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

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1. Introduction

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

2. Geological context

The Saglek-Hebron Complex, located on the east coast of Northern Labrador in Canada, is dominated by orthogneisses from the TTG series along with granitic rocks and includes meter to kilometer scale supracrustal enclaves (Fig. 1). Supracrustal rocks mainly include mantle-derived rocks (mafic metavolcanic and ultramafic rocks), and chemical or clastic metasediments (Baadsgaard et al., 1979; Bridgwater et al., 1975; Komiya et al., 2015; Nutman and Collerson, 1991). The metavolcanic rocks are divided into two distinct units, the Upernavik assemblage interpreted to be Mesoarchean and the Nulliak assemblage interpreted to be Eoarchean (Bridgwater and Schiøtte, 1991; Morino et al., 2018, 2017; Nutman et al., 1989; Schiøtte et al., 1992). Both units, however, are compositionally similar and interpreted as a series of tholeiitic basaltic flows displaying some extent of differentiation (Wasilewski et al., 2019). The SHC orthogneisses are divided into four main units including the Iqaluk grey gneiss (sometimes referred to as the Nanok gneiss), the Uivak gneiss, the Lister gneiss, and late granitic intrusions. The oldest TTG unit in the SHC is the Iqaluk grey gneiss dated at ~3900 Ma (Collerson, 1983a; Komiya et al., 2017; Regelous and Collerson, 1996; Shimojo et al., 2016). The Uivak gneiss are the predominant lithology in the SHC, and were originally subdivided into two units based on their age and mineralogy, with the older tonalitic Uivak I dated between 3863 Ma and 3732 Ma (Sałacińska et al., 2018; Schiøtte et al., 1989b; Vezinet et al., 2018) and the younger granodioritic Uivak II dated at ~3600 Ma (Hurst et al., 1975; Komiya et al., 2017; Sałacińska et al., 2018). However, a recent study argued that the Uivak gneiss rather includes five units produced throughout the Eoarchean, from 3890 to 3610 Ma, and thus form almost continuous protracted magmatism occurring over more than 250 million years (Komiya et al., 2017). The younger granitoids include Paleoarchean TTG called the Lister gneiss consisting of granodioritic intrusions emplaced between 3240 Ma and 3350 Ma (Komiya et al., 2017; Schiøtte et al., 1989b) and Neoarchean granitic rocks intruding the SHC with a magmatic peak occurring at ~2766 Ma (Collerson, 1983a, 1983b; Schiøtte et al., 1989b).

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

3. Methods

Forty-seven orthogneiss samples have been collected for this study. All samples have been analysed for whole-rock geochemistry. A subset of eighteen samples have been selected for coupled to U-Pb and Hf in-situ analysis in zircon. These samples have been selected to cover the full compositional range of SHC granitoids and to include all felsic lithologies previously described in the literature. This includes type localities for the Iqaluk gneiss described by Shimojo et al. (2016) in the Kangidluarsuk Inlet (St John's Harbour), the Uivak II gneiss described by Bridgwater et al.(1975) on the opposite coast of Nulliak Island and “White Point” and the Lister gneiss found on Lister Island. The outcrop studied by Shimojo et al. (2016) on which they obtained an age of 3920 ± 49 Ma and interpreted as the oldest rock of the complex was resampled because of its significance and given that its exact age was questioned due the relatively large imprecision (Whitehouse et al., 2019). Specific locations previously studied by Bridgwater et al., (1975) were also targeted given the paucity of geochronological data on the Uivak II and Lister gneiss from these localities. The detailed geochronology on multiple generations of orthogneisses provided the framework to the Hf isotope work and helped targeting the best-preserved zircon grains to analyse for Hf isotopic compositions which are used to better understand the crustal evolution of the SCH over the whole Archean Eon. Figure 1 shows the main sample locations and GPS coordinates for all samples are given in Table 1. Selected outcrop photographs are available in the supplementary material Figure S1). The full methods regarding whole-rock major and trace element geochemistry can be found in the supplementary material (Supplementary Table S1).

Eighteen samples from our set of 47 granitoids were analyzed for both in-situ U-Pb geochronology and Hf isotopes in zircon. Rock samples were crushed using a steel jaw crusher and disk mill and then sieved to collect grain sizes between 250-106 µm. Heavy minerals, such as zircon and metallic oxides, were separated using a water shaking table and methylene iodide heavy liquids. The heavy mineral fractions were then passed through a Frantz magnetic separator to remove the magnetic minerals. Between 100 and 120 zircon grains were then handpicked, mounted on an epoxy resin and polished. Cathodoluminescence (CL) images of the polished zircon grains were taken using the JEOL 6610LV scanning electron microscope (SEM) at the University of Ottawa, to identify the different zircon zones and guide the laser ablation work. Both U-Th-Pb and Hf isotope analyses were undertaken at the Laboratoire Magmas et Volcans (LMV) (Clermont-Ferrand, France). U-Th-Pb isotope analyses were performed on a Thermo Scientific Element XR-ICP-MS, and Hf isotope analyses were conducted on a Thermo Scientific Neptune Plus multicollector ICP-MS, both instruments were coupled with a Resonetics M50E 193 nm excimer laser ablation (LA) system. The analytical method for isotope dating by LA-ICP-MS described in Hurai et al.(2010) and Paquette et al.(2014) were followed and more details on analytic conditions are available in the supplementary material Table S1. Primary (GJ-1) and secondary (91500) zircon reference materials have been measured to ensure the quality of the data over the different analytic sessions and data are found in the supplementary material Table S1 and Table S2. Common Pb was not corrected owing to the large isobaric interference of ^{204}Hg. The ablated zircon zones were carefully chosen in order to avoid any mixed zones and Hf isotopic measurements were performed on the same spots than those chosen for the U-Pb analyses according to analytical techniques described in Moyen et al. (2017) and Paquette et al. (2017). The full Lu-Hf isotope data are available in the supplementary material Table S3. Initial ^{176}Hf/^{177}Hf ratios were calculated using the λ^{176}Lu decay constant of $1.867\text{x}10^{-11}.\text{yr}^{-1}$ of Söderlund et al. (2004) and the CHUR parameters of ^{176}Hf/^{177}Hf = 0.282785 and ^{176}Lu/^{177}Hf = 0.0336 (Bouvier et al., 2008) were used for calculation of εHf values.

4 Results

4.1 Petrography & Geochemistry

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

The TTG display a wide range of silica content (56.6 wt. % to 76.8 wt. %SiO_2), with most samples having below 72 wt. %SiO_2. The tonalite exhibits the lowest SiO_2 contents among the TTG and display high Mg concentrations relative to the other SHC granitoids (Fig. 3a), and therefore are here referred to as Mg-rich tonalite. Except for one sample, the granite shows high and uniform silica contents ranging from 73.0 to 77.5wt. %SiO_2 (Fig. 3a). Both the Mg-rich tonalite and the granodiorite display relatively high FeO_t and TiO_2 contents compared to the granite and the trondhjemite (Fig. 3b).

Compared to all TTG, the granite exhibits higher concentrations in K_2O (from 3 to 6 wt.%; Fig. 3c) and Rb, and lower CaO concentrations (<2 wt. %; Fig. 3e). The tonalite and trondhjemite exhibit higher Na_2O, CaO, and Al_2O_3but lower K_2O compared to the granite and granodiorite (Fig. 3c-d-e). The Mg-rich tonalite, however, exhibits lower A/CNK ratios relative to the trondhjemite, with the granite showing the highest A/CNK ratios (Fig. 3f).

All granitoids exhibit pronounced negative Nb-Ta anomalies (Fig. 4a) along with strong depletion in heavy rare earth elements (HREE) compared to light rare earth elements (LREE, Fig. 4b). Mg-rich tonalite and granodiorite are generally characterized by lower $(La/Yb)_N$ ratios than most trondhjemite and granite samples (Fig. 5). The trondhjemite samples exhibit higher $(La/Yb)_N$ ratios (up to 150) and most samples display positive Eu anomalies (Fig. 4b; Fig. 5). The trondhjemite samples collected on White Point, however, display higher HREE contents and thus lower $(La/Yb)_N$ ratios compared to other trondhjemite samples. The trondhjemite sample SG-019 collected on Ukkalek Island exhibits extremely low trace element and REE concentrations, a prominent positive Eu anomaly and mostly plots outside of the compositional field comprising the other trondhjemite samples (Fig. 4). The Lister gneiss (SG-265) is compositionally similar to the Mg-rich tonalite but exhibits the lowest REE concentrations (Fig. 4b). Granitic samples exhibit high Rb and Pb with lower Sr concentrations, relative to all other TTG. In general, granite samples are more enriched in LREE compared to the trondhjemitic rocks with large variability in $(La/Yb)_N$ ratios ranging from 0-300 (Fig. 5). The granite samples commonly display small positive or negative Eu anomalies while the leucogranitic samples SG-007 and SG-017 exhibit very low REE concentrations with pronounced positive Eu anomalies (Eu_N/Eu^* ~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. Analyse s 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 $^{207}Pb/^{206}Pb$ 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 range of ages, indicative of ancient Pb-loss, the oldest concordant zircon analyses were considered as the most representative of the crystallization age.

a. Trondhjemite

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

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

A total of 128 U-Pb analyses have been conducted on sample SG-024 (Fig. 7b). Zircon grains are dominated by two main populations that exhibit U concentrations <1000 ppm, associated with a wide range of Th/U ratios (0.01-2.83). A number of zircon grains from sample SG-024 are subeuhedral to euhedral and exhibit well defined oscillatory zoning (e.g. zircon #A17: Fig. 6a). The few concordant analyses plot between 3800 and 3870 Ma. However, most grains from this population are discordant and spread along a poorly defined discordia line (Fig. 7b). The two oldest concordant cores yield a Concordia age of 3869 ± 10 Ma ($MSWD_{(Concordance+Equivalence)}$ =1.5; *n* =2) considered as the crystallization age for this sample. A population of Neoarchean zircon is dominated by subeuhedral to rounded zircon grains that show clear sector zoning and exhibits a higher average Th/U ratio of 1.64 (e.g. zircon #A05:Fig. 6a), which altogether yield a weighted mean $^{207}Pb/^{206}Pb$ age of 2721 ± 8 Ma (MSWD= 0.89; *n* =42). A few Neoarchean oscillatory zoned rims and cores with lower Th/U ratios (<0.2) are also present. However, these Neoarchean zircon grains exhibit much more scattered ages and yield a weighted mean $^{207}Pb/^{206}Pb$ age of 2769 ± 19 Ma (MSWD= 4; *n* =28).

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

A total of 105 U-Pb analyses were conducted on sample SG-122 (Fig. 7d). Zircon grains from this trondhjemite sample are dominated by rounded grains with oscillatory zoned textures and homogeneous black grains in CL images. The population of oscillatory zoned zircon scatters along a discordia line with an upper intercept age at 3752 ± 33 Ma (MSWD=13; *n* = 62). Moreover, a number of concordant zircon analyses from this population spread along the Concordia curve between ~3790 and ~3700 Ma, suggestive of ancient Pb-loss. The oldest concordant zircon from this scattered population gives a Concordia age of 3781 ± 12 Ma, representing the crystallization age for this sample. This euhedral zircon grain shows a clear oscillatory zoning and a relatively high Th/U ratio (~0.2). A subordinate population of U-rich rims yields a younger weighted mean ^{207}Pb/^{206}Pb age of 2804 ± 9 Ma (MSWD= 0.81; *n* =28).

b. Mg-rich tonalite

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

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

A total of 138 analyses were done for the sample SG-027, which includes three main zircon populations (Fig. 7f). Zircon grains from the oldest population exhibit oscillatory zoning and igneous Th/U ratios from 0.2 to 0.5. Several zircon gains from this population are concordant, spreading on the Concordia curve between ~3600 and ~3500 Ma, which may reflect ancient Pb-loss. Therefore, the average age of the two oldest concordant zircon analyses of 3632 ± 9 Ma ($MSWD_{(C+E)}$ = 1.3; *n* =2) is taken as the crystallization age. The other zircon populations are Neoarchean, which include two groups characterized by metamict textures and black CL images, commonly occurring as rims around older cores. One Neoarchean population exhibits lower Th/U ratio and yields a weighted mean $^{207}Pb/^{206}Pb$ age of 2785 ± 7 Ma (MSWD= 1.3: *n* =46), while a secondary smaller population with higher Th/U ratios (>0.4) yields a weighted mean $^{207}Pb/^{206}Pb$ age of 2595 ± 20 Ma (MSWD= 0.93; *n* =6).

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

A total of 86 U-Pb analyses were performed on sample SG-265 that was collected on Lister Island. All zircon grains from this sample show clear and well-defined oscillatory zoning as well as relatively high Th/U ratios, between 0.3 and 0.5. Zircon analyses from a single population define a discordia line with an upper intercept age of 3229 ± 8 Ma (MSWD= 1.3; *n* =70; Fig. 7h). The most concordant zircon analyses yield a Concordia age of 3224 ± 7 Ma ($MSWD_{(C+E)}$ = 0.13; n = 13) and considered as the crystallization age.

c. Granodiorite

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

d. Granite

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

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

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

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

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

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

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

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

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

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

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

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

4.3 Zircon Lu-Hf isotopic compositions

For each rock sample, a subset of 10 to 30 representative zircon grains has been analyzed for in-situ Hf isotopes (supplementary material Table S4) and selected based on concordance (±2 %) and $^{207}Pb/^{206}Pb$ age (Table 2). Measurements were acquired on the same spot used for the U-Pb analyses (supplementary material Figure S3). Figure 8 shows the initial εHf values for each zircon grains, calculated by using their corresponding $^{207}Pb/^{206}Pb$ age. However, given that the studied samples are granitoids with a single crystallization age, the initial Hf isotopic composition for all of the zircon grains from a single sample should be at the same crystallization age. Therefore, initial Hf isotope compositions of each zircon analyses from the same sample were also calculated at the age of crystallization reported in Table 2 and the average initial εHf values (± 2SE) for each sample are also shown on Figure 8. This is performed in order to avoid calculating an initial Hf composition at an over or underestimated $^{207}Pb/^{206}Pb$ age for zircon grains that would have suffered ancient Pb-loss but remained on the Concordia curve (Fisher et al., 2014a, 2014b; Kemp et al., 2010) as illustrated in Figure 9. Figure 8 shows that Eoarchean samples (3883 to 3781 Ma) display initial Hf isotope compositions that are suprachondritic, with average initial εHf values ranging between +3.7 to +1.7. The Mg-rich tonalite sample SG-026 contains zircon grains exhibiting the most radiogenic Hf isotopic composition with an average initial εHf value of +3.7 ± 0.2 at 3820 Ma, falling on the depleted mantle evolution line. Most samples with positive εHf initial values are tonalitic or trondhjemitic in composition, except for granitic sample SG-007 showing a complex zircon population (Fig. 7j) and with initial $^{176}Hf/^{177}Hf$ ratios between 0.28027 and 0.28042 for all zircon grains analyzed and an average εHf initial value of +2.0 at 3883 Ma. However, given the complex and heterogeneous age populations of this sample, this initial εHf must be taken with cautious as its Eoarchean age is based on the three oldest zircon grains, that may be inherited or recrystallized. If the crystallization age of most zircon grains from this sample is rather Neoarchean, recalculation of initial εHf values at 3883 Ma for all zircon analyses would not be representative of its source. Five granitoids with Paleoarchean ages from 3632 to 3330 Ma display subchondritic initial $^{176}Hf/^{177}Hf$ ratios corresponding to average εHf initial values from -1.0 to -6.3. The low εHf initial value of -4.6 at 3576 Ma for the oldest zircon grain from the sample SG-017, however, is from an inherited grain and the main Neoarchean zircon population for this sample shows an average εHf of -12.7 ± 0.3 at 2802 Ma. Sample SG-203 is the only granodiorite sample analyzed but displays the same 3330 Ma age and Hf isotopic composition as the spatially associated granitic sample SG-208, with initial εHf values of around -6. The later ~3200 Ma Paleoarchean Mg-rich tonalite from Lister Island, sample SG-265, is characterized by a radiogenic initial Hf isotopic composition and does not appear to follow the general εHf trend *vs.* time that most other granitoids display on Figure 8. This sample has a slightly suprachondritic initial εHf value of +1.0 ± 0.3 at 3224 Ma. The late Mesoarchean granitic sample SG-087 is characterized by an initial εHf of -2.6 ± 0.2 at 2996 Ma. However, this sample has multiple and complex zircon generations (Fig. 7o) and geochronology results should be taken with caution as it also includes few inherited older zircon cores with initial εHf= +1.5 ± 0.4 at 3745 Ma.

Three Neoarchean granite samples dated between 2789 and 2744 Ma have been analyzed for zircon Hf isotopic composition. They all display low initial $^{176}Hf/^{177}Hf$ ratios corresponding to average εHf initial values of -11.2 to -14.3. These Neoarchean granite samples include a few inherited older grains which εHf values fall within the main time *vs.* initial εHf array with two ~3660 Ma cores with initial εHf = -0.2 to -1.7 and one 3303 Ma core with initial εHf= -7.8. The Neoarchean zircon population of the trondhjemitic sample SG-019 yielded an average εHf initial value of -11.3 if the 2732 Ma age if considered as the crystallization age. This sample also includes two zircon cores with older ages of 3477 and 3553 Ma that respectively yielded initial εHf values of -3.3 and +7.0. However, given the ambiguity of the crystallization age of this sample, these initial εHf values need to be considered with prudence. A few Eoarchean to Paleoarchean granitoids display Neoarchean zircon rims (sample SG-024, SG-025, and SG-027) with variable initial εHf values ranging from ~ -15 to ~ -22. These zircon grains, however, show variable and often low Th/U ratios, which could be indicative of metamorphic recrystallization. Therefore, the initial εHf values calculated at these ages may not be reflective of their crustal source.

5 Discussion

5.1 Composition and petrogenesis of the SHC felsic crust

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

Given the high-grade metamorphic conditions of the SHC, our interpretation is focused toward the least mobile elements, although the SHC granitoids appear to have relatively well-preserved whole-rock geochemical compositions, despite metamorphism reaching up to granulite facies. Granitoids from the SHC display variable geochemical compositions, but the samples analysed here are consistent with geochemical compositions of SHC felsic rocks from the literature (Bridgwater and Collerson, 1976; Schiøtte et al., 1993, 1989a) and can be divided into four main rock types including trondhjemite, Mg-rich tonalite, granodiorite and granite. Table 3 and Figure 10 present the main compositional characteristics of each granitoid groups. While there is no systematic relationship between the age of the SCH granitoids and their whole-rock chemical compositions, there appears to be a general temporal evolution of the SHC felsic rocks composition. The TTG are more commonly found in the Eoarchean and the Paleoarchean, whereas most granite samples are often Neoarchean (Table 2). This broad compositional secular evolution is observed in most Archean cratons (Laurent et al., 2014).

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

Several trondhjemite samples exhibits pronounced positive Eu anomalies that appear to correlate with La/Yb and Zr/Nd ratios (Fig. 11). Different petrogenetic processes have been proposed to explain positive Eu anomalies in Archean granitoids such as accumulation of feldspars, fractionation of small amounts of allanite or residual rocks from partial melting (Condie et al., 1985; Martin, 1987; Rudnick, 1992). Given the complex reworking history displayed by some SHC granitoids, supported by the high discordance of some Eoarchean zircon (Fig. 7b-c-d & supplementary material Table S3) and the relative abundance of secondary sector zoned zircon in the trondhjemite, it is possible that the trondhjemite which exhibit positive Eu anomalies represent residual rocks after some extent of partial melting. All trondhjemitic samples analyzed here for U-Pb include at least a small proportion of Neoarchean sector zoned igneous zircon, despite their Eoarchean crystallization ages. The main zircon population of sample SG-019, which displays the largest Eu anomaly, is Neoarchean with only 2 older zircon cores preserved. Field observations support the relatively high migmatitisation of the Saglek-Hebron Complex crust, which exhibits important, if not systematic, leucocratic veins of granitic melt (see supplementary material Fig. S1). Contrastingly, the Mg-rich tonalite and granodiorite do not exhibit such positive Eu anomalies and show higher REE contents, suggesting that their whole-rock geochemical composition may not have been affected by remelting event(s) to the same extent as the trondhjemitic rocks. The SHC granodiorite have a REE composition similar to modern granite as opposed to the high La/Yb ratios typically displayed by TTG (Fig. 5).

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

Figure 12 can be used to determine the potential source for felsic melts and therefore, the precursors of Archean granitoids. The SHC TTG appear to result from the melting of a mafic source characterized by variable K contents. The SHC granitic rocks, on the other hand, are more consistent with a felsic crustal source. The older Eoarchean and Paleoarchean tonalite and trondhjemite are the likely crustal precursor sources to the Neoarchean granitic rocks, which would explain the wide range of REE compositions of the granite with La/Yb ratios similar to both the Mg-rich tonalite and the trondhjemite. Several granite samples exhibit negative Eu anomalies which could be complementary with the positive anomalies found in the trondhjemitic (perhaps restitic) rocks. The ternary diagram shown in figure 12b was proposed by Laurent et al. (2014) to highlight the relative end-member petrogenetic processes involved in the formation of Archean granitoids. A high Na_2O/K_2O ratio indicates the melting of mafic rocks producing K-poor granitoids, a high A/CNK ratio is consistent with an Al-rich source such as metasediments, and the FMSB [FSBM = $(FeO_t + MgO) \times (Sr + Ba)$ wt. %] value is indicative of the interaction with a metasomatized mantle. Applied to the SHC granitoids, this diagram further supports the hypothesis that the likely source of the tonalite and trondhjemite is a mafic crust, whereas the granite is more consistent with derivation from the melting of Al-rich felsic crust. Some Mg-rich tonalite and granodiorite samples plot towards the FMSB end member, suggesting that they may have interacted with, or included, a mantle component. Similar geochemical characteristics are observed in Archean sanukitoid (Laurent et al., 2014) originally defined as diorite to granodiorite with high Mg# (>0.6), Ni (>100 ppm) and Cr contents (200-500 ppm), with variable TiO_2 contents and relatively high K, Sr, Zr and Nb concentrations (Shirey and Hanson, 1984). Despite the geochemical signature that could be supportive of interaction with the mantle and the relatively high abundance of

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

5.2 Geochronology of the SHC granitoids

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

To get a statistical overview of the age distribution of the felsic magmatism in the SHC, we have used Kernel density estimation diagrams (KDE; Fig. 13a-b) to highlight the probability of a zircon to be either metamorphic or magmatic, based on the textural and chemical characteristics of each zircon, at a given time. Probabilities from KDE, provide an overview of maximum magmatic production, thermal recrystallization and relative timing between different maximums, as well as show whether magmatic production is punctual (sharp Gaussian peak) or diffused in time (asymmetrical peak or large wavelength Gaussian distribution). It should also be noted that sample bias or missing data can be a caveat for interpretation of these KDE.

5.2.1 Magmatic history

The SHC is one of the rare geological terrains on Earth preserving Eoarchean rocks. Precise and accurate age determination of Earth's oldest rocks has important implications, as it brings timing constraints on various geological processes to understand the early Earth (e.g. Whitehouse et al., 2019). Shimojo et al. (2016) proposed age of 3920 ± 49 Ma for a banded grey gneiss sample (LAA995) they interpreted as from the Iqaluk gneiss. This would represent the second oldest occurrence of felsic crust on Earth, after the Acasta gneiss (Bowring and Williams, 1999; Reimink et al., 2016). However, over 300 spots from 231 zircon were analyzed by LA-ICP-MS for this sample and only the 6 oldest zircon were used to define this age, of which 5 yield over-concordant ages. Whitehouse et al. (2019) recently questioned the exactitude of this age and pointed out that a larger subset of analyses from this granitoids yields a statistically significant population with a mean $^{207}Pb/^{206}Pb$ age of 3865 ± 4 Ma. Sample SG-210c (this study) is the same banded grey gneiss, collected on the same outcrop as sample LAA995 (supplementary material Fig. S4). The Concordia age we obtained for the 34 oldest concordant zircon from sample SG-210c is 3869 ± 6 Ma (Fig. 7g). When we apply the same data filtering method used for SG-210c to select a subset of analyses from sample LAA995 (±2% concordance; Th/U >0.3; $^{207}Pb/^{206}Pb$ age older than 3800 Ma), it defines a consistent concordant population with a well-defined Gaussian distribution around a Concordia age of 3869±5 Ma (MSWD=1.1 n =181). Although it still represents one of the oldest rocks on Earth, the oldest Iqaluk granitoids unit is thus more consistent with a <3900 Ma age, which was displayed by other granitoids samples from the SHC, such as 3860 ± 10 Ma (Vezinet et al., 2018); 3851 ± 46 Ma, 3829 ± 27 Ma, 3869 ± 31 Ma, 3895 ± 33 Ma and 3897 ± 33 Ma (Komiya et al., 2017); 3849 ± 260 Ma (Collerson, 1979); 3863 ± 12 Ma (Schiøtte et al., 1989a).

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

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

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

5.2.2. Metamorphic history

The SHC has been subjected to a complex thermal history with prevalent metamorphism between 2700-2800 Ma, but later high-temperature events have also been recognized (Komiya et al., 2017; Kusiak et al., 2018; Sałacińska et al., 2019, 2018; Schiøtte et al., 1992; Van Kranendonk, 1990). Figure 13b shows the KDE diagram of the probability density ages for recrystallized zircon grains and those exhibiting evidence of age resetting. Only zircon grains yielding concordant (±2%) ages, and thus fully reset ages, are included on Figure 13b. Zircon grains which exhibit low Th/U ratios (>0.3) and homogeneously black texture or heterogeneous patterns (metamict) have also been used as criteria for metamorphic zircon. Eoarchean zircon grains consistent with a metamorphic recrystallization exhibit only one major probability peak that appears to be contemporaneous, if not slightly later, to the ~3600 Ma Uivak II magmatic event. This would be consistent with a thermal event described in previous work (Bridgwater et al., 1975; Collerson, 1983b; Sałacińska et al., 2018; Van Kranendonk, 1990). These concurrent igneous and metamorphic events are supported by the presence of 3600 Ma metamorphic recrystallization rims surrounding pristine igneous oscillatory zoned 3860 Ma old zircon cores (*e.g.* SG-210c; Fig. 6c & Fig. 7g; supplementary material Table S3), as well as the presence of inherited 3805 Ma igneous cores surrounded by a Uivak II oscillatory zoned crystallization (*e.g.* SG-080; Fig 6e & Fig 7m; supplementary material Table S3). An older subordinate Eoarchean population of metamorphic recrystallized zircon grains is seen at ~3750 Ma, but we suggest this could result from ancient Pb-loss in zircon from the Iqaluk gneiss rather than reflecting a distinct metamorphic event. Two minor metamorphic peaks can be observed at 3250 Ma and ~3000 Ma (Fig. 13b), but these late Paleo- to Mesoarchean metamorphic events remain equivocal due to the lack of resolution of the KDE diagram and the relative rarity of these populations.

The highest probability peak of recrystallization occurs during the Neoarchean (Fig. 13b), which is consistent with the extensive metamorphic event previously suggested at 2700 Ma (e.g. Kusiak et al., 2018; Schiøtte et al., 1989b). However, our maximum of recrystallized zircon ages appears to occur closer to 2800 Ma, perhaps shortly before the 2750 Ma Neoarchean peak defined by igneous zircon grains (Fig. 13a). This may suggest a slight delay (~50 Ma) between the maximum intensity of the thermal event and the maximum probability of granitic magmatism, which may correspond to crustal anatexis. The overall distribution of the Neoarchean recrystallized zircon shows an asymmetric peak shape that reaches its maximum at 2800 Ma, before gradually decreasing to lower probabilities until 2200 Ma, after which no zircon is produced in the SHC. This asymmetric peak could be an artifact from our selection criteria used to build the KDE diagram, which could comprise zircon grains that experienced incomplete Pb-loss. However, we cannot disregard the possibility that it suggests protracted high-grade metamorphism before gradually decreasing until 2400-2200 Ma. The apparent asynchrony between the igneous and metamorphic zircon ages may therefore be due to some metamorphic zircon grains that have not been fully reset. The zircon grains with ages <2700 show high U concentrations (Fig. 13b top right inset) and could be more susceptible to reopening of the U-Pb system due to the higher rate of U decay causing the breakdown the zircon crystal lattice (Lee et al., 1997). Therefore, a protracted thermal event that gradually decreased in intensity would keep the U-rich zircon in isotopically open conditions longer than those with lower U contents. The latter would then record older ages compared to the more U-rich zircon grains. This protracted thermal event is also in agreement with what was suggested by Kusiak et al. (2018) based on younger ages obtained on monazite and apatite in the SHC of 2600-2500 and 2200 Ma respectively.

5.3. Crustal sources and reworking history

The U-Pb-Hf isotopic composition of the Eoarchean and Hadean Jack Hills detrital zircons provided invaluable information about the early crust (e.g. Amelin et al., 1999; Blichert-Toft and Albarède, 2008; Harrison, 2005; Kemp et al., 2010), but the fact that their host rocks have been eroded away limits the constraints we can put on the earliest crustal history. Possible ancient Pb-loss in detrital zirconscan also result in younger apparent $^{207}Pb/^{206}Pb$ ages and consequently bias the calculated initial Hf isotopic composition (e.g. Amelin et al., 2000; Vervoort and Kemp, 2016). The SHC zircon grains studied here are all from rock samples with a single interpreted crystallization age. As shown on Figure 9, to avoid the possible calculation of the Hf isotopic compositions at an incorrect apparent Pb-Pb age, we are considering the average initial εHf value for all magmatic zircon grains from each granitoids, calculated at the crystallization age of their respective host rock (Table 2, Fig. 14). The composition we attribute to the precursor crustal source of the rocks hosting the zircon grains has an important implication on how the same dataset can be interpreted regarding crustal history. For example, a 3.5 Ga zircon with an initial εHf value of -5 could be produced from the remelting of a 3.8 Ga felsic crustal precursor or a 4.3 Ga mafic crustal precursor, assuming $^{176}Lu/^{177}Hf$ ratios of 0.01 and 0.025 respectively, and a chondritic initial $^{176}Hf/^{177}Hf$ ratio for both sources. To better constrain the Lu/Hf ratio for the crustal precursor(s) of the SHC granitoids, we use here the whole-rock geochemical composition of each of the host rock samples as a proxy for the composition of their crustal sources and evolution. As discussed previously, we interpret SHC TTG to be derived from the melting of a mafic precursor, while the granitic rocks are consistent with the melting of a felsic crustal source (Fig. 12a).

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

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

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

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

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

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

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

Another scenario than melting Hadean mafic crust that could possibly explain the low εHf values of the zircon grains from the ~3600 Ma Uivak II and ~3300 Ma Iluilik TTG, would be derivation from a mixed crustal source. One could imagine that by the Paleoarchean, the SHC crust was a mixture of mafic Nulliak metavolcanic-type rocks and felsic Iqaluk-Uivak I tonalite-trondhjemite. The contribution of an older felsic crustal component in the source of the Paleo- to Mesoarchean TTG could result in lower initial εHf values. Assuming that both mafic and felsic end-member sources were formed at 3880 Ma, with respective Hf concentrations of 1.1 and 3.5 ppm [average Hf concentrations for the Nulliak mafic rocks (Wasilewski et al., 2019) and SHC tonalite-trondhjemite (Table 1)] and evolved with respective $^{176}Lu/^{177}Hf$ ratios of 0.025 and 0.01, it would, require that the felsic component represent nearly 70% of the mixed crustal source to explain the zircon Hf isotopic composition of the 3630 Ma Uivak II tonalite. Even if a lower $^{176}Lu/^{177}Hf$ ratio of 0.001 [corresponding to the lowest Lu/Hf ratio measured in the SHC trondhjemite (Table 1)] is considered for the felsic source, more than 50% felsic component is needed to account for the εHf value of the Uivak II tonalite. This is however inconsistent with its whole-rock geochemical composition which infers a dominantly mafic source (Fig. 12). Mixing older SHC felsic rocks in the source of the 3300 Ma Iluilik granodiorite to account for its low zircon initial εHf value is also unlikely as the Iluilik rocks have the highest Hf concentrations (≥6.8 ppm) of all SHC felsic rocks and therefore, felsic crust assimilation or mixing would only have a limited effect on its Hf isotopic composition. Except for Neoarchean sample SG-087, all other SHC granite samples define an initial εHf *vs.* time trend consistent with the reworking of an Eoarchean felsic reservoir evolving with a $^{176}Lu/^{177}Hf$ of ~0.011 that originates from the Iqaluk/Uivak I gneiss (Fig. 14). Most inherited zircon grains are found in granitic samples and follow the same felsic reservoir trend (Fig. 14), which further supports the hypothesis that the granite is produced from the melting of the Eoarchean felsic crust. The combined whole-rock geochemical and Hf isotopic compositions of the granitic rocks of Iqaluk, Uivak I, Uivak II, Iluilik and Neoarchean ages, therefore suggest that they were produced from the remelting of the SHC Eoarchean tonalite/trondhjemite over ~1 billion years. The ~3000 Ma granitic sample SG-087, however, would be consistent with the reworking of the ~3200 Ma Lister gneiss rather than the Eoarchean TTG. Except perhaps for the trondhjemite sample SG-019 with an equivocal crystallization age, all Neoarchean granitoids analyzed here are granitic in composition with low initial zircon εHf values. This suggests that the ~2750 Ma felsic magmatism is dominated by remelting older felsic crust, with little to no input from a more juvenile source. The major Neoarchean high-grade metamorphic event that affected the SHC, therefore, appears to also have been accompanied by important crustal reworking and anatexis.

Given that the SHC is part of the North Atlantic Craton, it is often compared to the Itsaq Gneiss Complex of Southwest Greenland, also comprising Eoarchean TTG and supracrustal rocks (Bridgwater et al., 1990; Collerson, 1983a; McGregor, 1973; Morino et al., 2017; Næraa et al., 2012; Wasilewski et al., 2019). Figure 14 shows the evolution of Hf isotopic compositions for detrital and igneous zircon from southwest Greenland (Næraa et al., 2012; 2014) compared to the SHC igneous zircon. The Eoarchean zircon grains from Greenland overall display lower initial εHf values, with mostly chondritic to slightly subchondritic compositions, compared to the SHC where the 3700-3900 Ma zircon grains generally display positive initial εHf values (Fig. 14). The zircon Hf isotopic compositions from both SW Greenland and the SHC are mostly consistent with reworking of Eoarchean crust until the late-Paleoarchean when it abruptly shifts from low initial εHf values around 3300 Ma to suprachondritic compositions at ~3200 Ma (Fig. 14; Hoffmann et al., 2011; Næraa et al., 2012). This was interpreted by Næraa et al. (2012) to represent a change of geodynamic setting in SW Greenland that involved juvenile crust generation by plate tectonic processes. An important input of juvenile crust was recorded throughout the Mesoarchean in SW Greenland with mostly positive initial zircon εHf values between 3200 and 2800 Ma (Næraa et al., 2012), but the lack of Mesoarchean zircon-bearing rocks in the SHC does not allow to assess if the same processes were involved during that time. Similarly to the SHC Neoarchean granitoids, the Neoarchean Qôrqut Granite Complex of SW Greenland displays low zircon εHf values between -12 and -18 at ~2550 Ma. However, these are believed to be derived from an Eoarchean mafic source (Naeraa et al., 2014) rather than from reworking of older felsic crust, such as what we propose for the SHC Neoarchean granitoids. Therefore, the dominant crustal source of the SHC and SW Greenland felsic rocks seems to diverge in the Neoarchean.

5.4. Tectonic context of the SHC

One of the most highly debated subjects about the early Earth concerns the tectonic setting operating and responsible for the formation of the Archean cratons (e.g. Bédard, 2016; de Wit, 1986; Grove et al., 2003; Johnson et al., 2013; Moyen and Laurent, 2018; Nutman and Bennett, 2019; Smithies and Champion, 2000; Wiemer et al., 2018). The geochemical and isotopic compositions of ancient rocks have been widely used to constrain the tectonic environments in which they formed, but there is still no consensus on Earth's early geodynamics. Although the compositions of the SHC metavolcanic rocks do not exhibit geochemical signatures typically found in suprasubduction environments (Wasilewski et al., 2019), Komiya et al. (2015) suggested evidence of Eoarchean subduction settings in the SHC. Based on field observations, they describe the stratigraphy of the Nulliak supracrustal assemblage as an analog to the duplex structures observed in the Japanese trench, interpreted as large accretion prisms. In this model, the Uivak I granitoids would be produced from the melting of basaltic crust within an accretionary complex at shallow depths. However, Uivak I and Iqaluk TTG exhibit strong HREE depletion (Fig. 4 and Fig. 5) consistent with the melting of a garnet-bearing precursor at relatively high-pressures, which is at odds with shallow-level melting of a tholeiitic or possibly komatiitic crust. Although U-Pb and Hf isotopes cannot directly be linked to tectonic settings, these data can help to constrain the type of crustal precursor and its reworking history, to better establish the architecture of the reworked crustal sources and evaluate which tectonic context would be more likely. Figure 15 shows the rock or inherited zircon crystallization age *vs.* mantle extraction Hf model age for the source of the SHG TTG. The mantle extraction model ages are obtained using the zircon average initial $^{176}Hf/^{177}Hf$ ratios for each of the samples, back-calculated to the depleted mantle evolution line, and assuming a $^{176}Lu/^{177}Hf$ between 0.020 and 0.026 for a mafic crustal reservoir. Consequently, only the samples consistent with derivation from a mafic crustal precursor (i.e. the TTG) and inherited zircon grains from granitic samples (which likely crystallized within their TTG precursor) are included on Figure 15. Although it is more difficult to assess if the inherited zircon grains were affected by ancient Pb-loss, which could lead to erroneous calculation of their initial Hf isotopic composition, data from inherited grains are consistent with the other samples and follow the same trend on Figure 15. The zircon grains from Eoarchean granitoids are the only samples with relatively juvenile compositions, consistent with the remelting of almost contemporaneous mantle-derived mafic crust, potentially the Nulliak basaltic crust. The younger Uivak II and Iluilik granitoids appear to be derived from the melting of pre-Nulliak Hadean mafic crust. Except for the Lister gneiss (sample SG-265), the SHC granitoids derived from the melting of mafic crust show a negative correlation between their crystallization age and their Hf model age, suggesting that the latest granitoids were produced from the melting of the oldest crustal source. In a modern-style subduction setting, it would be unlikely that: -1) the subducting mafic crustal reservoir displayed such a wide age range (from ~3900 to ~4400 Ma) and -2) the oldest portions of the subducting mafic crust melted the latest. While one could suggest that a thick mafic crust, similar to an oceanic plateau, could include mafic crust with the range of ages suggested by the Hf model ages from Figure 15, it would be expected that the base of this thickened mafic crust, *i.e.* the oldest portions, melts before the younger overlying portions. This is inconsistent with the negative trend observed in Figure 15. Some authors have proposed a “subcretion” model as an alternative to modern-style subduction tectonics (Barr et al., 1999; Ducea et al., 2009; Grove et al., 2003; Hacker et al., 2015; Taramon et al., 2015). Sequential stacking or tectonic imbrication have been suggested as a mechanism for early Archean crustal growth (e.g. de Wit, 1986; Smithies and Champion, 2000) and the formation of the Eoarchean Itsaq Gneiss Complex of Southwest Greenland (Nutman et al., 2007; Nutman and Bennett, 2019). This tectonic regime has the potential to imbricate portions of mafic crust of variable ages in a configuration where older crust could overlie younger crust and perhaps melt later, in a thickened oceanic crust. Reworking of such an imbricated mafic crustal source could explain the crystallization age *vs.* Hf model age trend seen for the SHC TTG.

6 Conclusion

The SHC granitoids record more than 1 billion years of complex crustal history, including several episodes of felsic magmatism. The different generations of gneiss in the SHC were previously associated with certain geochemical characteristics, but they appear to be more geochemically and petrologically diverse. We, therefore, suggest that the different gneissic units should refer to distinct temporal magmatic events regardless of composition. Six distinct felsic magmatic events can be identified in the SHC: the oldest 3850 Ma Iqaluk gneiss, the 3750 Ma Uivak I gneiss, the 3600 Ma Uivak II gneiss, the 3330 Ma Iluilik gneiss, the 3230 Ma Lister gneiss, and the 2700-2800 Ma Neoarchean granitoids. The chondritic to subchondritic Hf compositions that Eoarchean zircons from most other early terrains exhibit has led to the suggestion that the depletion of the mantle through extraction of crustal material prior to 3.8 Ga was negligible (e.g. Vervoort and Kemp, 2016). The slightly suprachondritic Hf isotopic compositions of the Eoarchean zircon from the SHC granitoids, however, denote the existence of a long-term depleted source, more consistent with the evidence for an early depleted mantle as suggested from Nd isotopes (e.g. Bennett et al., 2007, 1993; Boyet and Carlson, 2006; Caro et al., 2003; O'Neil et al., 2016; Rizo et al., 2011). Most of the juvenile continental crust in the SHC appears to have been formed during the Eoarchean. Considering the geochemical composition of the SHC granitoids to constrain the nature of their crustal precursor, the Hf isotopic compositions of the different generations of TTG suggest the remelting of Eoarchean to Hadean mafic crust. The juvenile Iqaluk and Uivak I TTG are consistent with remelting of ~3800-3900 Ma mafic crust, while the younger Uivak II and Iluilik TTG would suggest melting of increasingly older (up to ~4300 Ma) mafic sources As observed in SW Greenland, the SHC granitoids show a marked transition between ~3300 and 3200 Ma from relatively unradiogenic to more juvenile compositions. The Neoarchean felsic magmatism in the SHC, however, appears to be dominated by the reworking of the Eoarchean TTG, without contribution of juvenile material.

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Captions

Figures:

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

Tables:

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

Table 2: Summary of geochronological data and initial εHf-zircon for the SHC granitoids. Age type = method used to calculate the age. *n* = number of zircon analyses used in age calculation. Age= interpreted crystallization, inherited or metamorphic age. $\varepsilon Hf_{(i)}$ is the average initial values of εHf of all zircon analyses calculated at the crystallization age of their host rock. Full dataset can be found in the supplementary material Tables S3 and S4.

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

Supplementary material:

FIGURES:

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

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

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

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

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

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

TABLES:

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

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

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

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

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