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Lithological, structural, and geochemical characteristics of the Mesoarchean Târtoq greenstone belt, South-West Greenland, and the Chugach-Prince William accretionary complex, southern Alaska: Evidence for uniformitarian plate-tectonic processes

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Abstract: The Mesoarchean Târtoq greenstone belt, South-West Greenland, consists of tectonically imbricated slices of metamorphosed basalt, gabbro, peridotite and sedimentary rocks, and is intruded by felsic rocks (now mylonites) with well-preserved duplex structures, representing a relict accretionary prism. The Târtoq greenstone belt is a remnant of a supra-subduction zone ophiolite that originated as back-arc basin oceanic crust. Following the initiation of intra-oceanic subduction, the back-arc oceanic crust accreted to the overriding plate, forming an accretionary prism. The felsic mylonites are compositionally akin to Archean tonalite-trondhjemite-granodiorite suites (TTG). Field observations, along with geochemical and zircon U-Pb age data, indicate that the protoliths of the felsic mylonites were derived from partial melting of back-arc basalts in the accretionary prism and emplaced along thrust faults between 3012±4 and 2993±6 Ma. It is proposed that the partial melting of the basalts likely occurred in response to ridge subduction. The Upper Cretaceous turbiditic greywackes of the Chugach-Prince William accretionary complex in southern Alaska are intruded by Paleogene felsic dykes. These felsic dykes appear to have been derived from partial melting of subducted and/or accreted oceanic crust during slab window magmatism. Archean granitoid-greenstone terrains share many geological characteristics of Phanerozoic subduction-accretion complexes such as the Alaskan and Altaid subduction accretion complexes, consistent with the operation of uniformitarian geological processes in the Archean. The Archean Earth might have been dominated by numerous smaller plates and greater ridge length than today that would have resulted in more frequent ridge-accretionary prism interactions and larger volumes of TTG generation in subduction-accretion complexes.
Introduction

The Archean Eon (4.0-2.5 Ga) constitutes about 30% of Earth’s history. However, the volume of preserved Archean crust is less than 10% of the present-day continental crust (see Condie 1981; Goodwin 1996; Hawkesworth et al. 2010; Dhuime et al. 2012). Crustal growth models suggest that 60-80% of the continental crust formed by 2.5 Ga (see Dewey and Windley 1981; Reymer and Schubert 1984; Taylor and McLennan 1995; Belousova et al. 2010). The occurrences of Archean rocks and recycled zircon grains in all major continents imply that a large volume of Archean continental crust has been reworked and recycled back into the mantle (Armstrong 1981; Fyfe 1978; Jacobsen 1988a, 1988b; Bowring and Housh 1995; Crowley et al. 2005; Hawkesworth and Kemp 2006a; Nebel-Jacobsen et al., 2010; Li et al. 2015; Paquette et al. 2015; Nutman et al. 2015a; Santosh et al. 2016). Thus, preserved Archean crust likely constitutes only a small fraction of the rocks formed between 4.0 and 2.5 Ga.

Recycling and melting of oceanic crust, growth of continental crust, and accretion of tectonic blocks (e.g., island arcs, oceanic plateaus and microcontinents) in the Phanerozoic Eon have occurred mainly at subduction zones (von Huene and Scholl 1991; Şengör and Natal’in 1996; Monger and Price 2002; Iizuka et al. 2010; Clift and Vannucchi 2004; Burke 2011; Gazel et al. 2015). Modern subduction zones are also major sites of crustal erosion (Scholl and von Huene 2010; Amato et al. 2013). Field observations, and geochemical and geochronological data from Archean cratons are also consistent with the growth of Archean continental crust at convergent plate margins through tectonic accretion of oceanic crust, including mid-ocean ridge, oceanic plateau and island arc crust, and by addition of melts originating from oceanic crust and upper mantle (Defant and Drummond 1990; Drummond and Defant 1990; Kusky and Kidd 1992; Calvert et al. 1995; Calvert and Ludden 1999; Kusky and Polat 1999; Foley et al. 2002; Rapp et al. 2003; Friend and Nutman 2005; Kerrich and Polat 2006; van der Velden et al. 2006; Harrison 2009; Percival et al. 2012; Furnes et al. 2013; Kusky et al. 2013; Polat et al. 2015). However, despite strong field, geochemical, isotopic, geochronological, geophysical and theoretical evidence indicating that plate tectonics operated in the Archean (see Burke et al. 1976; de Wit 1998; Garde 2007; van Kranendonk et al. 2007; O’Neil et al. 2011; Adam et al. 2012; Kisters et al. 2012; Næraa et al. 2012; Percival et al. 2012; Szilas et al. 2012; Arndt 2013;
Wang et al. 2013; Dziggel et al. 2014; Hynes 2014; Turner et al. 2014; Martin et al. 2014; Blichert-Toft et al. 2015; Nutman et al. 2015a, 2015b; Puchtel 2015; Komiya et al. 2015; Smart et al. 2016; Tang et al. 2016), some geologists still think that the Archean Eon was devoid of plate tectonics (e.g., Hamilton 2013; Stern 2015; Bédard and Harris 2014, 2015).

In the Phanerozoic, structural, magmatic, sedimentary and metamorphic processes involved in the generation of the continental crust are driven by plate tectonics at convergent plate margins (Isozaki et al. 1990; Şengör 1990; Monger and Price 2002; Arndt 2013; Şengör et al. 2014; Gazel et al. 2015). Despite the presence of similar rock associations and structures in the Archean and Phanerozoic Eons (Kisters et al. 2012; Friend and Nutman 2005; Kusky et al. 2013; Berger et al. 2014; Furnes et al. 2015; Polat et al. 2015), the nature of tectonic processes that operated in the former eon is still hotly debated (e.g., Hamilton 1998, 2013; Stern 2005; Bédard 2006; Johnson et al. 2014; Thébaud and Rey 2013; Moore and Webb 2013; François et al. 2014; Kamber 2015).

On the basis of mapping and structural data, Kisters et al. (2012) showed that the ca. 3.0 Ga Târtoq greenstone belt, also known as the Târtoq Group or Târtoq supracrustal belt, in South-West Greenland is an Archean subduction-accretion complex (Fig. 1). Szilas and co-workers (2013) presented extensive major and trace element data for the Târtoq greenstone belt and zircon U-Pb intrusion age (2986±4 Ma) data for the TTG gneisses, suggesting that the Târtoq greenstone belt represents a relict fragment of Mesoarchean supra-subduction zone oceanic crust. In this study, we present new field, major and trace element data for metamorphosed gabbros (meta-gabbro), mafic volcanic rocks (greenschist), syn-tectonic felsic intrusive rocks (felsic mylonite) and felsic dykes, and zircon U-Pb geochronological data for the felsic mylonites exposed in the Iterlak area of the Târtoq region (Fig. 1). This study differs from Kisters et al. (2012) and Szilas et al. (2013) in that we present new geodynamic interpretation for the mafic and felsic rocks in the Târtoq greenstone. In addition, we present new field, and major and trace element data for Upper Cretaceous turbidites (Valdez Group) and Paleogene felsic dykes in the Mesozoic subduction-accretion complex in southern Alaska for comparison. The purpose of this comparison is to revisit the models proposed by several studies suggesting that Archean greenstone belts share the geological characteristic of the Alaskan subduction-accretion complex (Williams et al. 1991; Percival and Williams 1989; Şengör and Natal’ in 1996). The new data from South-West Greenland and southern Alaska are used to place new geodynamic constraints
on the origin of Archean continental crust. Global Positioning System sample locations are given in Tables 1-4.

Analytical methods

Major and trace elements

Forty samples were analyzed for both major and trace elements, including twenty-seven samples from the Târtoq greenstone belt and thirteen samples from southern Alaska (Tables 2-4). Major and some trace element (Ba, Sr, Y, Sc, Zr, V) analyses were carried out by ICP, at Activation Laboratories Ltd. (Actlabs) in Ancaster, Ontario. Loss on ignition (LOI) was determined by weight loss upon heating to 1100°C for three hours. The samples were fused in lithium metaborate and lithium tetraborate that was followed by digestion with nitric acid, 5% solution (please see http://www.actlabs.com for further information). Totals for major element oxides are 100±1 wt.%. Detection limits and the results of standards and split sample analyses are given in Supplementary Data Table 1. Reproducibility of major elements analyses was generally better than ±5% of the amount present, except for TiO$_2$ where reproducibility was slightly higher (±7%) (Supplementary Data Table 1). Reproducibility of trace elements analyses was generally better than ±6% of the amount present, except for Sc where reproducibility was significantly higher (±25%) (Supplementary Data Table 1).

Trace element analyses, except for the Alaskan greywacke and felsic dyke samples, including large ion lithophile elements (LILE), rare earth elements (REE), high field strength elements (HFSE), and transition metals were conducted at Geoscience Laboratories (Geo Labs) in Sudbury, Ontario, using inductively coupled plasma-mass spectrometry (ICP-MS) following the analytical protocols of Burnham et al. (2002), Burnham and Schweyer (2004) and Schweyer (2006). The samples were digested by using multi-acid techniques (method code IMC-100). Replicate trace element analysis of sample 510309 is given in Supplementary Data Table 2. Reproducibility of individual trace elements (e.g., Rb, Sr, V, Zr) between the two laboratories was better than ±10% of the amount present for the majority of samples (Supplementary Data Table 3).

Both major and trace elements for the Alaskan greywackes and felsic dykes were analyzed at Activation Laboratories Ltd. (Actlabs) in Ancaster, Ontario, using Code 4B2 (trace element ICP/MS). Fused samples were diluted and analyzed by Perkin Elmer Sciex ELAN 6100 and
9000 ICP/MS. Three blanks and five controls (three before sample group and two after) were analyzed per group of samples. Duplicates were fused and analyzed every 15 samples. Instrument was recalibrated every 40 samples. Detection limits and the results of standard and duplicate sample analyses are given in Supplementary Data Table 4. Reproducibility of most element analyses was generally better than ±5% of the amount present.

The ratios of Eu/Eu*, Ce/Ce*, Nb/Nb*, Pb/Pb*, Zr/Zr* and Ti/Ti* were calculated following Taylor and McLennan (1985). Mg-numbers (%) were calculated as the molecular ratio of Mg/(Mg+Fe$^{2+}$), where Fe$^{2+}$ is assumed to be 90% of total Fe. Elements were normalized to chondrite (cn) and N-MORB (Sun and McDonough 1989).

**Zircon U-Pb dating**

Zircon analyses were conducted at the Department of Petrology and Economic Geology, Geological Survey of Denmark and Greenland (GEUS) and the details of analytical procedures are given in Frei and Gerdes (2009) and Dziggel et al. (2014). The crushed sample was poured onto a Wilfley shaking table where the heavy mineral grains were separated. Zircon grains were subsequently hand-picked from the final heavy mineral concentrate. The hand-picked zircon grains were cast into epoxy and polished to expose a central cross-section of each grain. The mount was documented prior to ablation using backscattered electron (BSE) imaging using a scanning electron microscope (SEM).

All U–Pb age data were acquired by laser ablation - single collector - magnetic sector field - inductively coupled plasma - mass spectrometry (LA-SF-ICP-MS) employing a Thermo Finnigan Element 2 mass spectrometer coupled to a New Wave Research UP213 frequency-quintupled solid state Nd:YAG laser system, employing a two-volume cell technology. The laser was operated at a repetition rate of 10 Hz and nominal energy output of 55%, corresponding to a laser fluency of c. 8 J cm$^{-2}$. All data were acquired with a single spot analysis on each individual zircon grain with a beam diameter of 30 µm and a crater depth of approximately 15–20 µm. For the spot diameter of 30 µm and ablation times of 30 s the amount of ablated material approximates 200–300 ng. The total acquisition time for each analysis was 60 s, with the first 30 s used to measure the gas blank. The instrument was tuned to give large, stable signals for the $^{206}$Pb and $^{238}$U peaks, low background count rates (typically around 150 counts per second for $^{207}$Pb) and low oxide production rates ($^{238}$U$^{16}$O/$^{238}$U generally below 2.5%). $^{202}$Hg, $^{204}$Pb (Pb + Hg),
$^{206}\text{Pb}$, $^{207}\text{Pb}$, $^{208}\text{Pb}$, $^{232}\text{Th}$ and $^{238}\text{U}$ intensities were determined through peak jumping using electrostatic scanning in low resolution mode and with the magnet resting at $^{202}\text{Hg}$. Each peak was determined at four slightly different masses and integrated sampling and a settling time of 1 ms for each isotope. Mass $^{202}\text{Hg}$ was measured to monitor the $^{204}\text{Hg}$ interference on $^{204}\text{Pb}$ where the $^{\text{Hg}}_{202}/^{\text{Hg}}_{204} \equiv 4.36$. $^{207}\text{Pb}/^{235}\text{U}$ was calculated from the $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ assuming $^{238}\text{U}/^{235}\text{U} \equiv 137.88$. Samples were analysed in sequences where three standards bracket each set of ten samples. The raw data were corrected for instrumental mass bias and laser-induced U-Pb fractionation through normalization to the GJ-1 zircon using in-house data reduction software (see Jackson et al. 2004). All isotope data were plotted and evaluated using ISOPLOT/EX 3.71 (Ludwig 2008). Model age calculation and error propagation follow Sambridge and Lambert (1997). Long term external reproducibility was monitored by repeated analyses of the Plešovice zircon standard (Sláma et al. 2008), yielding an average $^{238}\text{U}/^{206}\text{Pb}$ age of 339.4 ± 1.5 Ma (2σ) (n = 351 zircons, MSWD = 0.44), which is in perfect agreement with reported value by ID-TIMS of 338 ± 1 Ma (Aftalion et al. 1989).

After correcting the data for the effects of mass bias fractionation and Hg-interference, a filtering of the data was applied by which analyses with 2σ errors > 10% (abs) for the $^{206}\text{Pb}/^{238}\text{U}$ ratio were removed from the dataset. In some cases a correction for common Pb (cPb) was applied, but unless otherwise stated, these corrected data were not included in the plots or used to base any ages on. Plotting of concordia diagrams was done in an off-line Excel sheet using Isoplot/Ex version 4.15 (Ludwig 2008). Unless otherwise stated, all reported ages are $^{207}\text{Pb}/^{206}\text{Pb}$ ages and the reported errors are at the 2σ level or 95% confidence interval.

**Târtoq greenstone belt, South-West Greenland**

**Regional geology and tectonic evolution**

South-West Greenland contains the world’s best exposed Archean rocks ranging in age from 3850 to 2550 Ma, providing an excellent opportunity to study the characteristics of tectonic, magmatic and metamorphic processes that operated in the Archean (Fig. 1) (Bridgwater et al. 1974, Friend and Nutman 1991 2005; Nutman and Friend 2009; Windley and Garde 2009; Kolb et al. 2012; Dziggel et al. 2014; Polat et al. 2015). The Archean craton of South-West Greenland is composed mainly of TTG orthogneisses, granites, supracrustal belts, and layered anorthositic complexes (Nutman et al. 2004, 2013; Steenfelt et al. 2005; Polat et al. 2009, 2011; Windley and

Boundaries between the TTG orthogneisses and the supracrustal belts and layered anorthositic complexes are typically characterized by 5 to 20 meter wide high-strain shear zones, generally mylonitic in nature (Polat et al. 2015, and references therein). Intrusive relationships between TTGs and supracrustal rocks are also locally preserved. Supracrustal rocks occur as several kilometers wide and several kilometers long poly-deformed and metamorphosed mafic volcanic rocks with subordinate sedimentary rocks, gabbros and ultramafic lenses. Metasedimentary rocks, including mature, siliceous clastic sedimentary rocks, are very rare in the entire Archean craton of South-West Greenland. In addition, smaller size supracrustal rocks are dispersed as trains of 1 to 200 meters wide, 1 to 1000 meters long concordant lenses within the TTG orthogneisses (Bridgwater et al. 1976; Garde 1990; Windley and Garde 2009). The layered anorthositic complexes also occur as trains of meter- to kilometer-scale conformable layers of poly-deformed and metamorphosed anorthosite, leucogabbro, gabbro and ultramafic rocks (Myers 1985).

The geological history of the Archean craton of South-West Greenland is explained by either terrane (tectonic block) accretion models (Friend et al. 1987, 1988, 1996; Friend and Nutman 2001, 2005; Nutman et al. 2007; Kolb et al. 2012; Dziggel et al. 2014) or an arc crustal block accretion model (Windley and Garde 2009). Using extensive field and zircon U-Pb geochronological data, several studies (Friend et al. 1987, 1988, 1996; Friend and Nutman 2001, 2005; Nutman et al. 2007; Nutman and Friend 2007, 2009; Garde 2007; Polat et al. 2007, 2008, 2009; Nutman et al. 2004, 2009; Kolb et al. 2012; Dziggel et al. 2014) proposed tectonic models involving accretion of several fault-bounded Eoarchean to Neoarchean tectonic blocks to explain the geological record in South-West Greenland. Boundaries between different tectonic blocks are typically characterized by up to 200 meter wide sheared supracrustal rocks and serpentinites. The tectonic blocks consist mainly of juvenile TTG gneisses, supracrustal belts, and layered anorthositic complexes (only in the Mesoarchean terranes) and were assembled through several collisions in the Neoarchean (Nutman et al. 2004, 2007; Friend and Nutman 2005; Steenfelt et al.
Collisions were accompanied by magmatism and metamorphism (Friend et al. 1988; Windley and Garde 2009; Kolb et al. 2012; Dziggel et al. 2014).

The arc crustal accretion model proposed by Windley and Garde (2009) includes six fault-bounded tectonic blocks that display similar geological cross sections consisting of southerly upper and northerly lower zones. The upper zones are composed of amphibolite facies metamorphic rocks, whereas the lower zones exhibit granulite facies metamorphism. The tectonic blocks are interpreted as remnants of Archean island arcs and Andean-type continental margins. Both terrane and arc crustal accretion models imply that the Archean craton of South-West Greenland originated through prolonged, complex tectonothermal events that are comparable to those produced in Phanerozoic orogenic belts (Polat et al. 2015).

**Târtoq greenstone belt**

The Târtoq greenstone belt is located in the southern margin of the Archean craton of South-West Greenland (Fig. 1). The belt is exposed as four large fragments including the Nuuluk, Iterlak, Amitsuarsua, and Bikuben (Fig. 1) (Kisters et al. 2012; Szilas et al. 2013). The belt is part of the Kvanefjord block in the arc crustal accretion model of Windley and Garde (2009). The Târtoq greenstone belt is lithologically similar to other Archean supracrustal belts in South-West Greenland, consisting mainly of metamorphosed basaltic, gabbroic, ultramafic, and sedimentary rocks (Fig. 2) (Higgins 1968; 1990; Appel and Secher 1984; Evans and King 1993; Nutman and Kalsbeek 1994; Nutman et al. 2004; Kisters et al. 2012; Szilas et al. 2013, 2014). Field investigations indicate that the Târtoq greenstone belt underwent at least four phases of deformation and greenschist to upper amphibolite facies metamorphism (Kisters et al. 2012). The belt contains many fault-bounded domains and contacts between different domains are mainly marked by felsic mylonite and small-scale duplex structures (Figs. 3, 4). The felsic mylonites are interpreted to have been derived from syn-tectonic felsic rocks emplaced along thrust faults.

Contacts between different rock types often display multiple phases of folding, shearing, transposed fabrics, and hydrothermal carbonate, chlorite and quartz alteration, locally associated with orogenic gold mineralization (Fig. 3) (Polat and Dziggel 2011; Kolb et al. 2013). Due to intense shearing, hydrothermal alteration, and several generations of folding, pillow structures in mafic volcanic rocks have largely been destroyed and preserved only in certain locations (Fig. 4).
Ultramafic rocks, mostly serpentinites, are also strongly deformed and metamorphosed. The serpentinists are interpreted as lower arc cumulates (Szilas et al. 2014). Mesozonal orogenic gold mineralization occurs in several shear zone-hosted quartz-calcite vein systems in the belt (Appel and Secher 1984; Evans and King 1993; Kolb et al. 2013). Hydrothermal gold mineralization is hosted in the subduction-accretion structures similar to settings in the North American Cordillera (Kolb et al. 2013; Goldfarb et al. 1998).

The formation age of the Târtoq greenstone belt is uncertain but the best estimate is at 3189±65 Ma based on whole rock Lu-Hf isotope systematics, whereas Sm-Nd systematics only provide less well defined age constraints (Szilas et al. 2013). The belt has been intruded by the 2990 to 2950 Ma TTG sheets (Nutman and Kalsbeek 1994; Kisters et al. 2012). Some TTG sheets are tectonically imbricated with the volcanic rocks in the belt (Kisters et al. 2012; Szilas et al. 2013).

Kisters et al. (2012) divided the Iterlak part of the Târtoq greenstone belt into two tectonically juxtaposed units (Fig. 1). The upper unit is composed of greenstone, greenschist, meta-gabbro, serpentinite, and thin layers of metamorphosed banded iron formation. Both the greenstone and greenschist were derived from basaltic protoliths. The lower unit is lithologically similar to the upper unit but contains larger proportions of meta-gabbro and serpentinite. The belt is structurally underlain by TTG gneisses.

Extensive geochemical data reported by Szilas et al. (2013, 2014) indicate that the Târtoq greenstone belt formed in a Mesoarchean supra-subduction tectonic setting. On the basis of field characteristics and geochemical data, Szilas and co-workers (2013, 2014) interpreted the belt as a dismembered supra-subduction ophiolite in a subduction-accretion complex. Field studies suggest that the protoliths of the TTG gneisses and the felsic mylonites were derived mainly from partial melting of amphibolites originating from accreted oceanic crust (Kisters et al. 2012). The exposed rocks in the Târtoq greenstone belt represent relict fragments of accreted oceanic crust that escaped partial melting and tectonic reworking.

**Petrography**

On the basis of field observations, samples from the Târtoq greenstone belt are defined as meta-gabbro, greenschist (metabasalt), felsic mylonite, and felsic dyke (Figs. 1-4).
The meta-gabbro consists mainly of chlorite (30-40%), plagioclase (30-40%), amphibole (10-15%), quartz (5-10%), epidote (10-15%), calcite (5-7%), titanite (2-4%), apatite (2-3%), and pyrite (1-3%) (Figs. 5, 6). Plagioclase and amphibole are predominantly albite and actinolite, respectively. The greenschist is composed mainly of chlorite (40-50%), plagioclase (20-25%), quartz (10-15%), zoisite (5-10%), rutile (2-3%), sulfide (2-3%), titanite (2-3%), calcite (3-5%), magnetite (1-2%), and amphibole (3-5%) (Figs. 5, 6). Like in the meta-gabbro, plagioclase and amphibole in the greenschist are predominantly albite and actinolite, respectively. The felsic mylonite contains quartz (35-45%), muscovite (20-25%), albite (10-15%), chlorite (3-5%), epidote (3-5%), rutile (2-3%), apatite (2-4%), biotite (2-3%), titanite (2-3%), and K-feldspar (2-3%) (Figs. 5, 6). The felsic dykes are characterized mainly by quartz (50-60%), muscovite (30-40%), albite (10-15%), and opaque mineral (2-3%) (Figs. 5, 6).

Zircon Pb-Pb ages

Three samples from the felsic mylonite in the Iterlak area were selected for age determination by LA-ICP-MS dating at GEUS (Table 1).

Sample 510312: This sample consists mainly of quartz, feldspar and muscovite, and was collected from a 30-50 cm thick layer (Fig. 3c). The zircons are mostly anhedral to subhedral, and have prismatic habits with aspect ratios of around 1:3 (Fig. 7). Most grains show distinct oscillatory magmatic zoning defined by thin (a few µm), alternating relatively BSE-bright (U-rich) and BSE-dark (U-poor) layers. In total 67 spots were analyzed of which 16 were filtered out due to either poor analysis or presence of common Pb. Of the remaining 51 analyses only five fall outside the 90-100% concordancy range (Table 1). The majority of spots have high Th/U of 0.3–0.7 (Table 1). Based on a weighted average of 49 data points an age of 2997± 10 Ma (MSWD = 1.8, with 2 outliers removed) was calculated (Fig. 8). This age is taken as the crystallization age of the felsic protolith of the mylonite. A single slightly discordant but precise analysis gives an age of 2797 ± 23 Ma, which likely reflects the effects of ancient Pb-loss.

Sample 510325: This sample was collected from a 3-4 m thick layer and likewise consists of quartz, feldspar and muscovite. The outcrop has been overprinted by brittle deformation, as indicated by tectonic breccia. The zircons mostly constitute euhedral or subhedral grains with low aspect ratios of less than 1:2. The grains typically display a simple zoning pattern consisting in a relatively BSE-dark (U-poor) core surrounded by a BSE-bright (U-rich) rim of variable
thickness. Some grains show an hour-glass zonation pattern (Fig. 8; spot 25). Th/U ratios are rather constant throughout, commonly around 0.5 (Table 1). A total of 140 grains were analyzed in this sample and mostly highly concordant and constitute a cluster on the concordia line that yields a well-defined weighted Pb-Pb age of 3012±4 Ma (MSWD = 0.88, no outliers) (Fig. 8). The age is interpreted as the crystallization age of the protolith to the felsic mylonite.

Sample 510328: This sample was taken from a 4-5 m thick layer with boudins and pencil structures (Fig. 3f). The outcrop was overprinted by later brittle deformation. The zircon grains mostly show magmatic zoning and are euhedral to subhedral and prismatic. A few grains are distinctly anhedral and show no clear zoning (Fig. 7; spot 169). In the Tera-Wasserburg diagram the data mostly cluster on the concordia as a single age population. A weighted mean Pb-Pb age is calculated at 2993±6 Ma (MSWD = 0.79, with 4 outliers removed) (Fig. 8). This age is interpreted as the crystallization age of the protolith of the felsic mylonite.

Major and trace elements

Meta-gabbro

The meta-gabbros display moderately variable SiO$_2$ (46.5-50.7 wt.%), TiO$_2$ (0.76-1.15 wt.%), CaO (8.5-10.3 wt.%), MgO (5.0-7.9 wt.%), Al$_2$O$_3$ (12.6-16.5 wt.%), Fe$_2$O$_3$ (11.5-15.7 wt.%), and Na$_2$O (1.5-2.1 wt.) contents (Fig. 9; Table 2). Mg-numbers are between 39 and 58 (Table 2). They have large variations of Cr (23-272 ppm), Ni (45-214 ppm), V (210-415 ppm), Sc (32-51 ppm), Zr (45-75 ppm), and Y (15-28 ppm) concentrations (Fig. 10; Table 2). Al$_2$O$_3$/TiO$_2$ (8-22) ratios are sub-chondritic, Nb/Ta (15.1-21.8) ratios are sub-chondritic to super-chondritic, and Zr/Hf (35-36) ratios are chondritic (see Sun and McDonough 1989).

The meta-gabbros have the following trace element characteristics: (1) slightly depleted to moderately enriched LREE patterns (La/Sm$_{cn}$=0.84-1.41; La/Yb$_{cn}$=0.94-1.76) and slightly depleted HREE patterns (Gd/Yb$_{cn}$=1.11-2.12); (2) small negative Eu (Eu/Eu*=0.92-0.98) and Ce (Ce/Ce*=0.96-0.98) anomalies; (3) large positive Pb anomalies (Pb/Pb*=6.1-19.9); (4) variably negative Nb (Nb/Nb*=0.40-0.62) and Zr (Zr/Zr*=0.77-0.92) anomalies; and (5) small negative to positive Ti (Ti/Ti*=0.86-1.10) anomalies (Fig. 11; Table 2).
Greenschist rocks (metabasalt)

Major elements in the greenschist have larger variations than those in the meta-gabbro (SiO$_2$=44.0-51.5 wt.%; Al$_2$O$_3$=12.8-16.8 wt.%; MgO=4.8-14.9 wt.%; CaO=1.9-12.2 wt%; TiO$_2$=0.60-1.26 wt.%) (Fig. 9; Table 2). Mg-numbers (49-75) are higher than those of the meta-gabbro samples (38-58). They exhibit a large range of K$_2$O (0.04-0.94 wt.%), Na$_2$O (0.26-2.48 wt.%), Cr (43-788 ppm), Ni (42-260 ppm), Rb (1-33 ppm) and Sr (63-616 ppm), whereas Co (38-52 ppm), Sc (26-51 ppm), Zr (48-93 ppm) and V (162-322 ppm) contents are moderately variable (Fig. 10; Table 2). Al$_2$O$_3$/TiO$_2$ (13-25) ratios are mainly sub-chondritic. The ratios of Nb/Ta (14.6-17.6) and Zr/Hf (36-40) are similar to those of the meta-gabbro.

The greenschist samples display more fractionated REE patterns (La/Sm$_{cn}$=1.04-2.24; La/Yb$_{cn}$=1.13-5.34; Gd/Yb$_{cn}$=1.01-1.61; Eu/Eu*=0.82-1.04) than the meta-gabbro samples (Fig. 11, Table 2). Cerium (Ce/Ce*=0.96-1.00) anomalies are minor to absent. They have negative Nb (Nb/Nb*=0.24-0.51), Zr (Zr/Zr*=0.65-0.96) and Ti (Ti/Ti*=0.61-0.93), but positive Pb (Pb/Pb*=3-48) anomalies (Fig. 11). On the basis of REE patterns, the greenschist rocks are divided into two groups. Group 1 (La/Sm$_{cn}$=1.04-1.18; La/Yb$_{cn}$=1.13-1.43; Gd/Yb$_{cn}$=1.01-1.30) is less fractionated than Group 2 (La/Sm$_{cn}$=1.56-2.24; La/Yb$_{cn}$=1.98-5.34; Gd/Yb$_{cn}$=1.18-1.61) (Fig. 11b, c).

Felsic mylonite

The felsic mylonites are characterized by 64-71 wt.% SiO$_2$, 2.2-3.9 wt.% Fe$_2$O$_3$, 0.5-1.7 wt.% MgO, 14.1-18.8 wt.% Al$_2$O$_3$, 054-4.9 wt.% CaO, 1.91-4.98 wt.% Na$_2$O, and 016-0.65 wt.% TiO$_2$ (Fig. 9; Table 2). Mg-numbers vary between 29 and 57 (Table 2). They have variable Cr (5-133 ppm), Ni (9-71 ppm), Sc (6-14 ppm) and V (18-100 ppm) contents (Fig. 10; Table 2). Al$_2$O$_3$/TiO$_2$ (25-40) and Zr/Y (10.3-16.6) ratios are strongly super-chondritic, whereas Nb/Ta (11.0-12.6) ratios are sub-chondritic.

The felsic mylonites display the following trace element features: (1) strongly fractionated LREE patterns (La/Sm$_{cn}$=2.71-5.13; La/Yb$_{cn}$=6.35-18.91; Gd/Yb$_{cn}$=1.53-2.56); (2) negative to positive Eu anomalies (Eu/Eu*=0.69-1.18); (3) Ce anomalies (Ce/Ce*=0.94-1.02) are minor to absent; (4) large positive Pb anomalies (Pb/Pb*=2.7-7.2); (5) large negative Nb (Nb/Nb*=0.13-0.20) and Ti (Ti/Ti*=0.17-0.52) anomalies; and (6) slightly negative to moderately positive Zr anomalies (Zr/Zr*=0.94-1.51) (Fig. 11; Table 2).
Felsic dykes

Only two samples were analyzed from the felsic dykes in the Târtoq greenstone belt (Fig. 2f). They have 75.3-75.7 wt.% SiO$_2$, 0.03-0.05 wt.% TiO$_2$, 13.5-13.6 wt.% Al$_2$O$_3$, 1.3-1.5 wt.% Fe$_2$O$_3$, 0.4-0.6 MgO, 0.7-0.9 wt.% CaO, 5.01-5.79 wt.% Na$_2$O and 0.84-1.43 wt.% K$_2$O (Fig. 9; Table 2). They contain very low transition metal concentrations (Ni=3-6 ppm; Co=1-2 ppm; Cr=5-7 ppm, Sc=3-4 ppm; V=5 ppm) (Fig. 10; Table 2). Al$_2$O$_3$/TiO$_2$ (294-411) ratios are extremely high.

In comparison to the felsic mylonites, the felsic dykes have less fractionated REE patterns (La/Sm$_{cn}$=1.55-1.58; La/Yb$_{cn}$=1.98-2.25; Gd/Yb$_{cn}$=1.09), but stronger depletion of Eu (Eu/Eu*=0.20-0.41) and Ti (Ti/Ti*=0.02-0.03). Like the felsic mylonites, they have negative Nb anomalies (Nb/Nb*=0.29-0.33), positive Pb anomalies (Pb/Pb*=3.5-11.0), and sub-chondritic Nb/Ta (10.4-10.9) ratios (Fig. 11).

Chugach accretionary complex, southern Alaska

Regional geology and tectonic evolution

Southern Alaska consists of a series of accreted plateaus, arcs, with intervening accretionary prisms, ophiolites, and various arc and strike-slip related basins (Fig. 12) (Plafker et al. 1989). Here, we focus on the geochemistry of the sampled Paleogene felsic dykes and Upper Cretaceous greywackes, and their significance for the formation of continental crust (Fig. 13).

Southern Alaska represents one of the largest subduction-accretion complexes in the world and consists of a number of tectonic blocks or slivers ranging in age from Paleozoic to Cenozoic (Fig. 12) (Plafker et al. 1994). In the literature, these tectonic blocks are mainly referred to as “tectonic terranes” (Howell 1985; Coney and Jones 1985; Plafker et al. 1994). The terrane concept has been criticized by Şengör and Dewey (1990) because it obscures the details and significance of the tectonic history of orogenic belts. However, terrane terminology has been embedded in the literature on the Alaskan and Canadian Cordilleran geology over the past four decades. Although we agree with Şengör and Dewey (1990), to avoid adding additional confusion to the existing literature we will refer to terranes for descriptive purposes.

Tectonic events in southern Alaska during the Mesozoic and Cenozoic resulted in accretion of island arcs, subduction-accretion complexes, and ophiolites to the North American continental margin, as well as near-trench magmatic activity (Marshak and Karig 1977; Hudson 1983;
Plafker et al. 1994; Pavlis and Sisson 1995; Pavlis and Roeske 2007; Bradley et al. 2003; Sisson et al. 2003a, 2003b; Kusky et al. 2003, 2007; Trop and Ridgway 2007) (Fig. 12).

The Border Ranges fault system, as one of the major structures in the region (Fig. 12), divides southern Alaska into two main components (Plafker et al. 1994; Sisson et al. 2003a). The Wrangellia composite terrane is located to the north and includes the Alexander, Wrangellia and Peninsular terranes, the latter corresponding to the Triassic to Early Jurassic Talkeetna arc. To the south of the Border Ranges fault is a huge subduction-accretion complex. In the older literature (e.g., Plafker et al. 1994), this tract was subdivided into three terranes, namely the Chugach, Prince William and Yakutat. Most workers now combine the first two into a single entity, the Chugach-Prince William terrane (e.g., Bradley et al. 2003; Garver and Davidson 2015). Subduction-accretion proceeded episodically, during Jurassic, Cretaceous, and Cenozoic times.

Three fault-bounded lithotectonic assemblages have been grouped into the Mesozoic part of the subduction complex, i.e. the Chugach terrane of older literature. These are: (1) a Late Triassic to Jurassic greenschist to blueschist metamorphic assemblage (e.g., Liberty Creek and Seldovia); (2) the McHugh Complex, a mélange of Permian to Cretaceous rocks that was assembled in the Jurassic and Cretaceous; and (3) the Upper Cretaceous Valdez Group consisting mainly of greywacke turbidite deposits (Plafker et al. 1994). The Valdez turbidites were derived mainly from continental arc sources (Plafker et al. 1994; this study).

The McHugh Complex includes tholeiitic pillow basalts, volcaniclastic rocks, greywacke turbidites, argillite, and radiolarian chert, plus minor limestone, gabbro, and ultramafic rocks (Bradley and Kusky 1992; Kusky and Bradley 1999). Coherent, intact fault slices of these rock types are interleaved with mesoscale mélange derived from the same protoliths (Bradley and Kusky 1992; Kusky and Bradley 1999). Small fault-bounded slices of blueschists mark the inboard boundary of the McHugh Complex and its contact with Wrangellia, recording the early subduction conditions in this belt (Lopez-Carmona et al. 2011). The complex underwent zeolite to greenschist facies metamorphism, with locally higher conditions (amphibolite to granulite) recorded adjacent to plutons, and in the Chugach metamorphic complex (Sisson et al. 2003a; Kusky et al. 2003). Detrital zircon data have been interpreted to record pulses of clastic sedimentation at ca. 155-170 and 100-90 Ma, and episodes of subduction erosion at ca. 180-170 and ca. 125-105 Ma (Amato et al. 2013). The Upper Cretaceous Valdez Group is composed
mainly of turbiditic greywacke, siltstone, and argillite, plus minor local conglomerate, tholeiitic pillow basalts and diabase dykes. The Valdez Group is inferred to have been deposited in a trench or near-trench environment associated with a south-southwest facing magmatic arc (Plafker et al. 1994; Bradley et al. 2003). Interpretation of the basalts is problematic, as they are undated and structurally interleaved with accreted turbidites. They have been interpreted as fragments of an oceanic island arc (Page et al. 1986; Plafker et al. 1994).

The Valdez Group is flanked on its seaward side by the Cenozoic part of the accretionary complex. Turbidites of the Paleocene-Eocene Orca Group were deposited in the same trench or near-trench environment as the Valdez Group. Both the Valdez and Orca groups are variably folded and were metamorphosed under zeolite to greenschist facies conditions (Barker et al. 1992; Plafker et al. 1989, 1994).

The most seaward and most recently accreted part of the southern Alaska margin is referred to as the Yakutat terrane (Plafker et al. 1994). It is a subduction-accretion complex that formed far to the south and has been translated along margin-parallel strike-slip faults as far north as the great bend of the southern Alaska margin, where it is colliding with previously accreted rocks (Plafker et al. 1994; Perry et al. 2009).

The subduction-accretion complex of southern Alaska is intruded by near-trench felsic to mafic plutons and dykes (Figs. 12, 13e, f). The plutonic rocks are calc-alkaline granodiorite, tonalite, quartz diorite and granite, with minor gabbro (Harris et al. 1996; Bradley et al. 2003). Dykes display a wide range of compositions ranging from basaltic to rhyolitic (Lytwyn et al. 2000; Bradley et al. 2003). The origin of these rocks has been attributed to melting of the sedimentary rocks in the accretionary prism and underplating of mantle-derived mafic rocks during ridge subduction (Plafker et al. 1989; Barker et al. 1992; Harris et al. 1996). Emplacement of the plutons was controlled mainly by structures developed on the overriding plate during ridge subduction (Kusky et al. 2003). Near-trench magmatism in southern Alaska started at 61 Ma at Sanak Island in the west and reached Baranof Island in the east at ca. 50 Ma; this diachronous trend is consistent with a west to east migration of a trench-ridge-trench triple junction for an along-strike length of 2,200 km along the convergent margin (Bradley et al. 2003; Farris and Paterson 2009). In the Kenai Peninsula where our samples were collected, the igneous intrusions yield ages between 58 and 53 Ma (Bradley et al. 2000, 2003).
Petrography

The Upper Cretaceous Valdez Group greywackes in the Alaskan accretionary complex are composed of 20-30% fine-grained matrix (organic-rich clays, chlorite) and 70-80% silt- to sand-sized angular grains (Fig. 14). The grains are composed of 65-75% lithic, 15-20% quartz, and 10-20% feldspar fragments (Fig. 14). The Paleogene felsic dykes consist of quartz (30-40%), plagioclase (40-50%), amphibole (10-15%), chlorite (5-8%), calcite (4-6%), sericite (2-4%), and epidote (2-4%) (Fig. 14).

Major and trace elements

Felsic dykes

The Alaskan felsic dykes are dacitic in composition and have small variations in SiO$_2$ (67-70 wt.%), TiO$_2$ (0.19-0.21 wt.%), Fe$_2$O$_3$ (1.8-2.3 wt.%), Al$_2$O$_3$ (15.1-15.8 wt.%), and Sc (4-5 ppm), and moderate to large variations in MgO (1.4-2.2 wt.%), CaO (1.3-3.0 wt.%), Na$_2$O (4.3-5.8 wt.%), and K$_2$O (0.97-1.79 wt.%). They possess high Mg-numbers (56 to 69) but low transition metals (Ni=22-64 ppm; Cr=51-84 ppm; V=16-26 ppm). The ratios of Al$_2$O$_3$/TiO$_2$ (73-87), Zr/Y (14.3-20.1), and Sr/Y (86-178) are higher than those of the upper continental crust. Nb/Ta (16.5-21.0) ratios are sub-chondritic to super-chondritic, whereas Zr/Hf (29-30) ratios are sub-chondritic.

On the N-MORB-normalized diagram, the felsic dykes have the following trace element characteristics: (1) moderately enriched LREE (La/Sm$_{cn}$=2.08-2.74; La/Yb$_{cn}$=6.74-9.64) and depleted HREE (Gd/Yb$_{cn}$=2.12-2.25) patterns; (2) small negative to positive Eu anomalies (Eu/Eu*=0.88-1.09) and zero Ce anomalies (Ce/Ce*=0.99-1.02); (3) large negative Nb (Nb/Nb*=0.17-0.21) and Ti (Ti/Ti*=0.46-0.48) anomalies; and (4) positive Zr (Zr/Zr*=1.36-2.21) and Pb (Pb/Pb*=17.5-55.3).

Greywackes

Greywackes from the Upper Cretaceous Valdez Group have uniform SiO$_2$=66.6-69.8 wt.% contents; sample AL2006-30 has significantly lower SiO$_2$ (59.1 wt.%) content than the other samples. They have 1.96-3.11 wt.% MgO, 0.60-0.82 wt.% TiO$_2$, 5.4-7.4 wt.% Fe$_2$O$_3$, 1.40-3.84 wt.% CaO, 12.4-16.4 wt.% Al$_2$O$_3$, 0.77-2.52 wt.% K$_2$O and 1.62-3.96 wt.% Na$_2$O contents. The greywackes are characterized by moderate ranges of Ni (30-60 ppm), Cr...
(80-150 ppm), V (114-163 ppm), Co (13-19 ppm), Sc (12-16 ppm), Ba (420-997 ppm), and Sr (200-289 ppm). The ratios of \( \text{Al}_2\text{O}_3/\text{TiO}_2 \) (18-23) are slightly sub-chondritic, whereas Zr/Y (5.3-8.3) ratios are strongly super-chondritic (Table 4).

On the N-MORB-normalized diagram, they have the following trace element characteristics: (1) enrichment of LILE (e.g., Rb, Ba, K) relative to REE and HFSE; (2) enrichment of LREE (La/Sm\text{cn}=3.01-3.66; La/Yb\text{cn}=5.93-7.96; Gd/Yb\text{cn}=1.33-1.67) relative to HREE; (3) variably negative Eu (Eu/Eu*=0.75-0.91) anomalies; (4) strongly negative Nb (Nb/Nb*=0.15-0.20) and Ti (Ti/Ti*=0.40-0.50) anomalies; and (5) large positive Pb (Pb/Pb*=6.9-12.6) anomalies (Fig. 15).

**Discussion**

**Zircon age constraints on the Tårtoq greenstone belt**

The three samples of the felsic mylonite from the Iterlak fragment of the Tårtoq greenstone belt (Fig. 1) dated in this study yield ages of 2993±6 Ma, 2997±10 Ma and 3012±4 Ma. Whereas the former two samples (510312 and 510328) have identical ages within error, the third sample (510325) is ca. 20 million years older (outside analytical error). The younger age agrees well with previously reported ages from intrusive TTG sheets from the northern contact in the Amitsuarsua fragment (Fig. 1) that gave 2986±4 Ma and 2996±6 Ma (Szilas et al. 2013). An intrusive TTG sheet in the southern part of the Iterlak fragment was dated by Nutman and Kalsbeek (1994) providing a zircon U/Pb age of 2944±7 Ma. Compared to the above mentioned ages, sample 510325 provides an older age (3012±4 Ma), suggesting a prolonged time span of felsic magmatism in the larger Tårtoq region from 3012 to 2944 Ma. These ages are interpreted to reflect a long-lived (3012-2944 Ma) subduction zone activity in the region. The formation age of the Tårtoq greenstone belt has been inferred at c. 3190 Ma based on whole-rock Lu-Hf isotope systematics (Szilas et al. 2013) who reported whole-rock Lu–Hf and Sm–Nd errorchron ages of 3189±65 Ma and 3068±220 Ma, respectively, for mafic samples from different fragments in the belt. These ages either reflect open system behaviour of the Lu–Hf and Sm–Nd isotopes during metamorphism, or rocks from different fragments that are not cogenetic or coeval. Therefore, based on the 2993±6, 2997±10 and 3012±4 Ma zircon ages reported in this study, we suggest that most reliable minimum age of the Tårtoq greenstone belt is about 3012 Ma.

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Effects of metamorphism on element mobility

Petrographic observations indicate that the Upper Cretaceous greywackes of the Valdez Group and Paleogene felsic dykes underwent lower greenschist facies metamorphism (Fig. 14). Both the greywackes and felsic dykes display coherent LILE, REE and HFSE patterns and near zero Ce anomalies, suggesting that these elements were relatively immobile during metamorphism (Fig. 15).

The Târtoq greenstone belt underwent at least four phases of deformation (D1-D4) and greenschist to upper amphibolite facies regional metamorphism (Kisters et al. 2012; Szilas et al. 2013, 2014). It is therefore important to understand the effects of metamorphism on the chemical composition of the Târtoq rocks. The effects of metamorphism on element mobility in the Târtoq greenstone belt is discussed in detail by Szilas et al. (2013), based on a large number of samples collected from several areas. Interested readers are referred to this study. Therefore, we provide only a short discussion on this subject using our samples from the Iterlak area (Fig. 1).

The presence of quartz and calcite veins in all rock types in the Târtoq belt suggests that they all underwent silica and carbonate alteration (Figs. 5, 6). The meta-gabbro, greenschist and felsic mylonite samples display Ce anomalies (Ce/Ce*=0.94-1.06) close to unity (1.00) and coherent chondrite-normalized REE and N-MORB-normalized multi-element patterns, indicating that these elements were relatively immobile during metamorphism (cf., Polat and Hofmann, 2003; Fig. 11). One felsic dyke sample (510331) has a minor negative Ce (Ce/Ce*=0.88) anomaly (Tables 2, 3 and 4), consistent with disturbance of REEs in this sample (Fig. 11).

Geodynamic setting

Field and geochronological constraints

The Iterlak fragment of the Târtoq greenstone belt consists of tectonically interleaved greenschists and felsic mylonites, with several hundred meter long lenses of meta-gabbro, serpentinite, and banded iron formation (Figs. 1-3). Contacts between the felsic mylonites and greenschists are marked mainly by thrust shear zones and well-preserved small-scale duplex structures (Figs. 3, 4). Although the whole belt consists of numerous thrust fault-bounded tectonic slices, duplex structures are best developed at the contacts between the greenschists and the felsic mylonites where rocks with contrasting competence occur (Fig. 4). These duplex structures and major shear zones in the Iterlak area display consistently top-to-the east-southeast
tectonic transport (Figs. 3, 4). Given that the felsic mylonites are marked by thrust-imbricate structures, they are interpreted as syn-tectonic intrusions emplaced along thrust shear zones (Fig. 16). Their thickness varies from twenty centimeters to several tens of meters. Intrusion of TTGs along thrust zones during contractional tectonic events was a common process in the Archean (Myers 1976; de Wit et al. 1987). Many shear zones in the belt are marked by several meters thick, strongly deformed, discontinuously exposed carbonate lenses, reflecting their precipitation from carbonic fluids migrating along shear zones (Fig. 3). In addition, calcite veinlets are present throughout the belt, particularly along foliation planes (see also Evans and King 1993; Kisters et al. 2012). These carbonates are consistent with their precipitation from syn-tectonic hydrothermal fluids. Carbonate-bearing shear zones also contain mesozonal orogenic gold occurrences (Kolb et al. 2013).


On the basis of field observations, Kisters et al. (2012) suggested that the Târtoq greenstone belt formed as a subduction-accretion complex along an obliquely convergent margin. During subduction some part of the Târtoq greenstone rocks were buried to depths >25 km and underwent upper amphibolite facies metamorphism and ductile deformation (D1 and D2) (Kisters et al. 2012). High-grade metamorphic rocks were retrogressed during later ductile to
brittle deformation (D3 and D4) and juxtaposed with lower-grade metamorphic rocks. Melting of accreted oceanic crust that resulted in the intrusion of TTGs in the Târtoq region is attributed to higher temperature conditions in hotter Archean subduction zones (Kisters et al. 2012). However, crustal geothermal gradients in the Archean might not have been steeper than those of the modern Earth (see Burke and Kidd 1978; Burke et al. 1986). The average temperature of the mantle in the Archean was probably similar to that of the modern mantle. Komatiites with higher eruption temperatures (1450-1600 °C) are likely have been derived from mantle plumes originating at the core-mantle boundary (Campbell 2007; Burke et al. 2008; Nebel et al. 2014 and references therein). More heat production in the Archean Earth was likely buffered by greater melting of the shallow mantle to make thicker oceanic crust. When more heat was being generated in the Archean mantle and core, it was probably dissipated simply by faster mantle convection, generating smaller or faster moving plates. Therefore, Archean subduction zones were not necessarily characterized by hotter conditions than the present counterparts as often assumed (see Polat and Kerrich 2006). We speculate that the protoliths of the felsic mylonites alternatively would have originated from partial melting of accreted oceanic crust in response to ridge subduction and were emplaced as syn-tectonic intrusions along thrust faults in the accretionary prism (Fig. 16) (cf., Bradley et al. 2003; Sisson et al. 2003a; Xu et al. 2015).

The new ages (3012 to 2993 Ma) obtained by this study are interpreted to represent the time of subduction, imbrication and melting of Mesoarchean oceanic crustal fragments in the subduction channel during the D2 deformation phase. The time period between 3012±4 and 2993±6 Ma may reflect continuous ridge subduction or slab window formation over 15-20 million years. Kisters et al. (2012) suggested that felsic melts intruded along D2 shear zones may have acted as lubricant for return flow of the exhuming rocks in the subduction-accrretion complex.

Geochemical constraints

On the basis of trace element systematics of a large number of meta-volcanic rock (e.g., amphibolite, greenstone, greenschist) and meta-gabbro samples, Szilas et al. (2013) proposed that the Târtoq greenstone belt formed in a supra-subduction setting. Using geochemical data from the ultramafic rocks (serpentinites) in conjunction with field observations and the previously published geochemical data, Szilas et al. (2014) interpreted the Târtoq greenstone belt
as a subduction-related Archean ophiolite. Although the Târtoq greenstone belt lacks several major components of the classical Penrose-type ophiolite (Anonymous 1972), such as a sheeted dyke complex, layered gabbros and layered ultramafic rocks, we agree with Szilas et al. (2014) that the Târtoq greenstone belt can be defined as a supra-subduction zone ophiolite on the basis of the new ophiolite classification proposed by Furnes et al. (2015). Several studies suggested that the classical Penrose-type ophiolite definition should not be used as a guide to define ophiolites in poly-deformed and metamorphosed Precambrian orogenic belts, because this definition is not broad enough to encompass the major geological characteristics of dismembered oceanic lithosphere in these orogenic belts (Kusky 2004; Şengör and Natal’in 2004; Dilek and Polat 2008; Dilek and Furnes 2011; Furnes et al. 2015; Polat et al. 2015).

The geochemical data presented in this study for the greenschist, meta-gabbro and felsic mylonite in the Iterlak fragment are collectively consistent with a supra-subduction (arc, fore-arc or back-arc) geodynamic setting (Fig. 16; Table 3), supporting the interpretation of Szilas et al. (2013, 2014). Depletion of HFSE (Nb, Ti) relative to REE in the Târtoq greenschist rocks (meta-basalts) and meta-gabbros indicate that their mantle source was affected by subduction-derived melts or fluids. The Târtoq greenschist samples share trace element characteristics of modern back-arc basalts, such as the Lau Basin, Manus Basin and South Sandwich, and basalts from the Eocene East Sulawesi ophiolite (Fig. 17a, b) (Monnier et al. 1995; Gamble and Wright 1995; Hawkins 1995; Kadarusman et al. 2004; Leat et al. 2004; Pearce and Stern 2006). Many ophiolites and subduction-accretion complexes in Phanerozoic orogenic belts are interpreted as remnants of closed back-arc basins (Şengör and Yilmaz 1981; Şengör et al. 1984; Hsü 1988). Similarly, a number of Archean greenstone belts are interpreted as remnants of back-arc oceanic crust (Tarney et al. 1976; Hollings and Kerrich 2000; Kerrich et al. 2008; Polat 2009; Manikyamba et al. 2015).

The geochemical characteristics and petrogenesis of Archean TTGs and Cenozoic adakites have been extensively debated in the literature over the past thirty years (Martin 1987, 1999; Drummond and Defant 1990; Drummond et al. 1996; Bourdon et al. 2002; Foley et al. 2002; Martin and Moyen 2002; Guivel et al. 2003; Rapp et al. 2003; Condie 2005; Martin et al. 2005, 2014; Steenfelt et al. 2005; Richards and Kerrich 2006; Wang et al. 2006; Chiaradia et al. 2009; Moyen 2009; Adam et al. 2012; Nagel et al. 2012). Given their compositional diversity and
occurrence in various tectonic settings, the petrogenesis of both Archean TTGs and Cenozoic adakites still remains controversial (Martin et al. 2005, 2014; Moyen 2009; Kon et al. 2013).

The origin of Archean TTGs has been attributed to diverse petrogenetic and geodynamic processes, including melting of the sub-arc mantle peridotite, subducted slabs, the base of oceanic plateaus, accreted oceanic plateaus, and amphibolites in thickened arcs (Foley et al. 2002; Condie 2005; Steenfelt et al. 2005; Richards and Kerrich 2006; Moyen 2009; Martin et al. 2005, 2014; Hoffmann et al. 2014; Kamber 2015). On the basis of major and trace element data, earlier studies commonly interpreted Archean TTGs as product of partial melting of subducted oceanic crust, as Cenozoic adakites (Drummond and Defant 1990; Martin 1987, 1999). However, several later studies (Smithies 2000; Condie 2005; Martin et al. 2014) have revisited this interpretation and argued that Archean TTGs are not compositional equivalent of slab-derived Cenozoic adakites, suggesting that these two rock associations likely have different petrogenetic origins. Therefore, most Archean TTGs cannot be considered as analogs of Cenozoic adakites representing slab melts that reacted with sub-arc mantle peridotite during their descent.

Geochemically, adakites are defined on the basis mainly of \( \text{SiO}_2 \), \( \text{MgO} \), \( \text{Na}_2\text{O} \), \( \text{Al}_2\text{O}_3 \), \( \text{Ni} \), \( \text{Cr} \), \( \text{Sr} \), \( \text{Y} \) and \( \text{Yb} \) contents, Mg-number, and \( \text{K}_2\text{O}/\text{Na}_2\text{O} \), \( \text{Sr}/\text{Y} \) and \( \text{La}/\text{Yb} \) ratios (Martin 1999; Moyen 2009). Martin et al. (2005) divided adakites into high-silica (\( \text{SiO}_2 > 60 \) wt.%)) and low-silica (\( \text{SiO}_2 < 60 \) wt.%) groups. High-silica adakites are interpreted as slab melts that interacted with sub-arc mantle peridotite during their ascent, whereas the origin of low-silica adakites is attributed to melting of sub-arc mantle peridotite that has been metasomatized by adakitic melts.

Some Cenozoic adakites are spatially associated with high-Mg andesites and Nb-enriched basalts (see Polat and Kerrich 2006; Kerrich and Polat 2006; Ickert et al. 2009 and references therein). These rock associations are documented in arcs associated with subduction of young, hot oceanic lithosphere, ridge subduction, or slab windows (Defant et al. 1992; Yogodzinski et al. 1995; Kelemen 1995; Kepezhinskas et al. 1996; Ickert et al. 2009; Benoit et al. 2002; Kon et al. 2013). In this association, adakites are interpreted as slab-derived melts that variably interacted with sub-arc mantle peridotites, whereas high-Mg-andesites are products of more extensive interaction with sub-arc mantle peridotite, and Nb-enriched basalts that originated from the residue of interaction that melted at greater depths than high-Mg andesites (Defant et al. 1992; Yogodzinski et al. 1995; Kepezhinskas et al. 1996; Benoit et al. 2002; Kerrich and Polat 2006). On the basis of major and trace element, and Nd isotope data, Ickert et al. (2009)
attributed the origin of Eocene adakitic volcanic rocks in south-central British Columbia to melting of Mesozoic arc-derived mafic dykes and sills above a slab window. In this model, the mafic dykes and sills were emplaced into the lithospheric mantle beneath British Columbia. The origin of some adakites within continents is attributed to melting of delaminated lower mafic crust (Xu et al. 2002; Gao et al. 2004; Wang et al. 2006).

Kon et al (2013) showed that the late Miocene to Pliocene Taitao granitoids from Chile share the major element compositions of Archean TTGs, despite their less fractionated HREE patterns. The origin of these granitoids is attributed to melting of the subducted oceanic crust at garnet-free amphibolite conditions, at depths shallower than 30 km, suggesting that highly HREE depleted patterns in some TTGs should not be considered as unique evidence of subducted slab melting.

How did the protolith(s) of the Târtoq felsic mylonites form? The following six possible geological processes might have been responsible for their origin: (1) melting of subducted MORB-like oceanic crust (cf., Defant and Drummond 1990; Martin 1999); (2) melting of sub-arc mantle peridotite followed by fractional crystallization; (3) melting of delaminated lower crust (cf., Kay and Kay 1993; Xu et al. 2002; Gao et al. 2004); (4) melting of basaltic dykes or sills that were emplaced into sub-arc mantle peridotite (Ickert et al. 2009); (5) melting of mafic crust in a thickened arc (Adam et al. 2012); and (6) melting of accreted MORB or supra-subduction zone oceanic crust (e.g., arc, back-arc, fore-arc) in response to ridge subduction or generally higher Archean geothermal gradients.

On the basis of the following field observations and geochemical constraints, the first five processes are unlikely to account for the generation of the Târtoq felsic mylonites: (1) the protoliths of the Târtoq mylonites were emplaced into a subduction-accretion complex between a trench and magmatic arc (Kisters et al., 2012; this study), rather than emplaced into an arc crust; and (2) higher SiO₂ (by 64-71 wt.%), and lower Mg-numbers (29 and 57), and Cr (5-133 ppm) and Ni (9-71 ppm) contents, are inconsistent with slab-melt mantle peridotite interaction (Table 2). Accordingly, melting of accreted MORB or supra-subduction zone oceanic crust appears to be the most likely mechanism.

In order to further investigate this, the major and trace element compositions of the Târtoq felsic mylonites are compared with those of the Eocene adakitic dacites in south-central British Columbia, Canada, and the Miocene to Pliocene TTGs in the Taitao Peninsula, Chile
(Supplementary Figs. 1, 2). The Taitao TTGs are spatially and temporally associated with the subduction of the Chile ridge and derived from partial melting of subducting MORB (Kon et al., 2013). The Eocene British Columbia adakitic dacites stemmed from slab window formation (Ickert et al. 2009). On the SiO$_2$ versus MgO, TiO$_2$ and Al$_2$O$_3$, and La/Nb$_{morb}$ versus Th/Yb$_{morb}$ and La/Sm$_{cn}$ versus Th/Nb$_{morb}$ diagrams, the Târtoq samples plot mainly in the same field as the Taitao TTGs, whereas on the Y versus Zr and La/Sm$_{cn}$ versus Gd/Yb$_{cn}$ diagrams they plot between the British Columbia and Taitao samples. The Târtoq samples have higher Gd/Yb$_{cn}$ ratios than the Taitao samples, signifying subduction-derived, HREE-depleted source rocks, rather than MORB-like source rocks with flat HREE patterns. These geochemical characteristics suggest that the Târtoq felsic mylonites are unlikely to have been derived from partial melting of subducted MORB, as the Taitao TTGs, or from partial melting of the mafic dykes and sills in a sub-arc mantle wedge, as the Eocene British Columbia adakitic dacites.

Although on the Yb versus La/Yb diagram, the Târtoq felsic mylonites plot within the fields of TTGs and high-silica adakites (Fig. 17c), the absence of geochemical evidence for the interaction with mantle peridotite suggests that they do not meet the criteria of the term “adakite” defined by Defant and Drummond (1990) and Moyen (2009). Similarly, the Paleogene Alaskan dykes plot in the field of high-silica adakites but lack the chemical evidence of mantle interaction. The Mesoarchean Târtoq TTGs and the Paleogene Alaskan adakites have lower La/Yb and Gd/Yb ratios than most Archean TTGs and high-silica adakites, reflecting the shallow partial melting depths (<80 km) for the former rocks (cf., Herzberg 1995; Kon et al., 2013).

A number of recent studies postulated that ridge subduction played a major role in the petrogenesis of Phanerozoic adakites (Aguillón-Robles et al. 2001; Tang et al. 2010; Eyuboğlu et al. 2011, 2012, 2013; Xu et al. 2015). However, other studies on Phanerozoic adakites and Archean TTGs suggest diverse petrogenetic and geodynamic processes for the origin of these rocks, including melting of amphibolites in thickened arcs, and melting of subducted slabs, oceanic plateaus, sub-arc mantle peridotites, and delaminated lower continental crust (Richards and Kerrich 2006; Wang et al. 2006; Danyushevsky et al. 2008; Chiaradia et al. 2009; Moyen 2009; Hoffmann et al. 2014; Adam et al. 2012; Huang et al., 2013; Martin et al. 2014). Martin et al. (2014) argued that melting of subducted MORB and the base of thick oceanic plateaus are not realistic processes to account for the origin Archean TTG. These studies indicate that assigning
the origin of all adakites to slab melting and slab melt interaction with sub-arc mantle peridotites can be misleading.

**Tectonic evolution**

On the basis of field observations, geochemical and geochronological data, we propose the following geodynamic scenario for the Târtoq greenstone belt and spatially associated TTG gneisses:

1. The lithotectonic assemblage of metamorphosed basalt, gabbro and peridotite in the Târtoq greenstone belt formed in a back-arc basin behind either an oceanic island arc (e.g., Tonga-Kermadec) or a continental island arc (e.g., Japan). The mantle source of the back-arc oceanic crust was contaminated by subduction-derived chemical components, resulting in high Th/Nb, La/Nb, Pb/Ce and Sm/Ti ratios (Fig. 11). The presence of old recycled zircon grains (e.g., 3109, 3173, 3338, 3646, 3771 Ma) in the Târtoq meta-sedimentary rocks (Nutman et al. 2004) suggests that the island arc contained fragments of older crust. Alternatively, these zircons were transported from continents to the back-arc basin.

2. In order to explain the presence of back-arc oceanic crust fragments in the Târtoq subduction-accretion complex and intrusion of the complex by felsic rocks, we speculate that a subduction zone was initiated at the transition between the back-arc basin and island arc in response to either arc-arc or arc-continent collision (Fig. 16) (c.f., McKenzie 1969; Suppe 1984; Wang et al. 2015).

3. During the initiation of the new subduction zone, basalt, gabbro and peridotite from the subducted back-arc oceanic crust were peeled off and accreted to the overriding plate, forming the subduction-accretion complex represented by the Târtoq greenstone belt. Deeply buried (>25 km) back-arc oceanic crust melted in response to a ridge subduction resulting in thermal anomaly in the accretionary prism, producing TTG melts (Fig. 16). These melts migrated upward along syn-tectonic detachments, D2 shear zones of Kisters et al. (2012), producing felsic mylonites with small duplex structures.

**Comparison of the Târtoq and Alaskan felsic rocks**

Like the Mesoarchean Târtoq felsic mylonites, the Paleogene Alaskan felsic dykes are similar to high-silica adakites in terms of REE compositions (Fig. 17c; Tables 2 and 3; see...
Martin et al. 2005). Despite their similarities in major element compositions, there are, however, some geological and trace element differences between the Târtoq and Alaskan felsic rocks. The major geological differences between the two include: (1) the Târtoq felsic rocks were emplaced into an association of basaltic, gabbroic and ultramafic rocks, whereas the Alaskan counterparts were emplaced mainly into a siliciclastic sedimentary sequence; and (2) the Târtoq felsic rocks were emplaced as syn-tectonic intrusion along thrust faults, whereas the Alaskan counterparts were intruded into previously Mesozoic deformed rocks as dykes and sill.

Although the Alaskan and Târtoq felsic rocks have similar trace element patterns (Fig. 15b), the Alaskan samples have lower absolute abundances of REE, HFSE and LILE than the Târtoq counterparts, suggesting that they were derived from a more depleted source than the Târtoq samples (Fig. 15b). The origin of Paleogene granitic rocks in southern Alaska has been attributed to mixing of melts from two sources during ridge subduction: (1) accretionary sedimentary rocks and mafic rocks underplated the base of the accretionary prism (Barker et al. 1992; Harris et al. 1996; Sisson et al. 2003a; Bradley et al. 2003; Kusky et al. 2003). To assess the source characteristics of the Alaskan felsic dykes, we analysed the spatially associated greywacke turbidites from the Valdez Group accretionary complex. The petrographic (65-75% lithic fragments, 15-20% quartz grains, and 10-20% feldspar grains) and geochemical (Figs. 14, 15) characteristics of the Valdez turbidites are consistent with a continental arc source. As shown in Tables 3 and 4 and Fig. 15a, felsic dykes have significantly lower abundances of LILE, REE and HFSE than the Valdez Group greywacke turbidites, suggesting that the felsic dykes could not have been derived from partial melting of the greywacke turbidites. Accordingly, we suggest that the Alaskan felsic dykes were derived from partial melting of accreted or subducted oceanic crust during ridge subduction in the Paleocene to Eocene. However, partial melting of underplated mafic rocks (gabbros) cannot be ruled out as the source of the felsic dykes.

**Implications for Archean continental growth**

Growth and destruction of continental crust at present occurs predominantly at convergent plate margins through magmatic addition and tectonic accretion, and tectonic erosion, respectively (Şengör and Natal’în 1996; Yamamoto et al. 2009; Scholl and Huene 2010; Kusky et al. 2013). Numerous studies suggest that the geological characteristics of Archean granitoid-greenstone terrains necessitate horizontal tectonic movements involving collision of oceanic
island arcs, continental blocks and oceanic plateaus at Phanerozoic-like convergent plate margins (de Wit 1998; Kusky and Polat 1999; Friend and Nutman 2005; Furnes et al. 2015). Because of high heat production in the mantle and core, Archean tectonics was probably dominated by numerous smaller plates and greater ridge length than at present-day (Hargraves 1986; Ernst et al. 2015). Further consequences of the higher heat production in the Archean would be more melt production beneath ridges that would have produced a thicker oceanic crust (maybe 20-30 km instead of 7 km).

Several recent studies proposed that the time period between 3.2 and 3.0 Ga in Earth’s history marks the onset of Phanerozoic-type plate tectonics (Dewey 2007; van Kranendonk 2011; Shirey and Richardson 2011; Dhuime et al. 2015; Lee and McKenzie 2015; Tang et al. 2016). Neodymium and Hf isotope data obtained from Archean granitoids suggest that the 3.2-3.0 Ga time period in Earth’s history also represents a maximum rate of juvenile crust formation and crustal recycling into the mantle (Jacobsen 1988a, 1988b; Dhuime et al. 2012, Næraa et al. 2012). On the basis of Nd and Hf isotope data from the 3.48 Ga Barberton komatiites, Blichert-Toft et al. (2015) suggested that plate tectonic processes were in operation at about 3.5 Ga. Similarly, on the basis of nitrogen abundances, and nitrogen and carbon isotopic signatures in Archean placer diamonds from the Kaapvaal craton of South Africa, Smart et al. (2016) suggested that modern-style plate tectonic processes were active as early as 3.5 Ga. The trace element compositions of the Isua mafic to ultramafic rocks in South-West Greenland are consistent with the operation of subduction zone petrogenetic processes as early as 3.8 Ga (Polat et al. 2002; Furnes et al. 2007; Jenner et al. 2009). Field evidence from the Eoarchean terrains suggest that convergent plate margin processes were operative as early as 3.8 Ga (Turner et al. 2014; Komiya et al. 2015; Polat et al. 2015). Thicker Archean oceanic crust would have been more buoyant than oceanic crust today. Such crust might have had difficulties in subducting and rather ended up in “plate graveyards” i.e. sites of accreted plates (cf., Næraa et al. 2012), resulting in more extensive subduction-accretion complexes than post-Archean times. On the basis of density modelling of the oceanic lithosphere above a warmer Archean asthenosphere, however, Hynes (2014) showed that subduction of the oceanic lithosphere in the Archean was even slightly easier than it is today. Similarly, Rollinson (2010) argued that partial melting at Archean spreading ridges gave rise to Fe-rich basaltic crust and strongly depleted oceanic
lithosphere. This negatively buoyant basaltic crust underwent partial melting upon its subduction, producing TTG-dominated Archean continental crust at convergent plate margins.

There are striking geological similarities between Archean granitoid-greenstone terrains and Phanerozoic subduction-accretion complexes, including the Altaid and Alaskan accretionary complexes (Williams et al. 1991; Şengör and Natal’i in 1996; Kusky and Polat 1999; Polat et al. 2015). Melting of subducted oceanic crust, accretionary prisms, fore-arc crust or accreted oceanic plateaus may have played an important role in the growth of Archean continental crust. Archean subduction zones could also have been the main sites of crustal destruction (cf., Scholl and van Huene 2010). Nelson and Forsythe (1989) and Polat and Frei (2005) suggested that ridge subduction might have played a significant role in the growth of Archean continental crust. According to Nelson and Forsythe (1989), greater ridge length and smaller plates in the Archean probably resulted in more frequent ridge-trench collisions, contributing to the growth of the continental crust (Nelson and Forsythe 1989; Kusky and Polat 1999; Polat and Frei 2005; Yamamoto et al. 2009). Field evidence for ridge subduction has also been documented in Proterozoic orogenic belts (Garde and Hollis 2010). These data are consistent with studies showing that up to 80% of the volume of the present-day continental crust was built by 2.5 Ga (Taylor and McLennan 1995; Hawkesworth and Kemp 2006b; Hawkesworth et al. 2010). The small amount of preserved Archean continental crust, less than 10% of the present-day counterpart, likely reflects its recycling through plate tectonic processes.

Geological processes taking place in the Circum-Pacific belts involve initiation of intra-oceanic subduction zones, opening of back-arc basins, subduction reversals, oceanic ridge subduction, and accretion of island arcs and oceanic plateaus (Taira et al. 1992; Mann and Taira 2004; Şengör and Natal’in 1996; Sisson et al. 2003a). The structural, lithological and geochemical characteristics of Archean terrains are similar to those of the Circum-Pacific belts.

Paleogene interaction between a subducting oceanic ridge system and an overriding subduction-accretion complex resulted in a wide range of geological events in Alaska. The most obvious effect of this interaction is the intrusion of 61 to 50 Ma near-trench granitic to gabbroic plutons and dykes in the Alaskan accretionary prism (Bradley et al. 2003). Near-field effects of ridge subduction include anomalous magmatism in the trench and fore-arc region, ophiolite obduction, fore-arc deformation, and tectonic erosion of the fore-arc crust (Nelson and Forsythe 1989; Kusky et al. 1997; Bradley et al. 2003). Intrusion of the Mesozoic-Cenozoic subduction-
accretion complexes by granitoids in southern Alaska provides a good analog for granitoid intrusions in Archean granitoid-greenstone terrains. We postulate that ridge subduction caused partial melting of volcanic rocks in the Târtoq greenstone belt, resulting in intrusion of the precursor rocks of the felsic mylonites (Fig. 16).

Conclusions

On the basis of field observations, whole-rock major and trace element data, and zircon U-Pb dating, the following petrologic and geodynamic conclusions are drawn for the Mesoarchean Târtoq greenstone belt, South-West Greenland.

1. The Târtoq greenstone belt is composed of tectonically imbricated greenschist (metabasalt), meta-gabbro, serpentinite, banded iron formation, and siliciclastic sedimentary rocks. The belt displays well-preserved small-scale duplex structures that, together with the oceanic type rock units, are Archean analogs of duplex structures in Phanerozoic subduction-accretion complexes, suggesting that the belt originated as a subduction-accretion complex.

2. The greenschist, meta-gabbro and serpentinite are intruded by 3012±4 to 2993±6 Ma felsic rocks (SiO$_2=64$-71 wt.%; Al$_2$O$_3=14.1$-18.8 wt.%; La/Yb$_{cn}=6.4$-18.9) along thrust shear zones, forming felsic mylonite.

3. Greenschist, meta-gabbro, serpentinite and felsic mylonite are all characterized by supra-subduction zone (arc, fore-arc or back-arc) trace element compositions. To explain the geochemical characteristics and field relationships, we propose that the Târtoq greenstone belt originated in a Mesoarchean back-arc basin. It is suggested that following a reversal in subduction polarity, the back-arc basin oceanic crust started to subduct beneath the arc, forming a subduction-accretion complex.

4. The Upper Cretaceous turbiditic greywackes of the Chugach accretionary complex in southern Alaska are intruded by Paleogene felsic dykes (SiO$_2=67$-70 wt.%; Al$_2$O$_3=15.1$-15.8 wt.%; Sr/Y=86-178; La/Yb$_{cn}=6.7$-9.6). The greywackes were derived from continental arc rocks and the felsic dykes were derived from partial melting of either subducted or accreted oceanic crust, or underplated gabbro, during oceanic ridge subduction. Both the Mesoarchean Târtoq felsic mylonite and Paleogene Alaskan felsic dykes are similar to high-silica adakites in terms of their major element compositions (cf.,
Martin et al. 2005). However, their trace element characteristics are different from those of high-silica adakites.

5. The Mesoarchean Greenlandic and the Paleogene Alaskan felsic rocks have similar trace element patterns. However, the Greenlandic felsic rocks have higher abundances of REE and HSFE than the Alaskan counterparts, suggesting that the Greenlandic rocks were derived from more enriched (back-arc basalts) sources than the Alaskan rocks (MORB-like).

6. The geological characteristics of the Târtoq greenstone belt are similar to those of Phanerozoic subduction-accretion complexes such as the Circum-Pacific and Altaid subduction accretion complexes, consistent with the operation of uniformitarian geological processes in the Archean.

7. Higher heat production in the Archean likely resulted in the formation of thicker (20-30 km instead of 7 km) and more buoyant Archean oceanic crust than at present-day. Such thick crust might have had difficulties in subducting and rather ended up in the formation of extensive subduction-accretion complexes. Partial melting of these subduction-accretion complexes in response to more frequent ridge subduction, due to greater ridge length, may have played an important role in the production of Archean continental crust.

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Figure Captions

**Fig. 1.** (a) Location of the study area in South-West Greenland. (b) Simplified geological maps of the Târtoq region (modified from Berthelsen and Henriksen 1975 and Kisters et al. 2012) and (c) the Iterlak area (modified after Jensen 1973). Large fragments of the Târtoq greenstone belt are exposed in the Iterlak, Amitsuarsua, Nuuluk and Bikuben areas.

**Fig. 2.** Field photographs of the major rock types in the Târtoq greenstone belt: (a) medium-grained meta-gabbro outcrop; (b) metamorphosed pillow basalt with core and rim structures; (c) strongly sheared greenschist; (d) greenschist displaying isoclinal folds; (e) strongly deformed greenschist and felsic mylonite; and (f) greenschist containing sheared and fragmented felsic mylonite layers and cut by a felsic dyke (aplite). Figures (a), (b) and (d) are modified from Polat et al. (2015).

**Fig. 3.** Field photographs of the major rock types and thrust structures in the Târtoq greenstone belt: (a) contact between the TTG gneisses and the greenstone belt containing several meters to several tens of meters thick tectonic slivers of the gneisses within the greenstone belt; (b) and (c) tectonically intercalated felsic mylonite and greenschist; (d) a carbonate lens in greenschist; the carbonates occur as 1-3 meters thick and 1-50 meters long, discontinuous lenses along the thrust faults; and (e) and (f) tectonically intercalating felsic mylonite and greenschist. Figures (d), (e) and (f) are modified from Polat et al. (2015).

**Fig. 4.** Field photographs of small-scale duplex structures in the Iterlak part of the Târtoq greenstone belt, showing syn-tectonic intrusion of dacitic melts (now felsic mylonite) along thrust faults. These duplex structures are consistent with top-to-the south-southeast tectonic transportations.

**Fig. 5.** Photomicrographs illustrating the petrographic characteristics of the Târtoq greenstone belt: (a) meta-gabbro; (b-d) greenschist; and (e) and (f) felsic mylonite (amp: amphibole; chl: chlorite; qtz: quartz; zoi: zoisite; plg: plagioclase).
Fig. 6. Scanning electron microscope (SEM) backscatter electron (BSE) images showing the petrographic characteristics of the Târtoq greenstone belt: (a) and (b) meta-gabbro; (c) greenschist; (d) and (e) felsic mylonite; and (f) felsic dyke (amp: amphibole; chl: chlorite; qtz: quartz; tit: titanite; zoi: zoisite; apt: apatite; cal: calcite; rut: rutile; bio: biotite; mus: muscovite; kfl: K-feldspar).

Fig. 7. Back-scattered electron (BSE) SEM images of zircons from felsic mylonite sample 510312, 510325 and 510328 from the Târtoq greenstone belt. The white circles show LA-ICP-MS dating spots (30 µm diameter).

Fig. 8. Tera-Wasserburg plots showing the mean zircon U-Pb ages of felsic mylonite sample 510312, 510325 and 510328 from the Târtoq greenstone belt. Note that for 510312 and 510328 the light gray error ellipses are excluded from the calculation of the indicated age. Plotted on the right-hand side is for each sample the $^{207}$Pb/$^{206}$Pb age distribution curve.

Fig. 9. SiO$_2$ (wt.%), MgO (wt.%), TiO$_2$ (wt.%), Fe$_2$O$_3$ (wt.%), Al$_2$O$_3$ (wt.%), Na$_2$O (wt.%), and CaO (wt.%) variation diagrams for the meta-gabbro, greenschist, felsic mylonite, and felsic dyke in the Târtoq greenstone belt.

Fig. 10. SiO$_2$ (wt.%), Ni (ppm), Zr (ppm), La (ppm), Co (ppm), Th (ppm), and Nd (ppm) variation diagrams for the meta-gabbro, greenschist, felsic mylonite, and felsic dyke in the Târtoq greenstone belt.

Fig. 11. N-MORB-normalized trace element patterns for the meta-gabbro, greenschist, felsic mylonite, and felsic dyke in the Târtoq greenstone belt. Normalization values are from Sun and McDonough (1989).

Fig. 12. Geological map of south-central Alaska illustrating the accreted tectonic assemblages, near-trench magmatic rocks, and major structures (from Bradley et al. 2003).
**Fig. 13.** Field photographs of the major rock types in the Chugach-Prince William accretionary complex of southern Alaska: (a) and (b) late Cretaceous Valdez sedimentary rocks (greywacke turbidites); (c) pillow lavas in the McHugh complex; (d) mélange in the McHugh complex; and (e) and (f) Paleogene felsic dykes cutting across the Valdez sedimentary rocks (greywacke turbidites).

**Fig. 14.** Photomicrographs illustrating the petrographic characteristics of the Late Cretaceous Valdez sedimentary rocks (greywacke turbidites) (a-c) and Paleogene felsic dykes (d-f) in southern Alaska (qtz: quartz; fld: feldspar; chl: chlorite; lit: lithic grain; cal: calcite; plg: plagioclase).

**Fig. 15.** (a) N-MORB-normalized trace element patterns for the Late Cretaceous Valdez sedimentary rocks (greywacke turbidites) and Paleogene felsic dykes in the Alaska subduction-accretion complex. (b) Trace element patterns of the Mesoarchean felsic mylonites and Paleogene Alaskan felsic dykes. Normalization values are from Sun and McDonough (1989).

**Fig. 16.** A geodynamic model suggesting a possible origin of the Târtoq greenstone belt, implying that the basalt, gabbro and peridotite in the belt formed in a back-arc tectonic setting. Following the initiation of intra-oceanic subduction zone at the back-arc and arc transition, back-arc oceanic crust was imbricated and accreted to the overriding plate forming a subduction-accretion complex. Partial melting of the accreted oceanic crust generated the protolith of the felsic dykes. It is suggested that partial melting was facilitated by ridge subduction.

**Fig. 17.** (a) La/Sm$_{cn}$ versus Gd/Yb$_{cn}$ and (b) Nb/Yb versus Th/Yb variation diagram for the Mesoarchean Târtoq greenschist (metabasalts) and modern back-arc basin basalts. (c) Yb (ppm) versus La/Yb variation diagram for the Mesoarchean Târtoq felsic mylonites and the Alaskan Paleogene felsic dykes. (b) Modified from Pearce (2008) and (c) Fields from Moyen (2009).
Supplementary Fig. 1. SiO$_2$ (wt.%), TiO$_2$ (wt.%), and Al$_2$O$_3$ (wt.%), MgO (wt.%) versus SiO$_2$ (wt.) variation diagrams for the Tårtoq felsic mylonites, Alaskan felsic dykes, Eocene dacites from British Columbia (Ickert et al. 2009), and Late Miocene to Pliocene Chile (Taitao) TTGs (Kon et al., 2013).

Supplementary Fig. 2. (a) Y (ppm) versus Zr (ppm), (b) La/Nb$_{morb}$ versus Gd/Yb$_{cn}$, (c) La/Sm$_{cn}$ versus Gd/Yb$_{cn}$, (d) La/Nb$_{morb}$ versus Th/Yb$_{morb}$, (e) Zr/Y and TiO$_2$ and (f) La/Sm$_{cn}$ versus Th/Nb$_{morb}$ variation diagrams for the Tårtoq felsic mylonite, Alaskan felsic dykes, Eocene dacites from British Columbia (Ickert et al. 2009), and Late Miocene to Pliocene Chile (Taitao) TTGs (Kon et al., 2013). Subscript cn: Chondrite-normalized; and subscript morb: Average N-MORB. Normalization values are from Sun and McDonough (1989).
Orthogneisses

Greenschist ~500 m

Felsic mylonite

Greenschist ~100 m

Felsic mylonite

Greenschist ~2 m

Felsic mylonite

Greenschist ~50 m

Felsic mylonite
510312

#: 129
2984 ± 61 Ma
(103) Th/U = 0.40

#: 65
3006 ± 63 Ma
(97) Th/U = 0.33

#: 78
3031 ± 84 Ma
(94) Th/U = 0.38

#: 80
3069 ± 41 Ma
(95) Th/U = 0.40

510325

#144:
2988 ± 64 Ma
(93) Th/U = 0.36

#25:
2983 ± 57 Ma
(94) Th/U = 0.50

#180:
3033 ± 97 Ma
(102) Th/U = 0.53

#129:
3074 ± 67 Ma
(103) Th/U = 0.52

510328

#: 7
3001 ± 42 Ma
(100) Th/U = 0.51

#: 53
3024 ± 53 Ma
(93) Th/U = 0.29

#: 24
3020 ± 63 Ma
(99) Th/U = 0.43

#: 169
2994 ± 81 Ma
(95) Th/U = 0.36
(a) Meta-gabbro

(b) Greenschist Group 1

(c) Greenschist Group 2

(d) Felsic mylonite

(e) Felsic dyke
Felsic dyke in the Valdez sedimentary rocks

McHugh mélange complex

Felsic dyke

McHugh complex pillow basalts

Valdez sedimentary rocks

~1 meter
(a) Alaskan felsic dykes
Red Alaskan turbidites

(b) Alaskan felsic dykes
Red Tårtq felsic mylonites
(a) Accretionary prism, Magmatic arc, Back-arc basin, ESE Rifted margin

(b) Suture zone, Magmatic arc, Târtoq greenstone belt (accretionary prism), Felsic schist, Greenschist, Subducted oceanic ridge-mantle window
(a) Gd/Yb vs. La/Sm$_{cn}$

(b) Ti/Yb vs. Nb/Yb

(c) La/Yb vs. Yb (ppm)

Legend:
- N-MORB
- E-MORB
- Back-arc basalts
- Tarqoq greenschist
- OIB
- Archean TTGs
- High-silica adakites
- Archean adakites
- Normal arc rocks
- Tarqoq felsic mylonites
- Tarqoq felsic dykes
- Alaskan felsic dykes
Table 2. Major (wt.%) and trace (ppm) element compositions of the mafic to felsic rocks in the Tartoq greenstone belt.

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| Ratio   | La/Sm<sub>cn</sub> | Gd/Yb<sub>cn</sub> | La/Yb<sub>cn</sub> | Al<sub>2</sub>O<sub>3</sub>/TiO<sub>2</sub> | Zr/Y | Sr/Y | Zr/Hf | Nb/Ta | Ce/Ce* | Eu/Eu* | Nb/Nb* | Pb/Pb* | Zr/Zr* |
|---------|-------------------|-------------------|-------------------|------------------------|------|------|-------|-------|--------|--------|--------|--------|--------|--------|
|         | 2.18              | 2.12              | 6.74              | 84                     | 20.1 | 178  | 30.6  | 16.8  | 0.99   | 0.88   | 0.17   | 25.5   | 2.19   |
|         | 2.08              | 2.25              | 6.81              | 87                     | 20.1 | 145  | 28.9  | 16.5  | 1.03   | 0.89   | 0.17   | 55.3   | 2.20   |
|         | 2.74              | 2.23              | 9.64              | 73                     | 14.3 | 86   | 29.7  | 21.0  | 1.02   | 1.09   | 0.21   | 17.5   | 1.36   |

https://mc06.manuscriptcentral.com/cjes-pubs
\[
\begin{array}{cccc}
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|---|---|---|---|
| North | $60^\circ 46'.36.1''$ | $60^\circ 46'.36.1''$ | $60^\circ 46'.36.1''$ | \\
| West  | $148^\circ 41'.50.5''$ | $148^\circ 41'.50.5''$ | $148^\circ 41'.50.5''$ | \\
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Table 4. Major (wt.%) and trace (ppm) element composition of the siliciclastic sedimentary rocks (flysch) from the Upper Cretaceous Valdez Group, Alaska.

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Table 4. Major (wt.%) and trace (ppm) element composition of the siliciclastic sedimentary rocks (flysch) from the Upper Cretaceous Valdez Group, Alaska.

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149° 25' 11.2"
149° 25' 11.2"
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Table 3. Major (wt.%) and trace (ppm) element composition of the Paleogene felsic dykes, Alaska.

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Table 4. Major (wt.%) and trace (ppm) element composition of the siliciclastic sedimentary rocks (flysch) from the Upper Cretaceous Valdez Group, Alaska.

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https://mc06.manuscriptcentral.com/cjes-pubs
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Table 4. Major (wt.%) and trace (ppm) element composition of the siliciclastic sedimentary rocks (flysch) from the Upper Cretaceous Valdez Group, Alaska.

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Table 4. Major (wt.%) and trace (ppm) element composition of the siliciclastic sedimentary rocks (flysch) from the Upper Cretaceous Valdez Group, Alaska.

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https://mc06.manuscriptcentral.com/cjes-pubs
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Table 4. Major (wt.%) and trace (ppm) element composition of the siliciclastic sedimentary rocks (flysch) from the Upper Cretaceous Valdez Group, Alaska.

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