An overview of anorthosite-bearing layered intrusions in the Archaean craton of southern West Greenland and the Superior Province of Canada: implications for Archaean tectonics and the origin of megacrystic plagioclase

Ali Polat, Fred J. Longstaff, and Robert Frei

Department of Earth and Environmental Sciences, University of Windsor, Windsor, Canada; Department of Earth Sciences, The University of Western Ontario, London, Canada; Department of Geoscience and Natural Resource Management, University of Copenhagen, København, Denmark

ABSTRACT
Anorthosite-bearing layered intrusions are unique to the Archaean rock record and are abundant in the Archaean craton of southern West Greenland and the Superior Province of Canada. These layered intrusions consist mainly of ultramafic rocks, gabros, leucogabbros and anorthosites, and typically contain high-Ca (>An$_{40}$) megacrystic (2–30 cm in diameter) plagioclase in anorthosite and leucogabbrro units. They are spatially and temporally associated with basalt-dominated greenstone belts and are intruded by syn-to post-tectonic granitoid rocks. The layered intrusions, greenstone belts and granitoids all share the geochemical characteristics of Phanerozoic subduction zone magmas, suggesting that they formed mainly in a suprasubduction zone setting. Archaean anorthosite-bearing layered intrusions and spatially associated greenstone belts are interpreted to be fragments of oceanic crust, representing dismembered subduction-related ophiolites. We suggest that large degrees of partial melting (25–35%) in the hotter (1500–1600 °C) Archaean upper mantle beneath rifted arcs and backarc basins produced shallow, kilometre-scale hydrous magma chambers. Field observations suggest that megacrystic anorthosites were generated at the top of the magma chambers, or in sills, dykes and pods in the oceanic crust. The absence of high-Ca megacrystic anorthosites in post-Archaean layered intrusions and oceanic crust reflects the decline of mantle temperatures resulting from secular cooling of the Earth.

1. Introduction
Although they are volumetrically small, Archaean anorthosite-bearing layered intrusions are widespread both in western Greenland and the Superior Province of Canada (Ashwal, 1993, 2010; Ashwal & Bybee, 2017; Daigneault, ST-Julien, & Allard, 1990; Dymek & Owens, 2001; Gilbert, 2007; Hoffmann, Svaenhberg, Pizlozo, Scherstén, & Münker, 2012; Myers, 1985; Owens & Dymek, 1997; Peck, Messing, Halden, & Chandler, 1998; Percival et al., 2012; Windley & Garde, 2009; Windley, Herd, & Bowden, 1973; Zhou et al., 2016). Anorthosites are also a ubiquitous component of Proterozoic orogenic belts and Phanerozoic ophiolites (Table 1), and have been recovered from modern oceanic crust (Ashwal, 2010). Archaean anorthosites are distinct from Proterozoic and Phanerozoic counterparts in that they commonly contain equidimensional, rounded high-Ca (An$_{40}$-$>$An$_{53}$) plagioclase ranging from 2 to 30 cm in diameter, and are associated with layers of leucogabbro, gabbro, and ultramafic rocks (Ashwal, 1993; Ashwal & Bybee, 2017; Myers, 1985; Peck et al., 1998; Polat et al., 2009). These layered intrusions are typically associated with tholeiitic to calc-alkaline basalt-dominated volcanic rocks in greenstone belts, and are intruded mainly by syn- to post-tectonic tonalite-trondhjemite-granodiorite suites (TTGs) (Daigneault et al., 1990; Davis et al., 1988; Leclerc et al., 2011; Polat, Wang, & Appel, 2015; Zhou et al., 2016).

Despite their widespread occurrence, Archaean megacrystic anorthosites are among the least understood rock types in the Earth’s crust (Ashwal, 1993, 2010; Ashwal & Bybee, 2017; Phinney, Donald, & David, 1988). Several outstanding petrogenetic and geodynamic questions remain to be resolved to constrain their origin. These questions include: (1) How did anorthosites originate in the first place? (2) How were they emplaced into the crust? (3) What type of crust (oceanic versus continental) did they intrude? (4) Why do they have plagioclase crystals up to 30 cm in diameter? (5) What was the geodynamic setting(s) that produced these rocks? (6) Why are they rare in post-Archaean times?

CONTACT Ali Polat polat@uwindsor.ca
The supplemental data for this article can be accessed at https://doi.org/10.1080/09853111.2018.1427408

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Table 1. List of known anorthosite-bearing Phanerozoic ophiolites and their interpreted geodynamic setting.

| Ophiolite                          | Age             | Interpreted geodynamic setting          | References                  |
|------------------------------------|-----------------|-----------------------------------------|-----------------------------|
| Camaguey Ophiolite, Cuba           | Jurassic        | Suprasubduction zone                    | Henare et al. (2010)        |
| Mazhalik Ophiolite, Russia         | Ordovician      | Suprasubduction zone, island arc        | Borodina, Egorova, and Izokh (2004) |
| Zambales Ophiolite, the Philippines| Eocene          | Suprasubduction zone                    | Hawkins (2007)              |
| Talazhinskiy Ophiolite, Russia     | Paleozoic       | Suprasubduction zone                    | Yurichev and Chernysheva (2014) |
| Chilas Complex, Pakistan           | Jurassic - Cretaceous | Suprasubduction zone, island arc    | Khan, Jan, Windley, Tarney, and Thirlwall (1989) |
| Bay of Islands Ophiolite, Newfoundland (Canada) | Ordovician | Suprasubduction zone                    | Jia, Liu, Zhang, Zhang, and Yu (2003) |
| Zhaheba Ophiolite, China           | Cambrian-Ordovician | Suprasubduction zone             | Kite and Stoddard (1984)    |
| Halifax County Complex, North Carolina (USA) | Early Palaeozoic | Suprasubduction zone, island arc       | Zimer, Ballhaus, Muenker, & Marien, 2013 |
| Troodos Ophiolite, Cyprus          | Cretaceous      | Suprasubduction zone, backarc basin    | Faznia and Alizade (2013)    |
| Urmia Ophiolite, Iran              | Devonian        | Mid ridge to island arc                | Liu, Liu, Kroener, Windley, and Shi (2010) |
| Sonokler Ophiolite, China and Mongolia | Permian-Triasic | Suprasubduction zone, island arc       | Newman, Stolper, and Stern (2000); Hawkins (2007) |
| Kahnjou Ophiolite, Iran            | Jurassic-Cretaceous | Mid ocean ridge  | Arvin, Babi, Ghadami, Dargahi, and Ardekan (2005) |
| Mariana Arc, the Philippines       | Cenozoic        | Suprasubduction zone, backarc basin    | Peterson, Ryan, Becker, Cochrane, and Collins (2009) |
| Buck Creek Ophiolite, North Carolina (USA) | Paleozoic | Suprasubduction zone                    | Eyuboglu, Dilek, Bozkurt, Bektas, and Rojai (2010) |
| Karayasmak Alaskan-type intrusion, Turkey | Carboniferous | Suprasubduction zone                   | Yang, Shi, Wu, Su, and Chen (2008) |
| Hongliugou Ophiolite, China        | Cambrian-Ordovician | Suprasubduction zone                  | Leake and Styles (1984)     |
| Lizard Ophiolite, England          | Devonian        | Suprasubduction zone                    | Boudier and Nicolas (2011b) |
| Oman Ophiolite, Oman               | Cretaceous      | Suprasubduction zone                    | Jian, Wang, He, and Wang (1999) |
| Jinshajiang Ophiolite, China       | Carboniferous?  | Unkonwn                                 | Austheim and Prestrup (2008) |
| Leka Ophiolite, Norway             | Cambrian-Ordovician | Unkonwn                | Mitis and Economou-Eloipoulos (2001) |
| Otrhis Ophiolite, Greece           | Mesozoic        | Suprasubduction zone                    | Parkinson (1998); Kadarusman, Miyashita, Maruyama, Parkinson, and Ishikawa (2004) |
| East Sulawesi Ophiolite, Indenosis  | Cenozoic        | Suprasubduction zone, mid ocean ridge   |                             |
| Elder Creek Ophiolite, California (USA) | Mesozoic        | Suprasubduction zone, forearc          | Shervais (2003)             |

In this contribution, we use the geological and geochemical characteristics of well-studied Mesoarchean to Neoarchean anorthosite-bearing layered intrusions in western Greenland (Ivisaartoq, Fiskenæsset, Naajat Kuuat complexes) and the Superior Province of Canada (Bad Vermilion Lake and Doré Lake complexes) to address the above questions and propose a new geodynamic model for the petrogenetic origin of these rocks (Figures 1 and 2). Emphasis is placed on the megacrystic anorthosites and leucogabbros in the Fiskenæsset Complex where most stages of anorthosite and leucogabbro development are well exposed (Figures 3 and 4). In addition, we present new major and trace element and petrographic data from the Sinarssuk area in the Fiskenæsset region to provide new constraints on the origin of megacrystic anorthosite and leucogabbros. Analytical methods of major and trace elements for the new analyses are the same as described in Polat et al. (2011). The method used for Scanning Electron Microscope–Energy Dispersive Spectroscopy (SEM–EDS) analyses is given in Price (2012).

2. Regional geology

2.1. Archaean craton of southern West Greenland

The Archaean terrain of West Greenland is composed of Eoarchean to Neoarchean tectonically-accreted crustal blocks, consisting predominantly of TTG and granitic gneisses, basalt-dominated greenstone belts (volcanic and sedimentary rock associations), anorthosite-bearing layered intrusions, and granites (Friend & Nutman, 2005; Polat et al., 2015; Steenfelt, Garde, & Moyen, 2005; Windley & Garde, 2009). The poly-deformed and metamorphosed greenstone belts and anorthosite-bearing layered intrusions occur as meter- to kilometer-scale conformable layers within the TTG gneissies. Contacts between the TTG gneisses and greenstone belts and anorthosite-bearing layered intrusions are characterized predominantly by 5 to 20 meter wide mylonitic shear zones with rare intrusive relationships (Polat et al., 2015; Windley & Garde, 2009). The anorthosite-bearing layered intrusions are commonly spatially associated with volcanic rocks and are interpreted to have been emplaced into oceanic basaltic to gabbroic rocks, and they are in turn intruded by syn-tectonic TTGs (Hoffmann et al., 2012; Myers, 1976; Windley & Garde, 2009). The greenstone belts consist of tholeiitic to calc-alkaline basalts, boninites and picrites with minor siliciclastic sedimentary rocks. The anorthosite-bearing layered intrusions are composed mainly of cumulate layers of anorthosite, leucogabbro, gabbro and ultramafic rocks (Figure 4). Field, geochronological and geochemical investigations suggest that the Archaean craton of southern West Greenland grew at convergent plate margins through accretion of island arcs and continental blocks (Dziggel, Diener, Kolb, & Kokfett, 2014; Friend & Nutman, 2005; Garde, 2007; Kisters, van Hinsberg, & Szilas, 2012;
Nutman, Friend, & Bennett, 2002; Polat et al., 2015; Szilas et al., 2012, 2013; Windley & Garde, 2009). Despite polyphase deformation and amphibolite to granulite facies metamorphism, primary structures (e.g. pillows) in the volcanic rocks and cumulate textures and igneous layering in the anorthosite-bearing layered intrusions are well preserved (Chadwick, 1985, 1990; Myers, 2001; Polat et al., 2008, 2011, 2015; Windley & Garde, 2009).

Windley and Garde (2009) divide the Archaean craton of southern West Greenland into six Mesoarchaean to Neoarchaean (ca. 3000–2720 Ma) tectonic blocks that display similar geological cross-sections (Figure 1). These blocks consist of a southerly upper and a northerly lower zone (Windley & Garde, 2009). These blocks are composed predominantly of TTG orthogneisses containing numerous layers of anorthosite-bearing layered intrusions and amphibolite facies metavolcanic rocks.

### 2.2. Superior Province of Canada

The Archaean Superior Province of Canada consists of approximately east-west-striking TTG dominated-plutonic, greenstone-granitoid, high-grade TTG-gneissic and meta-sedimentary subprovinces (Figure 2) (Card & Ciesielski, 1986; Percival et al., 2012; Stott, 1997). These subprovinces represent a tectonic collage of numerous fragments of continental blocks, oceanic island arcs, oceanic plateaus and orogenic flysch deposits. These
Figure 2. (a) Simplified tectonic map of the Superior Province (modified from Percival et al., 2012). A: Ashuanipi; B: Bienville; DH: Douglas Harbour; E: Eastmain; ERB: English River Belt; G: Goudalie; HBT: Hudson Bay Terrane; ILD: Island Lake Domain; KU: Kapuskasing Uplift; LG: La Grande; LM: Lac Minto; MRVT: Minnesota River Valley Terrane; MT: Marimon Terrane; NCT: North Caribou Terrane; O: Opatica; Op: Opinaca; OSD: Oxford-Stull Domain; PB: Pontiac Belt; Q: Qalluviartuqu; QB: Quetico Belt; T: Tikkerutuk; U: Utsalik; UD: Uchi Domain; WAT: Wawa-Abitibi Terrane; WRT: Winnipeg River Terrane; WWT: Western Wabigoon Terrane.

Figure 3. Field photographs of megacrystic leucogabbro and anorthosites in the Fiskenæsset Anorthosite Complex, western Greenland, showing the various stages of anorthosite development. Permanent marker has a length of 15 cm. (b) modified from Polat et al. (2009); (c) modified from Huang et al. (2012).
In addition, gabbros in the Ivisaartoq belt contain up to 15 cm long, deformed anorthositic inclusions (xenoliths), consisting primarily (>90%) of Ca-rich plagioclase and 5–10% amphibole (Figure 5(a, b); Polat et al., 2008). These xenoliths are interpreted to be anorthositic cumulates that were transported to the oceanic crust by upwelling gabbroic magmas. The xenoliths display Th- and LREE-enriched trace element patterns with large positive Eu but negative Nb and Ti anomalies (Figure 6(a)), and have large positive initial εNd (+4.8 to +6.0) values, consistent with a long-term depleted mantle source (Polat et al., 2008). Volcanic and intrusive rocks of the Ivisaartoq greenstone belt share the geochemical characteristics of Phanerozoic subduction zone magmas (Ordóñez-Calderón et al., 2009; Polat et al., 2008; Szilas et al., 2016). The Mesoarchaean Naajat Kuuat Anorthosite Complex in southern West Greenland is composed of megacrystic anorthosite and leucogabbro, gabbro and ultramafic rocks, and is spatially associated with metavolcanic amphibolites (Figure 1) (Hoffmann et al., 2012). The meter- to kilometer-long slices of the Naajat Kuuat Anorthosite Complex are interleaved with the intrusive TTGs. The anorthosites have arc-like trace element patterns, whereas the ultramafic rocks and metavolcanic amphibolites display MORB- to arc-like geochemical characteristics (Hoffmann et al., 2012). Like those in the Archaean craton of southern West Greenland, the anorthosite-bearing layered intrusions in the Superior Province display well-preserved megacrystic textures (Ashwal, 1993) and are spatially and temporally associated with tholeiitic to calc-alkaline basalt-dominated volcanic rocks, and are intruded by syn- to post-tectonic TTGs (Bédard, Leclerc, Harris, & Goulet, 2009; Davis et al., 1988; Leclerc et al., 2011; Polat, Frei, Longstaffe, & Woods, 2017; Wu et al., 2016).

3. Summary of the geological and geochemical characteristics

3.1. West Greenland

An assemblage of leucogabbros, anorthosites, and differentiated sills in the Sinarssuk area of the Fiskenæsset Anorthosite Complex, West Greenland. (a) modified from Polat et al. (2011); (b) and (d) modified from Polat et al. (2009); (c) and (f) modified from Huang et al. (2012). The Mesoproterozoic lithotectonic terrains were assembled through subduction-driven, >1500 km long, accretionary and collisional processes, which took place diachronously from north to south between 2720 and 2680 Ma (Percival et al., 2012; Stott, 1997). Layered intrusions consisting mainly of gabbro, anorthosite and ultramafic rocks are exposed in many areas particularly in the northern Abitibi and the western Wabigoon greenstone-granitoid subprovinces and the North Caribou terrane (see Figure 3 in Percival et al., 2012). Like those in the Archaean craton of southern West Greenland, the anorthosite-bearing layered intrusions in the Superior Province display well-preserved megacrystic textures (Ashwal, 1993) and are spatially and temporally associated with tholeiitic to calc-alkaline basalt-dominated volcanic rocks, and are intruded by syn- to post-tectonic TTGs (Bédard, Leclerc, Harris, & Goulet, 2009; Davis et al., 1988; Leclerc et al., 2011; Polat, Frei, Longstaffe, & Woods, 2017; Wu et al., 2016). The Mesoproterozoic lithotectonic terrains were assembled through subduction-driven, >1500 km long, accretionary and collisional processes, which took place diachronously from north to south between 2720 and 2680 Ma (Percival et al., 2012; Stott, 1997). Layered intrusions consisting mainly of gabbro, anorthosite and ultramafic rocks are exposed in many areas particularly in the northern Abitibi and the western Wabigoon greenstone-granitoid subprovinces and the North Caribou terrane (see Figure 3 in Percival et al., 2012). Like those in the Archaean craton of southern West Greenland, the anorthosite-bearing layered intrusions in the Superior Province display well-preserved megacrystic textures (Ashwal, 1993) and are spatially and temporally associated with tholeiitic to calc-alkaline basalt-dominated volcanic rocks, and are intruded by syn- to post-tectonic TTGs (Bédard, Leclerc, Harris, & Goulet, 2009; Davis et al., 1988; Leclerc et al., 2011; Polat, Frei, Longstaffe, & Woods, 2017; Wu et al., 2016).

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mantle Nd and Hf isotopic compositions. On the basis of field observations and geochemical data, the Naajat Kuuat Anorthosite Complex and spatially associated amphibolites and TTGs are interpreted to have formed in a suprasubduction zone setting (Hoffmann et al., 2012).

The ca. 2970 Fiskenæsset Anorthosite Complex, Western Greenland, is probably the largest and best-preserved Archaean anorthosite-bearing layered intrusion in the world, consisting of ca. 550 m-thick cumulate layers of anorthosite, leucogabbro, gabbro and hornblende-bearing ultramafic rocks (Figures 3 and 4) (Huang, Fryer, Polat, & Pan, 2014; Huang, Polat, Fryer, Appel, & Windley, 2012; Myers, 1985; Polat, Frei, Scherstén, & Appel, 2010; Polat et al., 2009, 2011, 2012; Rollinson, Reid, & Windley, 2010; Windley & Smith, 1974; Windley et al., 1973). Field characteristics indicate that the Fiskenæsset Anorthosite Complex was emplaced into Mesoarchaean oceanic crust as multiple sills and dykes of magma and crystal mush, and is intruded by 2950–2750 Ma TTGs (Huang, Polat, & Fryer, 2013; Polat et al., 2010, 2011). The Fiskenæsset anorthosites and leucogabbros display diverse REE patterns with variable La/Sm and Gd/Yb ratios, and have variably negative Nb and Ti but positive Eu anomalies (Huang et al., 2012; Polat et al., 2009, 2011). All rock types in the complex share the trace element characteristics of modern subduction zone igneous rocks, and have depleted mantle Nd (average initial εNd=+3.3), Pb and O primitive isotope (δ18O = 5.8 ± 0.5‰) compositions (Huang et al., 2012; Polat & Longstaffe, 2014; Polat et al., 2009, 2010, 2011).

For this study, we analysed seven samples of megacrystic plagioclase from the leucogabbros of the Sinarssuk area (Figure 1; Table 2). These plagioclase megacrysts have small variations in SiO2 (46.7–48.2 wt.%), CaO (14.9–16.2 wt.%), and Al2O3 (31.9–3.5 wt.% (Table 2). They have low MgO (0.18–0.33 wt.%), Fe2O3 (0.48–1.1 wt.%), TiO2 (0.01–0.02 wt.%) and Na2O (1.65–2.55 wt.%). Mg-numbers span from 37 to 52 (Table 2). Anorthite (An%) contents vary from 78 to 86. Al2O3/TiO2 (1440–4440) ratios are extremely super-chondritic. The Sinarssuk megacrystic plagioclase samples are characterized by strongly fractionated REE patterns (La/Sm=3.0–13.9; La/Yb=6.2–83.7; Gd/Yb=1.79–4.92), and positive Eu (Eu/Eu*=2.91–8.3) anomalies (Figure 6(b); Table 2). Cerium anomalies are minor to absent (Ce/Ce*=0.96–1.05). On the N-MORB-normalized diagram, they have large positive Pb (Pb/Pb*=186–3740), and negative Nb (Nb/Nb*=0.06–0.55) and Ti (Ti/Ti*=0.13–0.29) anomalies. Zirconium anomalies (Zr/Zr*=0.36–6.65) vary from negative to positive (Figure 6(c)).

3.2. Superior Province

The Superior Province contains several well-preserved Neoarchaean anorthosite complexes including the Bad Vermillion Lake, Doré Lake, Shawmere, Pipestone, Pikwitonei and Bird River complexes (Ashwal, 1993;
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has depleted mantle-like Nd ($\varepsilon_{Nd} = \text{ca.} +2$) and near-prim-itive O ($\delta^{18}O = +5.5 \text{ to } +6.7 \text{‰}$) isotope signatures and subduction zone-like trace element patterns (Figure 6(c); Ashwal et al., 1983; Zhou et al., 2016). The spatially asso-ciated mafic to felsic volcanic rocks with the complex also have subduction zone trace element characteristics (Wu et al., 2016).

The ca. 2728 Ma Doré Lake Complex is located in the northeastern segment of the Northern Volcanic Zone of the Abitibi subprovince and is composed predominantly of megacrystic anorthosite and leucogabbro, and gabbro (Bédard et al., 2009; Daigneault et al., 1990; Leclerc et al., 2011; Polat et al., 2017). The complex has depleted mantle-like Nd ($\varepsilon_{Nd} = \text{ca.} +2$) and near-prim-itive O ($\delta^{18}O = +5.5 \text{ to } +6.7 \text{‰}$) isotope signatures and subduction zone-like trace element patterns (Figure 6(c); Ashwal et al., 1983; Zhou et al., 2016). The spatially asso-ciated mafic to felsic volcanic rocks with the complex also have subduction zone trace element characteristics (Wu et al., 2016).

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**Figure 6.** N-MORB-normalized trace element patterns for the Ivisaartoq (a) Fiskenæsset (b), Bad Vermilion Lake (c), and Doré Lake (d) anorthosites. Data for (a), (b), (c) and (d) from Polat et al. (2008, 2009), Zhou et al. (2016), and Polat et al. (2017), respectively. Normalization values are from Sun and McDonough (1989).
Like the Bad Vermilion Lake anorthosites and spatially associated volcanic rocks, the Doré Lake anorthosites like the Bad Vermilion Lake anorthosites and spatially associated volcanic rocks, the Doré Lake anorthosites and spatially associated volcanic rocks, the Doré Lake anorthosites and spatially associated volcanic rocks (see Figures 8–10 in Leclerc et al., 2011; Polat et al., 2017) also have subduction zone trace element signatures (Figure 6(d)), although Leclerc et al. (2011) do not attribute the origin of Th- and LREE-enriched patterns with negative Nb, Ta and Ti anomalies in the tholeiitic to calc-alkaline volcanic rocks to Archaean subduction zone processes.

4. Geodynamic origin of Archaean anorthosites

Numerous studies on the thermal history of the Earth suggest that the ambient temperature of the Archaean mantle (1500–1600 °C) was 200–300 °C higher than today's mantle (1300–1400 °C) (e.g. Davies, 2009; Herzberg, Condie, & Korenaga, 2010; Korenaga, 2008; Lee, Lufi, Plank, Dalton, & Leeman, 2009; Nisbet, Cheadle, Arndt, & Bickle, 1993). The higher mantle temperatures beneath the Archaean oceanic spreading centres would have produced 25–35% partial melting, resulting in three to four times more voluminous basaltic magma.

Table 2. Major (wt.%) and trace element (ppm) compositions and significant element ratios of megacrystic plagioclase in leucogabbros at Sinarsuk, Fikseneset Complex.

| SiO2  | TiO2  | Al2O3 | Fe2O3 | MgO  | CaO  | Na2O | P2O5 | LOI  | Rb  | Sr  | Cs  | Ba  | V  | Ta  | Nb  | Zr  | Hf  | Th  | U  | Y  | La  | Ce  | Eu  | Gd  | Tb  | Dy  | Ho  | Er  | Tm  | Yb  | Lu  | Ce  | Ga  | Pb  | Li  | La/Sm  | Gd/Yb | Eu/Eu* | Ce/Ce* | Pb/Pb* | Al2O3/TiO2 | Nb/Nb* | Zr/Zr* | Ti/Ti* | North | West |
|-------|-------|-------|-------|------|------|------|------|------|-----|-----|-----|-----|----|----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
| 47.72 | 0.01  | 32.76 | 0.48  | 0.18 | 15.34| 1.65 | 0.01 | 2.64 | 43  | 82  | 0.44 | 232 | 8.1 | 0.018| 0.201| 8.15 | 0.881| 0.446| 0.23 | 1.712| 2.722| 0.238 | 0.706 | 0.097 | 0.140 | 0.087 | 0.009 | 0.004 | 0.008 | 0.022 | 0.003 | 0.015 | 0.003 | 6.0 | 87 |
| 46.90 | 0.01  | 32.99 | 0.90  | 0.30 | 16.14| 1.75 | 0.03 | 1.77 | 40  | 86  | 0.52 | 115 | 18  | 0.007| 0.092 | 7.13 | 0.038| 0.018| 0.016| 0.016 | 0.027 | 0.006 | 0.017 | 0.017 | 5.0 | 64 | 50 |
| 46.71 | 0.01  | 33.51 | 0.52  | 0.22 | 16.77| 1.77 | 0.01 | 1.33 | 46  | 86  | 0.14 | 51  | 5.1 | 0.015| 0.100 | 1.64 | 0.023| 0.001| 0.022| 0.022 | 0.008 | 0.003 | 0.002 | 0.002 | 8.5 | 47 | 49 |
| 47.68 | 0.01  | 33.22 | 0.53  | 0.31 | 14.94| 1.80 | 0.01 | 1.66 | 54  | 18  | 0.79 | 51  | 7.1 | 0.018| 0.162 | 9.35 | 0.048| 0.097| 0.31  | 0.22 | 0.012 | 0.005 | 0.007 | 0.007 | 1.9 | 47 | 49 |
| 48.72 | 0.02  | 31.32 | 0.83  | 0.30 | 15.31| 2.55 | 0.01 | 0.99 | 10 | 54  | 0.12 | 143 | 5.0 | 0.034| 0.342 | 2.07 | 0.021| 0.013| 0.67  | 0.77 | 0.012 | 0.005 | 0.015 | 0.015 | 32.7 | 47 | 51 |
| 48.22 | 0.02  | 32.35 | 0.80  | 0.22 | 15.79| 2.28 | 0.01 | 0.97 | 10 | 18 | 0.13 | 177 | 7.1 | 0.035| 0.310 | 9.23 | 0.134| 0.053| 0.77  | 0.43 | 0.009 | 0.015 | 0.010 | 0.010 | 1.5 | 49 | 51 |
| 47.09 | 0.02  | 32.86 | 1.10  | 0.22 | 16.22| 1.97 | 0.02 | 0.74 | 12 | 10 | 0.09 | 122 | 7.1 | 0.012| 0.102 | 3.02 | 0.093| 0.018| 0.43  | 0.15 | 0.015 | 0.015 | 0.015 | 0.015 | 0.5 | 51 | 54 |
| 47.58 | 0.02  | 32.80 | 0.74  | 0.27 | 15.79| 1.97 | 0.02 | 0.74 | 12 | 10 | 0.09 | 122 | 7.1 | 0.012| 0.102 | 3.02 | 0.093| 0.018| 0.43  | 0.15 | 0.015 | 0.015 | 0.015 | 0.015 | 0.5 | 51 | 54 |

1990; Leclerc et al., 2011; Ludden, Francis, & Allard, 1984; Mortensen, 1993). The Nd (εNd=+2.6 to +5.0) and O (δ18O=-6.05 to +7.85%) isotope systematics are consistent with a depleted mantle source (Polat et al., 2017).
than the 7–10% partial melting beneath present-day spreading centres (see Herzberg, 2004; Herzberg et al., 2010; McKenzie & Bickle, 1988). Such high degrees of partial melting would have generated 25–35 km thick oceanic crust, in contrast to the 7–10 km thick crust produced at modern ocean ridges (Bickle, 1986; Herzberg et al., 2010; Sleep & Windley, 1982). Although petrological models predicting higher mantle temperatures for the Archaean are based on the composition of non-arc basalts (e.g. Herzberg et al., 2010; Lee et al., 2009), the presence of arc picrites, boninites and high-Mg andesites in Archaean greenstone belts might also imply higher ambient mantle temperatures beneath Archaean arcs and backarcs (Polat & Kerrich, 2006). We propose that 25–35% degrees of partial melting in the Archaean upper mantle would have resulted in a large volume of basaltic magmas underplating rifting arcs and/or backarc basins, generating large, shallow magma chambers (Figure 7). Higher geothermal gradients in the Archaean oceanic crust would have allowed these shallow magma chambers to cool slowly, providing optimal conditions for differentiation and stratification to produce cumulates of dunite, peridotite, chromitite, pyroxenite, (± hornblende-dite), gabbro, leucogabbro and anorthosite (Figure 7) (Namur et al., 2015; Polat et al., 2009). The Fiskenæsset Complex contains numerous small-scale (50 centimetres to 5 meters thick) examples of differentiated sills representing smaller versions of magma chambers that solidified to form such cumulate layers, including hornblende, in the oceanic crust (Figure 4(a, b); see Myers, 1985; Polat et al., 2011).

Archaean plagioclase megacrysts, including those in the Fiskenæsset Complex, are typically characterized by high-Ca content (An$_{70-93}$) (see Ashwal, 2010; Ashwal & Bybee, 2017). Previous studies have shown that plagioclase grains in the anorthosites and leucogabbros of the Fiskenæsset Complex typically have high anorthite (An) contents, ranging between 75 and 98% (Huang

**Figure 7.** Simplified geodynamic model for the origin of Archaean anorthosite-bearing layered intrusions (modified from Polat et al., 2008). (c) shows an example of mineralogically stratified sill, representing a small version of Archaean magma chambers.
et al., 2014; Myers & Platt, 1977; Rollinson et al., 2010; Windley & Smith, 1974). The plagioclase megacrysts from the Sinarsuk area of the Fiskenaesset region (Figures 1, 3) also have high-Ca contents (An_{78-86}). Additional analyses of plagioclase in leucogabbros and hornblendites on the island of Qeqertarsuatsiaq also indicate high An contents (An_{66-91}) (see Data Repository Table 1). These high An contents are attributed to hydrous parental melts in a subduction zone environment (see Ashwal, 2010; Müntener, Kelemen, & Grove, 2001; Takagi, Sato, & Nakagawa, 2005).

Virtually all rock types and structures that are associated with Archaean anorthosite-bearing layered intrusions (Daigneault et al., 1990; Davis et al., 1988; Myers, 1976, 1985; Polat et al., 2015; Wu et al., 2016) have analogues in Phanerozoic convergent plate margins and orogenic belts. Although the details of tectonic processes that produced Archaean anorthosite-bearing layered intrusions and spatially associated volcanic rocks and TTG intrusions remain to be determined, trace element and field observations indicate that their origin can be explained best by subduction zone tectonic processes (Chown, Daigneault, Mueller, & Mortensen, 1992; Daigneault, Mueller, & Chown, 2002; Mueller, Daigneault, Mortensen, & Chown, 1996; Percival et al., 2012; Polat et al., 2009, 2011, 2017; Windley & Garde, 2009). Available Nd and O isotope data for Archaean anorthosite complexes suggest that they originated from long-term depleted mantle sources that were overprinted by subduction-derived melts and/or fluids (Polat & Longstaffe, 2014; Polat et al., 2008, 2010, 2017; Zhou et al., 2016). None of the Archaean anorthosite-bearing layered intrusions discussed in this contribution displays evidence for extensive crustal contamination. Hence, the Th- and LREE-enriched and Nb- and Ti-depleted trace element patterns in these rocks (Figure 6) reflect subduction zone geodynamic processes rather than emplacement into older continental crust. Collectively, we interpret the anorthosite-bearing layered intrusions in the Archaean craton of western Greenland and the Superior Province as remnants of rifted oceanic arcs or backarc basins (Figure 7).

5. Anorthosite-bearing layered intrusions and the nature Archaean oceanic crust

Given their absence in the Phanerozoic rock record, Archaean anorthosite-bearing layered intrusions and associated greenstone belts are not generally considered part of oceanic crust because they do not have a Penrose-type ophiolite lithological association. Thus, these rock associations are not interpreted to be ophiolites (Hamilton, 1998; Kamber, 2015; Stern, 2005). Recent studies on Phanerozoic ophiolites and Archaean greenstone belts have shown that the original Penrose ophiolite definition (Anonymous, 1972) is too simplistic and has major limitations for understanding the evolution of the oceanic crust and its diversity in the rock record (Dilek & Furnes, 2011; Furnes, Dilek, & de Wit, 2015; Kusky, 2004; Kusky et al., 2013; Şengör & Natal'in, 2004). Therefore, this definition is not recommended to be used as a guide to define ophiolites in Archaean terrains that are typically characterized by polyphase deformation, and multiple generations of metamorphism and granitoid intrusion. Given that the Penrose ophiolite definition cannot be readily applied to the Precambrian rock record, Dilek and Furnes (2011) proposed a broad definition ophiolite as “suites of temporally and spatially associated ultramafic to felsic rocks related to separate melting episodes and processes of magmatic differentiation in particular tectonic environments”. On the basis of new definition, ophiolites are divided into two major types: (1) subduction-related ophiolites that form in backarc, forearc and arc tectonic settings; and (2) subduction-unrelated ophiolites that originate in rifted continental margins, mid-ocean ridges and plume-derived oceanic plateaus. The subduction-related ophiolites constitute about 85% of the ophiolites in the rock record (Furnes et al., 2015), reflecting mainly their biased preservation. Because of its buoyancy, suprasubduction crust is less prone to the subduction and recycling than “normal” oceanic crust produced along mid-ocean ridges. Most Archaean greenstone belts have Phanerozoic subduction zone-like lithological and geochemical characteristics, representing dismembered fragments of oceanic island arcs and backarc basins (Furnes et al., 2015).

On the basis of their lithological and geochemical characteristics, we interpret Archaean anorthosite-bearing layered intrusions and spatially associated greenstone belts to be subduction-related Archaean ophiolites using the new ophiolite concept developed by Dilek and Furnes (2011) and Furnes et al. (2015). These rock associations represent relict fragments of Archaean suprasubduction zone (arc-backarc-forearc) crust, marking the closure of ocean basins and formation of suture zones in Archaean orogenic belts (Polat et al., 2017).

Modern oceanic lithosphere forming at spreading centres consists generally of a mantle section (dunite, lherzolite, harzburgite) at the bottom, mafic to ultramafic cumulates (gabbros, pyroxenites) in the middle, and a mafic crustal section (isotopic gabbros, sheeted dykes and basalts) at the top. The thickness of the mafic rocks in modern oceanic crust and Phanerozoic ophiolites varies between 5 and 10 km, which is much smaller than the predicted 25–35 km thickness of Archaean oceanic crust (Herzberg et al., 2010; Sleep & Windley, 1982). In contrast to Phanerozoic oceanic crust, Archaean oceanic crust was likely composed of an association of 25–35 km thick basaltic flows and sills of gabbros, anorthosites, leucogabbros and differentiated ultramafic rocks (Figure 7(b)) (Dilek & Polat, 2008).
6. Magma chamber dynamics and the origin of megacrystic plagioclase

Textural analyses of layered intrusions and theoretical studies of magma chamber dynamics suggest that if a magmatic system with a single mineral, like plagioclase growing in a liquid at the top of a magma chamber, remains at high temperatures close to the liquidus of that mineral, large crystals will develop at the expense of small ones (Ostwald ripening), resulting in a megacrystic grain size (Higgins, 2005, 2011). Field observations of the Fiskenæsset Complex indicate that plagioclase crystals separated from the residual liquid and floated to the top.

Figure 8. Field photographs of plagioclase megacrysts in the Fiskenæsset Complex. Dark to grey cores are relict igneous plagioclase, whereas clear to white rims and patches represent the recrystallized metamorphic plagioclase. (b) modified from Huang et al. (2014); (d) modified from Polat et al. (2009).

Figure 9. Photomicrographs across two megacrystic plagioclase grains, illustrating the preservation of igneous plagioclase in the Fiskenæsset Complex despite deformation and metamorphism. Notes: Optical continuity (e.g. albite twinning) in igneous plagioclase grains suggests that they formed as 6–8 cm long single crystal in a magma chamber. Albite twins in large igneous crystals are overprinted by smaller metamorphic crystals.
of the magma chamber, forming a crystal mush of leucogabbro and anorthosite (Figure 4(a–c); see Higgins, 2005; Namur et al., 2011). Floatation of plagioclase was likely facilitated by the presence of denser residual liquid, which mostly crystallized to hornblende (Figures 3 and 4). The presence of plagioclase laminations, erosional surfaces, slump structures and trough layers (Myers, 1985), and the injection of plagioclase crystal mush into a gabbroic magma (Figure 4(d–f)) are consistent with a dynamic, vigorously convecting magma chamber (see Higgins, 2005; Namur et al., 2015). The presence of amphibole inclusions in chromite grains in the Fiskenæsset Complex is attributed to a hydrous magma composition (Polat, 2012; Rollinson et al., 2010). The Fiskenæsset Complex provides field evidence for most stages of crystallization of plagioclase in hydrous melts to form megacrystic anorthosites and leucogabbros (Figure 3). This hydrous magmatic system likely remained at high temperatures (1000–1200 °C) for a long period of time, resulting in the formation of megacrystic (2–15 cm in diameter) plagioclase grains. Plagioclase grain size varies from 2 mm to 30 cm and some megacrysts appear to have formed by coalescence of smaller megacrysts (Figures 3 and 8). Most plagioclase grains display field and petrographic evidence for metamorphic recrystallization (Figure 8). The continuity of albite twinning for 6–8 cm in the least deformed and recrystallized megacrysts in samples 508144 and 508146 collected from the least-deformed outcrops in the Sinarsuk area (Figures 1, 9; Table 2), however, clearly demonstrates their formation as single crystals, rather than coalesced, in the magma chamber. The megacrystic magma either solidified in place or was transported as crystal mush into the overlying oceanic crust to form sills, dykes or pods of anorthosites and leucogabbros (Figure 7). These plagioclase megacrysts grew further by interaction with either new melts and/or melts expelled from the lower parts of the magma chamber, producing grains up to 30 cm in length (Figure 3). Collectively, large grain sizes in Archaean megacrystic anorthosites and leucogabbros can be interpreted as the product of slow cooling conditions close to the plagioclase liquidus temperatures that stemmed from the higher geothermal gradients in the Archaean oceanic crust.

7. Conclusions

Anorthosite-bearing layered intrusions in the Archaean craton of southern West Greenland and the Superior Province of Canada were emplaced into basalt-dominated greenstone belts, and were intruded by syn- to post-tectonic granitoids. All these rock associations share the trace element signatures of modern subduction-related magmas, suggesting that they originated in a suprasubduction zone setting, representing Archaean subduction-related ophiolites. Anorthosite-bearing layered intrusions and spatially associated greenstone belts are interpreted as fragments of suture zones, marking the closure of Archaean ocean basins. These intrusions in Greenland and Canada are mostly located close to the major tectonic boundaries separating different tectonic blocks/terrane (Figures 1 and 2), suggesting a close relationship between the formation of anorthosite-bearing layered intrusions and major tectonic processes that led the amalgamation of different blocks/terrane.

Ashwal (2010) and Ashwal and Bybee (2017) addressed the temporality of anorthosites, showing that high-Ca (An = 61–94%, average An = 80%) megacrystic Archaean and massif-type Proterozoic anorthosites (An = 30–70%, average An = 53%) are restricted in space and time, whereas continental layered mafic intrusions and ophiolites do not show clear time-restriction. Field studies indicate that Archaean anorthosites originated mostly in oceanic settings and are spatially and temporally associated with basalt-dominated greenstone belts, whereas Proterozoic counterparts are spatially associated with granitoid rocks and thought to have formed in continental arcs. Anorthosites in Phanerozoic ophiolites (Table 1) originated mainly in suprasubduction zone settings (e.g. oceanic arcs, backarc basins, and forearcs) and are characterized by millimetre grain size and centimetre to decimetre thick layers, in contrast to hundreds of meters to several kilometres thick Archaean megacrystic anorthosites.

We suggest that changes in the nature of anorthosites occurring in oceanic settings in Archaean and post-Archaean times reflect the thermal evolution of the Earth. Irreversible heat loss of the Earth led to upper mantle cooling by the end of the Archaean to a level at which it could no longer produce 25–35% partial melting, except within mantle plumes, beneath oceanic arcs and spreading centres. Large degrees of partial melting in the Archaean mantle led to the formation of mineralogically-stratified, km-scale magma chambers beneath spreading centres and magmatic arcs. These magma chambers consisted mainly of olivine-, pyroxene-, hornblende-, and plagioclase-rich crystal mushes (Figure 7), and cooled slowly in response to higher geothermal gradients in the oceanic crust and magmatic arcs, providing optimum petrological conditions for the formation of megacrystic plagioclase in anorthosites and leucogabbros (Figure 7). In addition to higher mantle temperatures, the presence of water in Archaean suprasubduction zone magmas might have played an important role in the formation of plagioclase megacrysts by enhancing the magma cooling time and element diffusion rates. Because of declining ambient mantle temperatures and geothermal gradients in the oceanic crust, the generation of large, mineralogically stratified magma chambers decreased and eventually ceased by the end of the Archaean Eon, leading to the termination of the formation of high-Ca megacrystic anorthosites in post-Archaean oceanic environments.
Acknowledgments

Tim Kusky, Claude Herzberg and Michael Higgins are acknowledged for their insightful, helpful and constructive comments on the previous version of the paper. This research is supported by NSERC grants to A. Polat and F.J. Longstaffe.

Disclosure statement

No potential conflict of interest was reported by the authors.

Funding

This work was supported by the Natural Sciences and Engineering Research Council of Canada.

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