inuit name for the area described as the "opposite coast of Nulliak
552 Island" by Komiya et al. (2017, 2015), where the 3330 Ma samples were collected. Although
553 recognizably distinct, the Iluilik magmatic event appears to be a minor component of the SHC mostly
554 located around the Nulliak Island and its Iluilik opposite coast.

555 A prominent Neoproterozoic event is dominated by granitic intrusions and defines a single magmatic peak
556 with an age of 2750 Ma (Fig. 13a). This suggests that the Neoproterozoic thermal event generally
557 associated with peak metamorphism in the SHC (e.g. Komiya et al., 2017; Kusiak et al., 2018; Nutman
558 and Collerson, 1991; Schiøtte et al., 1989b; Van Kranendonk, 1990), was also accompanied by an
559 important magmatic activity.

560 **5.2.2. Metamorphic history**

561 The SHC has been subjected to a complex thermal history with prevalent metamorphism between
562 2700-2800 Ma, but later high-temperature events have also been recognized (Komiya et al., 2017;
563 Kusiak et al., 2018; Sałacińska et al., 2019, 2018; Schiøtte et al., 1992; Van Kranendonk, 1990).
564 Figure 13b shows the KDE diagram of the probability density ages for recrystallized zircon grains and
565 those exhibiting evidence of age resetting. Only zircon grains yielding concordant ($\pm 2\%$) ages, and
566 thus fully reset ages, are included on Figure 13b. Zircon grains which exhibit low Th/U ratios (>0.3)
567 and homogeneously black texture or heterogeneous patterns (metamict) have also been used as criteria
568 for metamorphic zircon. Eoarchean zircon grains consistent with a metamorphic recrystallization
569 exhibit only one major probability peak that appears to be contemporaneous, if not slightly later, to the
570 ~ 3600 Ma Uivak II magmatic event. This would be consistent with a thermal event described in
571 previous work (Bridgwater et al., 1975; Collerson, 1983b; Sałacińska et al., 2018; Van Kranendonk,
572 1990). These concurrent igneous and metamorphic events are supported by the presence of 3600 Ma
573 metamorphic recrystallization rims surrounding pristine igneous oscillatory zoned 3860 Ma old zircon
574 cores (e.g. SG-210c; Fig. 6c & Fig. 7g; supplementary material Table S3), as well as the presence of
575 inherited 3805 Ma igneous cores surrounded by a Uivak II oscillatory zoned crystallization (e.g. SG-
576 080; Fig 6e & Fig 7m; supplementary material Table S3). An older subordinate Eoarchean population

577 of metamorphic recrystallized zircon grains is seen at ~3750 Ma, but we suggest this could result from
578 ancient Pb-loss in zircon from the Iqaluk gneiss rather than reflecting a distinct metamorphic event.
579 Two minor metamorphic peaks can be observed at 3250 Ma and ~3000 Ma (Fig. 13b), but these late
580 Paleo- to Mesoproterozoic metamorphic events remain equivocal due to the lack of resolution of the KDE
581 diagram and the relative rarity of these populations.

582 The highest probability peak of recrystallization occurs during the Neoproterozoic (Fig. 13b), which is
583 consistent with the extensive metamorphic event previously suggested at 2700 Ma (e.g. Kusiak et al.,
584 2018; Schiøtte et al., 1989b). However, our maximum of recrystallized zircon ages appears to occur
585 closer to 2800 Ma, perhaps shortly before the 2750 Ma Neoproterozoic peak defined by igneous zircon
586 grains (Fig. 13a). This may suggest a slight delay (~50 Ma) between the maximum intensity of the
587 thermal event and the maximum probability of granitic magmatism, which may correspond to crustal
588 anatexis. The overall distribution of the Neoproterozoic recrystallized zircon shows an asymmetric peak
589 shape that reaches its maximum at 2800 Ma, before gradually decreasing to lower probabilities until
590 2200 Ma, after which no zircon is produced in the SHC. This asymmetric peak could be an artifact
591 from our selection criteria used to build the KDE diagram, which could comprise zircon grains that
592 experienced incomplete Pb-loss. However, we cannot disregard the possibility that it suggests
593 protracted high-grade metamorphism before gradually decreasing until 2400-2200 Ma. The apparent
594 asynchrony between the igneous and metamorphic zircon ages may therefore be due to some
595 metamorphic zircon grains that have not been fully reset. The zircon grains with ages <2700 show high
596 U concentrations (Fig. 13b top right inset) and could be more susceptible to reopening of the U-Pb
597 system due to the higher rate of U decay causing the breakdown the zircon crystal lattice (Lee et al.,
598 1997). Therefore, a protracted thermal event that gradually decreased in intensity would keep the U-
599 rich zircon in isotopically open conditions longer than those with lower U contents. The latter would

600 then record older ages compared to the more U-rich zircon grains. This protracted thermal event is also
601 in agreement with what was suggested by Kusiak et al. (2018) based on younger ages obtained on
602 monazite and apatite in the SHC of 2600-2500 and 2200 Ma respectively.

603 **5.3. Crustal sources and reworking history**

604 The U-Pb-Hf isotopic composition of the Eoarchean and Hadean Jack Hills detrital zircons provided
605 invaluable information about the early crust (e.g. Amelin et al., 1999; Blichert-Toft and Albarède,
606 2008; Harrison, 2005; Kemp et al., 2010), but the fact that their host rocks have been eroded away
607 limits the constraints we can put on the earliest crustal history. Possible ancient Pb-loss in detrital
608 zircons can also result in younger apparent $^{207}\text{Pb}/^{206}\text{Pb}$ ages and consequently bias the calculated initial
609 Hf isotopic composition (e.g. Amelin et al., 2000; Vervoort and Kemp, 2016). The SHC zircon grains
610 studied here are all from rock samples with a single interpreted crystallization age. As shown on
611 Figure 9, to avoid the possible calculation of the Hf isotopic compositions at an incorrect apparent Pb-
612 Pb age, we are considering the average initial ϵHf value for all magmatic zircon grains from each
613 granitoids, calculated at the crystallization age of their respective host rock (Table 2, Fig. 14). The
614 composition we attribute to the precursor crustal source of the rocks hosting the zircon grains has an
615 important implication on how the same dataset can be interpreted regarding crustal history. For
616 example, a 3.5 Ga zircon with an initial ϵHf value of -5 could be produced from the remelting of a 3.8
617 Ga felsic crustal precursor or a 4.3 Ga mafic crustal precursor, assuming $^{176}\text{Lu}/^{177}\text{Hf}$ ratios of 0.01 and
618 0.025 respectively, and a chondritic initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratio for both sources. To better constrain the
619 Lu/Hf ratio for the crustal precursor(s) of the SHC granitoids, we use here the whole-rock geochemical
620 composition of each of the host rock samples as a proxy for the composition of their crustal sources
621 and evolution. As discussed previously, we interpret SHC TTG to be derived from the melting of a

622 mafic precursor, while the granitic rocks are consistent with the melting of a felsic crustal source (Fig.
623 12a).

624 The Eoarchean Iqaluk and Uivak I gneiss exhibit slight positive initial ϵ_{Hf} values from +1.7 to +3.7,
625 between 3700 and 3900 Ma (Fig. 14), suggesting that the precursor source of these granitoids had a
626 suprachondritic Hf isotopic composition in the Eoarchean. This contrasts with zircon from most other
627 Eoarchean TTG, generally displaying chondritic or subchondritic initial ϵ_{Hf} values (e.g. Guitreau et al.
628 2012; Iizuka et al., 2009; Næraa et al., 2012; O'Neil et al., 2013; Reimink et al., 2016), leading some
629 authors to suggest that widespread chemical depletion of the mantle, did not take place prior to
630 ~3800 Ma (e.g. Vervoort and Kemp, 2016). Vezinet et al. (2018) recently concluded that a 3860 Ma
631 TTG sample from the SHC included zircon with chondritic initial ϵ_{Hf} values within uncertainty.
632 However, when the ϵ_{Hf} values for each zircon analysed are all calculated at the crystallization age of
633 the host rock of 3860 Ma, such as the approach from study, it yields an average initial ϵ_{Hf} value of
634 $+1.6 \pm 0.2$ (2 S.E.). This is consistent with the slight suprachondritic Hf isotopic compositions we have
635 obtained for all SHC Eoarchean granitoids (Fig. 14). We, therefore, argue that the oldest granitoids
636 from the SHC were sourced from reservoir characterized by a suprachondritic Lu/Hf ratio. The
637 3820 Ma Iqaluk sample SG-026 yields the highest initial ϵ_{Hf} value of +3.2. To evolve to such positive
638 ϵ_{Hf} values by 3800 Ma, a reservoir formed at 4568 Ma with chondritic $^{176}\text{Hf}/^{177}\text{Hf}$ would need a
639 $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of 0.0435. The other SHC Eoarchean granitoids have suprachondritic initial ϵ_{Hf}
640 values corresponding to time-integrated $^{176}\text{Lu}/^{177}\text{Hf}$ ratios for their source ranging from 0.0393 to
641 0.0412. This would suggest that the crustal precursor source of the SHC Eoarchean granitoids was
642 derived from a mantle with a comparable to slightly higher degree of depletion than the present-day
643 depleted mantle with $^{176}\text{Lu}/^{177}\text{Hf} = 0.03933$ (Blichert-Toft and Puchtel, 2010).

644 The Uivak gneiss has previously been interpreted to result from the melting of the Nulliak mafic
645 supracrustal rocks (Komiya et al., 2017, 2015; Morino et al., 2018; Nutman and Collerson, 1991;
646 Shimojo et al., 2016). A Lu-Hf isochron, mainly including ultramafic rocks interpreted to be from the
647 Nulliak assemblage, yielded an initial ϵHf value of +5.1 at 3794 Ma (Morino et al., 2018) with
648 $\epsilon\text{Hf}_{(3770\text{Ma})}$ for individual samples as high +12.8. No zircon analyses from SHC granitoids have
649 however yielded such high initial ϵHf values (this study; Vezinet et al., 2018). The highly positive ϵHf
650 values of the Nulliak rocks may in part be due to some extent of disturbance of the Lu-Hf isotopic
651 system, as suggested by the high MSWD value of 142 of the Lu-Hf isochron (Morino et al., 2018). It
652 nevertheless suggests that the SHC includes Eoarchean mafic/ultramafic crust with suprachondritic Hf
653 isotopic compositions, which would be a possible crustal source for the Iqaluk and Uivak I TTG. The
654 Iqaluk granitoids, however, appear to predate the Nulliak mafic rocks, which would argue against
655 their derivation from melting of the Nulliak metabasalts. All long-lived isotopic systems used to
656 constrain the age of the Nulliak mafic rocks, such as Sm-Nd, Lu-Hf or Re-Os (Collerson et al., 1991;
657 Ishikawa et al., 2017; Morino et al., 2018, 2017), however, show evidence of some degree of
658 disturbance with large errors on isochron ages (hundreds of million years) and we therefore cannot rule
659 out the Nulliak mafic rocks as the possible source of the Iqaluk gneiss.

660 Although trondhjemite and Mg-rich tonalite are more dominant within the Eoarchean Iqaluk and
661 Uivak I gneiss, tonalitic samples also occur within the ~3600 Ma Uivak II and ~3200 Ma Lister gneiss.
662 The Uivak II tonalitic sample has a slight subchondritic initial ϵHf value of -1.1 at 3632 Ma. While
663 reworking of Eoarchean felsic crust evolving with a low $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of ~0.011 could explain the
664 zircon Hf isotopic composition of this Uivak II tonalite sample (Fig. 14), its whole-rock geochemical
665 composition rather suggests derivation from a mafic crustal source (Fig. 12a). Previous work showed
666 that the mafic amphibolites from the SHC display a wide range of $^{176}\text{Lu}/^{177}\text{Hf}$ [e.g, 0.016 to 0.032

667 (Morino et al., 2018)] but assuming a $^{176}\text{Lu}/^{177}\text{Hf}$ ratio between 0.020 and 0.026 for this mafic
668 precursor, more typical of a basaltic crust (Amelin et al., 2000; Blichert-Toft and Albarède, 2008;
669 Kemp et al., 2010), it would suggest a Hadean age between ~4000 and ~4200 Ma for the crustal source
670 of the Uivak II tonalite, if this mafic crust was derived from a long-term depleted mantle-like reservoir
671 (Fig. 14).

672 The initial ϵHf value of +1.0 for the 3224 Ma Lister tonalitic gneiss contrasts with the lower initial ϵHf
673 value of ~-6 for the 3330 Ma Iluilik granitoids, suggesting that although these granitoids units were
674 produced only ~100 Ma apart, they were derived from sources with distinct early histories. If we
675 consider a mafic composition for the crustal precursor of the tonalitic Lister gneiss (Fig. 12a), it would
676 suggest a ~3500 Ma to ~3600 Ma mafic crustal source (Fig. 14). Morino et al. (2018, 2017) proposed
677 the occurrence of Paleoproterozoic mafic rocks in the SHC, but their chondritic $^{176}\text{Hf}/^{177}\text{Hf}$ composition
678 at ~3400 Ma suggest it would not be a suitable precursor source for the Lister gneiss. Other potential
679 mafic crustal sources for the Lister tonalite would include the ~3800 Ma Nulliak mafic supracrustal
680 rocks or perhaps the Paleoproterozoic Saglek dikes (Baadsgaard et al., 1979). Alternatively, the Lister
681 gneiss could be derived from the melting of older mafic crust, similar to the source of the Iqaluk-Uivak
682 I or Uivak II, but with contribution of juvenile material or interaction with the depleted mantle. The
683 Lister gneiss, however, does not exhibit the geochemical features that are typically interpreted as
684 reflecting an interaction between TTG melts and the mantle, such as high Cr and Ni (e.g. Moyen and
685 Martin, 2012) nor evidence for a high component of metasomatized mantle (Fig 12b). Regardless of
686 the exact source of the Lister gneiss, its zircon Hf isotopic composition denotes the contribution of a
687 more juvenile component at the end of the Paleoproterozoic.

688 Both the granite and granodiorite ~3300 Ma Iluilik samples yielded comparable initial ϵHf values of ~-
689 6. While the granitic sample could be produced by reworking of the older SHC Eoarchean felsic crust

690 (Fig. 14), the whole-rock geochemical composition of the granodiorite would be more consistent with
691 a high-K mafic crustal source (Fig 12a). A mafic crustal reservoir would need to have been formed
692 before ~4.1 Ga to evolve to an initial ϵHf value of -6 at 3300 Ma (Fig. 14). No evidence of Hadean
693 mafic crust has been observed in the SHC, but the existence of Hadean hydrothermally altered
694 enriched mafic crust has been proposed as the source of the Jack Hills detrital zircons (Kemp et al.,
695 2010), the Acasta gneiss (e.g. Reimink et al. 2016, 2019) and the Nuvvuagittuq TTG (O'Neil et al.,
696 2013; O'Neil and Carlson, 2017).

697 Another scenario than melting Hadean mafic crust that could possibly explain the low ϵHf values of
698 the zircon grains from the ~3600 Ma Uivak II and ~3300 Ma Iluilik TTG, would be derivation from a
699 mixed crustal source. One could imagine that by the Paleoproterozoic, the SHC crust was a mixture of
700 mafic Nulliak metavolcanic-type rocks and felsic Iqaluk-Uivak I tonalite-trondhjemite. The
701 contribution of an older felsic crustal component in the source of the Paleo- to Mesoarchean TTG
702 could result in lower initial ϵHf values. Assuming that both mafic and felsic end-member sources were
703 formed at 3880 Ma, with respective Hf concentrations of 1.1 and 3.5 ppm [average Hf concentrations
704 for the Nulliak mafic rocks (Wasilewski et al., 2019) and SHC tonalite-trondhjemite (Table 1)] and
705 evolved with respective $^{176}\text{Lu}/^{177}\text{Hf}$ ratios of 0.025 and 0.01, it would, require that the felsic
706 component represent nearly 70% of the mixed crustal source to explain the zircon Hf isotopic
707 composition of the 3630 Ma Uivak II tonalite. Even if a lower $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of 0.001
708 [corresponding to the lowest Lu/Hf ratio measured in the SHC trondhjemite (Table 1)] is considered
709 for the felsic source, more than 50% felsic component is needed to account for the ϵHf value of the
710 Uivak II tonalite. This is however inconsistent with its whole-rock geochemical composition which
711 infers a dominantly mafic source (Fig. 12). Mixing older SHC felsic rocks in the source of the 3300
712 Ma Iluilik granodiorite to account for its low zircon initial ϵHf value is also unlikely as the Iluilik

713 rocks have the highest Hf concentrations (≥ 6.8 ppm) of all SHC felsic rocks and therefore, felsic crust
714 assimilation or mixing would only have a limited effect on its Hf isotopic composition. Except for
715 Neoproterozoic sample SG-087, all other SHC granite samples define an initial ϵ_{Hf} vs. time trend
716 consistent with the reworking of an Eoarchean felsic reservoir evolving with a $^{176}\text{Lu}/^{177}\text{Hf}$ of ~ 0.011
717 that originates from the Iqaluk/Uivak I gneiss (Fig. 14). Most inherited zircon grains are found in
718 granitic samples and follow the same felsic reservoir trend (Fig. 14), which further supports the
719 hypothesis that the granite is produced from the melting of the Eoarchean felsic crust. The combined
720 whole-rock geochemical and Hf isotopic compositions of the granitic rocks of Iqaluk, Uivak I, Uivak
721 II, Ilulik and Neoproterozoic ages, therefore suggest that they were produced from the remelting of the
722 SHC Eoarchean tonalite/trondhjemite over ~ 1 billion years. The ~ 3000 Ma granitic sample SG-087,
723 however, would be consistent with the reworking of the ~ 3200 Ma Lister gneiss rather than the
724 Eoarchean TTG. Except perhaps for the trondhjemite sample SG-019 with an equivocal crystallization
725 age, all Neoproterozoic granitoids analyzed here are granitic in composition with low initial zircon ϵ_{Hf}
726 values. This suggests that the ~ 2750 Ma felsic magmatism is dominated by remelting older felsic crust,
727 with little to no input from a more juvenile source. The major Neoproterozoic high-grade metamorphic
728 event that affected the SHC, therefore, appears to also have been accompanied by important crustal
729 reworking and anatexis.

730 Given that the SHC is part of the North Atlantic Craton, it is often compared to the Itsaq Gneiss
731 Complex of Southwest Greenland, also comprising Eoarchean TTG and supracrustal rocks
732 (Bridgwater et al., 1990; Collerson, 1983a; McGregor, 1973; Morino et al., 2017; Næraa et al., 2012;
733 Wasilewski et al., 2019). Figure 14 shows the evolution of Hf isotopic compositions for detrital and
734 igneous zircon from southwest Greenland (Næraa et al., 2012; 2014) compared to the SHC igneous
735 zircon. The Eoarchean zircon grains from Greenland overall display lower initial ϵ_{Hf} values, with

736 mostly chondritic to slightly subchondritic compositions, compared to the SHC where the 3700-
737 3900 Ma zircon grains generally display positive initial ϵ_{Hf} values (Fig. 14). The zircon Hf isotopic
738 compositions from both SW Greenland and the SHC are mostly consistent with reworking of
739 Eoarchean crust until the late-Paleoarchean when it abruptly shifts from low initial ϵ_{Hf} values around
740 3300 Ma to suprachondritic compositions at ~ 3200 Ma (Fig. 14; Hoffmann et al., 2011; Næraa et al.,
741 2012). This was interpreted by Næraa et al. (2012) to represent a change of geodynamic setting in SW
742 Greenland that involved juvenile crust generation by plate tectonic processes. An important input of
743 juvenile crust was recorded throughout the Mesoarchean in SW Greenland with mostly positive initial
744 zircon ϵ_{Hf} values between 3200 and 2800 Ma (Næraa et al., 2012), but the lack of Mesoarchean
745 zircon-bearing rocks in the SHC does not allow to assess if the same processes were involved during
746 that time. Similarly to the SHC Neoproterozoic granitoids, the Neoproterozoic Qôrqut Granite Complex of
747 SW Greenland displays low zircon ϵ_{Hf} values between -12 and -18 at ~ 2550 Ma. However, these are
748 believed to be derived from an Eoarchean mafic source (Næraa et al., 2014) rather than from
749 reworking of older felsic crust, such as what we propose for the SHC Neoproterozoic granitoids.
750 Therefore, the dominant crustal source of the SHC and SW Greenland felsic rocks seems to diverge in
751 the Neoproterozoic.

752 **5.4. Tectonic context of the SHC**

753 One of the most highly debated subjects about the early Earth concerns the tectonic setting operating
754 and responsible for the formation of the Archean cratons (e.g. Bédard, 2016; de Wit, 1986; Grove et
755 al., 2003; Johnson et al., 2013; Moyen and Laurent, 2018; Nutman and Bennett, 2019; Smithies and
756 Champion, 2000; Wiemer et al., 2018). The geochemical and isotopic compositions of ancient rocks
757 have been widely used to constrain the tectonic environments in which they formed, but there is still

758 no consensus on Earth's early geodynamics. Although the compositions of the SHC metavolcanic
759 rocks do not exhibit geochemical signatures typically found in suprasubduction environments
760 (Wasilewski et al., 2019), Komiya et al. (2015) suggested evidence of Eoarchean subduction settings
761 in the SHC. Based on field observations, they describe the stratigraphy of the Nulliak supracrustal
762 assemblage as an analog to the duplex structures observed in the Japanese trench, interpreted as large
763 accretion prisms. In this model, the Uivak I granitoids would be produced from the melting of basaltic
764 crust within an accretionary complex at shallow depths. However, Uivak I and Iqaluk TTG exhibit
765 strong HREE depletion (Fig. 4 and Fig. 5) consistent with the melting of a garnet-bearing precursor at
766 relatively high-pressures, which is at odds with shallow-level melting of a tholeiitic or possibly
767 komatiitic crust. Although U-Pb and Hf isotopes cannot directly be linked to tectonic settings, these
768 data can help to constrain the type of crustal precursor and its reworking history, to better establish the
769 architecture of the reworked crustal sources and evaluate which tectonic context would be more likely.
770 Figure 15 shows the rock or inherited zircon crystallization age *vs.* mantle extraction Hf model age for
771 the source of the SHG TTG. The mantle extraction model ages are obtained using the zircon average
772 initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios for each of the samples, back-calculated to the depleted mantle evolution line,
773 and assuming a $^{176}\text{Lu}/^{177}\text{Hf}$ between 0.020 and 0.026 for a mafic crustal reservoir. Consequently, only
774 the samples consistent with derivation from a mafic crustal precursor (i.e. the TTG) and inherited
775 zircon grains from granitic samples (which likely crystallized within their TTG precursor) are included
776 on Figure 15. Although it is more difficult to assess if the inherited zircon grains were affected by
777 ancient Pb-loss, which could lead to erroneous calculation of their initial Hf isotopic composition, data
778 from inherited grains are consistent with the other samples and follow the same trend on Figure 15.
779 The zircon grains from Eoarchean granitoids are the only samples with relatively juvenile
780 compositions, consistent with the remelting of almost contemporaneous mantle-derived mafic crust,

781 potentially the Nulliak basaltic crust. The younger Uivak II and Iluilik granitoids appear to be derived
782 from the melting of pre-Nulliak Hadean mafic crust. Except for the Lister gneiss (sample SG-265), the
783 SHC granitoids derived from the melting of mafic crust show a negative correlation between their
784 crystallization age and their Hf model age, suggesting that the latest granitoids were produced from the
785 melting of the oldest crustal source. In a modern-style subduction setting, it would be unlikely that: -1)
786 the subducting mafic crustal reservoir displayed such a wide age range (from ~3900 to ~4400 Ma) and
787 -2) the oldest portions of the subducting mafic crust melted the latest. While one could suggest that a
788 thick mafic crust, similar to an oceanic plateau, could include mafic crust with the range of ages
789 suggested by the Hf model ages from Figure 15, it would be expected that the base of this thickened
790 mafic crust, *i.e.* the oldest portions, melts before the younger overlying portions. This is inconsistent
791 with the negative trend observed in Figure 15. Some authors have proposed a “subcretion” model as an
792 alternative to modern-style subduction tectonics (Barr et al., 1999; Ducea et al., 2009; Grove et al.,
793 2003; Hacker et al., 2015; Taramon et al., 2015). Sequential stacking or tectonic imbrication have been
794 suggested as a mechanism for early Archean crustal growth (e.g. de Wit, 1986; Smithies and
795 Champion, 2000) and the formation of the Eoarchean Itsaq Gneiss Complex of Southwest Greenland
796 (Nutman et al., 2007; Nutman and Bennett, 2019). This tectonic regime has the potential to imbricate
797 portions of mafic crust of variable ages in a configuration where older crust could overlie younger
798 crust and perhaps melt later, in a thickened oceanic crust. Reworking of such an imbricated mafic
799 crustal source could explain the crystallization age *vs.* Hf model age trend seen for the SHC TTG.

800 **6 Conclusion**

801 The SHC granitoids record more than 1 billion years of complex crustal history, including several
802 episodes of felsic magmatism. The different generations of gneiss in the SHC were previously

803 associated with certain geochemical characteristics, but they appear to be more geochemically and
804 petrologically diverse. We, therefore, suggest that the different gneissic units should refer to distinct
805 temporal magmatic events regardless of composition. Six distinct felsic magmatic events can be
806 identified in the SHC: the oldest 3850 Ma Iqaluk gneiss, the 3750 Ma Uivak I gneiss, the 3600 Ma
807 Uivak II gneiss, the 3330 Ma Iluilik gneiss, the 3230 Ma Lister gneiss, and the 2700-2800 Ma
808 Neoproterozoic granitoids. The chondritic to subchondritic Hf compositions that Eoarchean zircons from
809 most other early terrains exhibit has led to the suggestion that the depletion of the mantle through
810 extraction of crustal material prior to 3.8 Ga was negligible (e.g. Vervoort and Kemp, 2016). The
811 slightly suprachondritic Hf isotopic compositions of the Eoarchean zircon from the SHC granitoids,
812 however, denote the existence of a long-term depleted source, more consistent with the evidence for an
813 early depleted mantle as suggested from Nd isotopes (e.g. Bennett et al., 2007, 1993; Boyet and
814 Carlson, 2006; Caro et al., 2003; O'Neil et al., 2016; Rizo et al., 2011). Most of the juvenile
815 continental crust in the SHC appears to have been formed during the Eoarchean. Considering the
816 geochemical composition of the SHC granitoids to constrain the nature of their crustal precursor, the
817 Hf isotopic compositions of the different generations of TTG suggest the remelting of Eoarchean to
818 Hadean mafic crust. The juvenile Iqaluk and Uivak I TTG are consistent with remelting of ~3800-
819 3900 Ma mafic crust, while the younger Uivak II and Iluilik TTG would suggest melting of
820 increasingly older (up to ~4300 Ma) mafic sources. As observed in SW Greenland, the SHC granitoids
821 show a marked transition between ~3300 and 3200 Ma from relatively unradiogenic to more juvenile
822 compositions. The Neoproterozoic felsic magmatism in the SHC, however, appears to be dominated by
823 the reworking of the Eoarchean TTG, without contribution of juvenile material.

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834 **Captions**

835 **Figures:**

836 Figure 1: Simplified geological map of the Saglek-Hebron Complex (SHC) modified from **Ryan and**
837 **Martineau (2012)** and **Komiya et al. (2015)**. Sample locations (yellow circles) show the main localities
838 where multiple samples were collected for this study. Coordinates are in UTM NAD 27 zone 20.

839

840 Figure 2: Ab-An-Or ternary diagram (**Barker, 1979**) for the SHC granitoids. Shaded fields with smaller
841 light colored samples represent SHC granitoids compositions from the literature (**Bridgwater and**
842 **Collerson, 1976; Collerson and Bridgwater, 1979; Schiøtte et al., 1993, 1989b**).

843

844 Figure 3: Selected major element compositions of the SHC granitoids. a), c) and d) selected elements
845 respectively *vs.* MgO b) TiO_2 *vs.* FeO_t e) and f) have MgO + FeO_t on their abscissa showing
846 covariations with CaO (wt. %) and A/CNK (relative aluminum concentration in molar unit calculated
847 as followed $\text{A/CNK} = \text{Al}/(2\text{Ca} + \text{Na} + \text{K})$). Symbols are as in Figure 2.

848

849 Figure 4: a) Primitive mantle normalized trace element diagrams , b) Chondrite normalized rare earth
850 element diagrams for the SHC granitoids (McDonough and Sun, 1995). Shaded fields show the data
851 for SHC granitoids from the literature (Bridgwater and Collerson, 1976; Collerson and Bridgwater,
852 1979; Schiøtte et al., 1993, 1989b). Symbols and colors for fields are as in Figure 2. Normalization
853 values for primitive mantle are from Lyubetskaya and Korenaga (2007) and for chondrite are from
854 McDonough and Sun (1995). Some specific samples discussed in the text are emphasised by thicker
855 lines and are labeled.

856 Figure 5: $(\text{La}/\text{Yb})_N$ *vs.* Yb_N diagram reflecting LREE/HREE ratios of all SHC TTG and granite
857 compared to typical TTG, sanukitoid and modern granite from the literature (Moyen and Martin
858 (2012) and references therein). Normalization values are after Masuda et al. (1973). Symbols are as in
859 Figure 2.

860

861 Figure 6: Cathodoluminescence images of zircon grains with U-Pb (red dashed circles) and Lu-Hf
862 (blue dashed circles) LA-ICP-MS spots. a) Trondhjemite sample SG-024: grain #A05 shows typical
863 sector zoning of zircon that crystallized during the 2700-2800 thermal events. Zircon #A17 exhibiting
864 igneous oscillatory zoning and a younger core age relative to the rim. b) Mg-rich tonalite sample SG-

865 026 from Ukkalek Island: grains #B16 and #E12 show respectively oscillatory and sector zoned cores
866 with discordant black colored rims. c) Mg-rich tonalite sample SG-210c from the Kangidluarsuk fjord:
867 grain #A01 exhibits multiple ages decreasing from the core to the rim, suggesting the spread of zircon
868 ages observed on the Concordia diagram can be explained by ancient Pb-loss. Zircon #A02 is a
869 discordant metamict U-rich grain. d) Granite sample SG-007 from Ukkalek Island: grain #C05 yields
870 an Eoarchean magmatic age and grain #C06 from the same sample is a black U-rich metamict zircon.
871 e) Granite sample SG-080 from the Kangidluarsuk fjord: grain #C15 is a good example of an
872 Eoarchean inherited zircon within an early-Paleoarchean oscillatory zoned crystal. f) Granite sample
873 SG-143: U-rich zircon #B11 shows oscillatory zoning and zircon B12 is a metamict crystal typically
874 found in Neoproterozoic rocks.

875

876 Figure 7: Concordia diagrams of trondhjemite (a-b-c-d), Mg-rich tonalite (e-f-g-h), granodiorite (i) and
877 granite (j-k-l-m-n-o-p-q-r) samples from the SHC. All diagrams show interpreted crystallization ages,
878 inherited grains, and metamorphic secondary populations. Colors for the ellipses reflect the different
879 age populations which are classified from the criteria described in the results (section 2.4); green:
880 inherited zircon; red: primary igneous zircon; orange: secondary igneous zircon; yellow: metamorphic
881 re-crystallized zircon; pink: zircon used for Concordia ages; light blue: Concordia ages. Dashed empty
882 ellipses show altered zircon yielding analyses that cannot be grouped in a consistent population. These
883 analyses were rejected for any age calculation. Ellipses represent 2σ uncertainties.

884

885 Figure 8: Initial ϵ_{Hf} values vs. $^{207}\text{Pb}/^{206}\text{Pb}$ ages for the analyzed zircon grains. Grey diamonds represent
886 individual analyses, and the colored symbols show the average initial ϵ_{Hf} values calculated at the

887 interpreted crystallization age of the host rocks. Symbols are as in Figure 2. Note that average ϵ_{Hf}
888 values only include relevant analysis for the defined population, such that metamorphic and inherited
889 zircon grains are not included in the average epsilon calculation. A ratio of $^{176}\text{Lu}/^{177}\text{Hf} = 0.03915$ was
890 used for the long-term evolution of the depleted mantle, starting with a chondritic $^{176}\text{Hf}/^{177}\text{Hf}$
891 composition at 4568 Ma. 2 SD errors on the average ϵ_{Hf} values and age for the colored symbols are
892 smaller than the symbols.

893

894 Figure 9: Cartoon showing how average initial ϵ_{Hf} values are calculated in Figure 8, from the
895 interpreted age of crystallization of the rocks, to avoid apparent ϵ_{Hf} vs. $^{207}\text{Pb}/^{206}\text{Pb}$ trends caused by
896 ancient Pb-loss. Panel a) shows a fictitious example of a sample with a crystallization age of 3880 Ma,
897 which was affected by ancient Pb-loss and for which most zircon analyses are still relatively
898 concordant. Panel b) illustrates that for the same fictitious sample, if initial ϵ_{Hf} values are calculated
899 for individual zircon grains at their respective “concordant” ages (yellow circles), it produces a steep
900 ϵ_{Hf} vs. $^{207}\text{Pb}/^{206}\text{Pb}$ age array consistent with the low Lu/Hf of zircon, rather than representative of the
901 evolution of a crustal source. The blue cross symbol shows the average ϵ_{Hf} value for all zircon
902 analyses calculated at the 3880 Ma crystallization age.

903

904 Figure 10: Proposed major element composition discrimination diagram for all granitoids from the
905 SHC. Symbols and colored fields are as in Figure 2.

906

907 Figure 11: Zr/Nd_N vs. Eu_N/Eu^* diagram showing the correlation between Eu anomalies and
908 incompatible trace element ratio (Zr/Nd). Eu is normalized to chondrite, Zr and Nd are normalized to
909 the primitive mantle. Normalization values are from McDonough and Sun (1995). Symbols are as in
910 Figure 2.

911

912 Figure 12: Discrimination ternary diagrams for granitoids proposed by Laurent et al. (2014). a) Major
913 element ternary diagram showing the possible crustal source(s) for the granitoids. The arrow labeled
914 “restitute” shows the suggested restitic nature of some samples from the SHC that exhibit pronounced
915 positive Eu anomalies (e.g. SG-007; SG-017; SG-019). b) Ternary diagram showing the petrogenetic
916 processes involved in the formation of Archean granitoids. Contribution of mantle component in the
917 melt is highlighted by the FMSB value $=[(FeO_t + MgO)wt.\% \times (Sr + Ba)wt.\%]$. Symbols and colored
918 fields are as in Figure 2.

919

920 Figure 13: Kernel density estimate (KDE) and frequency diagrams for zircon grains analyzed for this
921 study. Igneous zircon grains (panel a) and metamorphic zircon grains (panel b) have been
922 discriminated based on CL imaging and Th/U ratios. The top right inset on figure 13b shows a U
923 (ppm) vs. $^{207}Pb/^{206}Pb$ age diagram for all the Neoproterozoic zircon grains that illustrates the increase of
924 the U concentrations in younger zircon grains. KDE analysis was performed under “IsoplotR”
925 (Vermeesch, 2018) software using a combination of the Botev et al. (2010) bandwidth selector and the
926 Abramson (1982) adaptive kernel bandwidth modifier. More detail available at: [https://cran.r-](https://cran.r-project.org/web/packages/IsoplotR/IsoplotR.pdf)
927 [project.org/web/packages/IsoplotR/IsoplotR.pdf](https://cran.r-project.org/web/packages/IsoplotR/IsoplotR.pdf) .

928

929 Figure 14: Average zircon initial ϵ_{Hf} values vs. $^{207}\text{Pb}/^{206}\text{Pb}$ age diagram for the SHC rocks. The green
930 field shows data for the SW Greenland TTG and detrital zircons from [Næraa et al. \(2012\)](#). Data from
931 [Veziñet et al. \(2018\)](#) shows the average ϵ_{Hf} value for all analyzed zircon grains recalculated at the
932 crystallization age of the host TTG. Data from [Morino et al. \(2018\)](#) shows the initial ϵ_{Hf} values from
933 their whole-rock Lu-Hf isochrons. Evolution arrays for mafic sources are shown in purple. Evolution
934 array for an Eoarchean felsic source is shown in pink. Evolution array for a Paleoarchean felsic source
935 starting from the Lister gneiss sample is shown with a dash line. The KDE from Figure 13 for igneous
936 zircon analyses is shown in blue at the bottom of the diagram. A ratio of $^{176}\text{Lu}/^{177}\text{Hf} = 0.03915$ was
937 used for the long-term evolution of the depleted mantle, starting with a chondritic $^{176}\text{Hf}/^{177}\text{Hf}$
938 composition at 4568 Ma. This reference line for the depleted mantle corresponds to a present-day
939 $^{176}\text{Hf}/^{177}\text{Hf}$ of 0.2833 consistent with modern high degree melt MORB average value (Salters and
940 Stracke, 2004) Symbols for the granitoids are as figure 2, except for trondhjemite sample SG-019
941 shown by a yellow diamond given its equivocal Neoproterozoic age. 2 SD errors on the average ϵ_{Hf} values
942 and age are smaller than the symbols.

943

944 Figure 15: Mantle extraction Hf model ages vs. crystallization ages of the SHC granitoids and inherited
945 zircon grains from the SHC granite sample. The vertical bars for each sample represent the variation of
946 model ages using $^{176}\text{Lu}/^{177}\text{Hf}$ ratios of 0.020 and 0.026, with the symbols plotted as the average.

947

948 **Tables:**

949 Table 1: Whole-rock major (wt. %) and trace (ppm) element analysis for the SHC granitoids. GPS
950 coordinates are in UTM NAD 27 zone 20. Major element compositions are recalculated as anhydrous
951 compositions.

952

953 Table 2: Summary of geochronological data and initial ϵ_{Hf} -zircon for the SHC granitoids. Age type =
954 method used to calculate the age. n = number of zircon analyses used in age calculation. Age=
955 interpreted crystallization, inherited or metamorphic age. $\epsilon_{\text{Hf}(i)}$ is the average initial values of ϵ_{Hf} of all
956 zircon analyses calculated at the crystallization age of their host rock. Full dataset can be found in the
957 supplementary material Tables [S3](#) and [S4](#).

958

959 Table 3: Key characteristics discriminating the four compositional types of granitoids from the SHC,
960 based on this study and compositions from the literature ([Bridgwater and Collerson, 1976](#); [Collerson,](#)
961 [1979](#); [Schjøtte et al., 1993, 1989a](#)).

962

963 **Supplementary material:**

964 **FIGURES:**

965 Figure S1 : Field photographs of different rock types and points of interest (rock hammer head is 20 cm and the
966 sledgehammer is 1 meter long). a) Sampling location for sample SG-210c of Iqaluk tonalitic gneiss dated at
967 3869 ± 6 Ma (the frame size is about a meter). Same outcrop as the sample dated at 3920 ± 49 by Shimojo et al.
968 (2016). b) Banded grey gneiss of trondhjemite from the Nulliak Island (SG-227). c) Mg-tonalite (SG-265; SG-

969 266) dated at 3230 Ma from Lister Island crosscut by granitic migmatite. d) Crosscutting relationship observed
970 at the White Point location showing Mg-tonalite (melaosome e.g. SG-272) of supposedly Lister in age (3200
971 Ma) crosscut by migmatitic melts (leucosome e.g. SG-271), both crosscut by a mafic dike. e) Typical Iluilik
972 augen granodiorite (SG-203; SG-204) described in the literature as the Uivak II (Bridgwater and Schiøtte, 1991;
973 Hurst et al., 1975). f) Iluilik banded granodiorite that exhibits sheared plagioclase and migmatitic veins. g)
974 Trondhjemite observed on Big Island infiltrated by migmatites (SG-260). h) Similar grey gneiss to that in shown
975 in "g", with higher content of Fe and magnetic minerals (SG-258). i) Migmatitized trondhjemite SG-024 found
976 on Ukkalek Island, only melanosome was analysed. j) Migmatitized Mg-rich tonalite (SG-026) found on
977 Ukkalek Island, only melanosome was analysed. k) Iluilik banded granodiorite that exhibits migmatitic veins.
978 Meter scale veins have been analyzed consisting in the sample SG-208 and SG-209.

979 Figure S2 : 15 representative photomicrographs of SHC granitoids. a) cross-polarized photomicrograph of the
980 SG-134 granite and plane-polarized pair in "b". c) cross-polarized photomicrograph of a garnet grain in the SG-
981 127 granite and plane-polarized pair in "d". e) cross-polarized photomicrograph of a SG-122 trondhjemite and
982 plane-polarized pair in "f". g- h) Good example of sagenitic texture in SG-024 biotite formed by secondary
983 exsolution of titanium oxides in the crystal lattices seen in plane-polarized light. i) Indicator of alteration and
984 high deformation in the feldspar found in the SG-024 trondhjemitic sample. j) Highly deformed quartz in granite
985 SG-080 from the Kangidluasuk inlet. k) Cross-polarized photomicrograph of a SG-026 showing the presence of
986 clinopyroxene and plane-polarized pair in "l". m-n) Cross-polarized photomicrograph of a SG-027 which
987 similarly showing the presence of clinopyroxene. o) Cross-polarized SG-027 shows sub-grain formation in
988 plagioclase. Qtz = quartz, Bio = biotite, Apt = apatite, Grt = garnet, Pl = plagioclase, Ep = epidote, Or =
989 Orthose, Zr = zircon, Cpx = clinopyroxene, Amph = amphibole.

990 Figure S.3: Cathodoluminescence images of all analyzed zircon with U-Pb and Hf analyses laser
991 ablation spots.

992 Figure S.4: Schematic of the outcrop from Shimojo et al. (2016) where the Iqaluk gneiss has been described
993 and dated at 3920 Ma. The precise location of sample SG-210c dated at 3869 Ma (this study) and the Sample
994 LAA995 (Shimojo et al., 2016) of Iqaluk gneiss are shown respectively in red and green. The top right inset is a
995 photograph of the drawn outcrop.

996 Figure S.5: Sr/Y vs. La/Yb diagram for the SHC granitoids with the fields for TTG produced by
997 variable melting pressures (Moyen and Martin, 2012).

998 TABLES:

999 Table S.1: Detailed summary of the analytic procedure and conditions for the in-situ U-Pb geochronological
1000 analyses and the in-situ Lu-Hf isotopic analyses. Primary and secondary standard reproducibility of U-Pb
1001 analysis. The primary standard is the GJ-1 and the secondary is the 91500

1002 Table S.2: Full set of U-Pb analysis on zircon primary standard GJ-1 and secondary standard 91500. ρ is the
1003 error correlation coefficient.

1004 Table S.3: Full set of U-Pb analysis on zircon from the 18 samples analyzed here. Abbreviations:
1005 Osci.=Oscillatory; Inh= Inherited; Ext.= External; Int.= Internal; Meta.= Metamict; Z.= Zoned; Conc =
1006 concordance. In the internal structure column when the rim or core is not mentioned, the mentioned
1007 characteristic describes the whole grain. ρ is the error correlation coefficient.

1008 Table S.4: Full data set for the in-situ Lu-Hf isotopic analyses of zircon.

1009

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