Lithosphere thermal thickness and geothermal heat flux in Greenland from a new thermal isostasy method

Irina M. Artemieva
IGN, University of Copenhagen, Denmark.

A R T I C L E   I N F O
Keywords:
Lithosphere thickness
Iceland hotspot
Cratonic mantle
Continental crust
Kimberlite
Mantle temperature

A B S T R A C T
Lithosphere thermal structure in Greenland is poorly known and models based on seismic and magnetic data are inconsistent, while growing awareness in the fate of the ice sheet in Greenland requires reliable constraints on geothermal heat flux (GHF) from the Earth's interior in the region where conventional heat flux measurements are nearly absent. The lithosphere structure of Greenland remains controversial, while its geological evolution is constrained by direct observations in the narrow ice-free zone along the coasts. The effect of the Iceland hotspot on the lithosphere structure is also debated.

Here I describe a new thermal isostasy method which I use to calculate upper mantle temperature anomalies, lithosphere thickness, and GHF in Greenland from seismic data on the Moho depth, topography and ice thickness. To verify the model results, the predicted GHF values are compared to available measurements and show a good agreement. Thick (200–270 km) cratonic lithosphere of SW Greenland with GHF of ca. 40 mW/m² thins to 180–190 km towards central Greenland without a clear boundary between the Archean and Proterozoic blocks, and the deepest lithosphere keel is observed beneath the largest kimberlite province in West Greenland. The NW-SE belt with an anomalously thin (100–120 km) lithosphere and GHF of 60–70 mW/m² crosses north-central Greenland from coast to coast and it may mark the Iceland hotspot track. In East Greenland this anomalous belt merges with a strong GHF anomaly of > 100 mW/m² in the Fjordland region. The anomaly is associated with a strong lithosphere thinning, possibly to the Moho, that requires advective heat transfer such as above active magma chambers, which would accelerate ice basal melting. The anomaly may extend 500 km inland with possibly a significant contribution of ice melt to the ice-drainage system of Greenland.

1. Introduction
Greenland hosts many unresolved questions and even its topography remains enigmatic. East and South Greenland are flanked by mountains, reaching > 2 km in elevation (Fig. 1a). The presence of the Tertiary basalts at mountain tops suggests a young rapid uplift (Japsen and Chalmers, 2000) away from the plate boundaries, but the fastest recent vertical movements are south from the highest topography (Bevis et al., 2012). Ice melting (Velicogna et al., 2014) causes post-glacial isostatic adjustments, but there is a poor fit between these models (Khan et al., 2016) and the uplift rates in Greenland (Bevis et al., 2012).

The crustal structure in Greenland, best constrained by seismic P- and S-receiver function (PRF, SRF) studies (Dahl-Jensen et al., 2003b; Kumar et al., 2007), is controversial. PRF give systematically deeper Moho (by 6–10 km under the ice cap) than SRF (Fig. 2a) despite they are constrained by the same seismic data. The PRF results are in general agreement with the Rayleigh wave tomography model for the Moho depth (Mordret, 2018) and with a joint P- and SRF inversion (Kraft, 2016). It is likely that while PRF sample the Moho boundary, SRF sample some intracrustal high-contrast velocity interface, such as the top of the underplated material (Thybo and Artemieva, 2013). In such a case, the difference in depths reported by PRF and SRF would constrain the thickness of the underplated material, 6–10 km thick in Central Greenland (Fig. 2b). This is consistent with tomography inversion along the first controlled-source seismic refraction profile in Central Greenland, TopoGreenland (Shulgin and Thybo, 2015), which shows a ca. 50 km thick crust with a ca. 13 km thick high-velocity (7.0–7.3 km/s) lower crustal layer below the Summit station (Shulgin and Thybo, 2014). Therefore, the following analysis is constrained by PRF for the Moho depth in Greenland (Fig. 1b). For PRF, the Moho depth is well correlated with the Vp/Vs ratio (Fig. 2b), in support of an interpretation that magmatic underplating increases both the crustal thickness and the bulk crustal Vp/Vs ratio. Such correlation is absent for SRF.

The proximity to Iceland and the presence of the Tertiary basalts in East and West Greenland (Fig. 1a) formed basis for hypotheses on the
passage of the Iceland hotspot beneath Greenland in Paleogene (Forsyth et al., 1986; Lawver and Müller, 1994; O’Neill et al., 2005; Doubrovine et al., 2012; Torsvik et al., 2015). Reduced Vs speeds in the upper mantle of north-central Greenland may be a manifestation of the Iceland hotspot track, but the locations vary (Lebedev et al., 2018; Antonijevic and Lees, 2018; Pourpoint et al., 2018). Regional seismic tomography models on the deep structure of Greenland remain highly controversial (Fig. 3) and show a very heterogeneous lithosphere structure at all depths. Waveform tomography for the Arctic region imaged two nuclei of the cratonic lithosphere in Greenland centered at ca. 70°N and 80°N separated by a belt of slower mantle at ca. 73–75°N (Fig. 3b) which may be a manifestation of the Iceland hotspot track (Lebedev et al., 2018). The presence of a low velocity anomaly in central-eastern Greenland interpreted as the Icelandic plume track is also recognized in a recent Rayleigh wave group velocity tomography model (Pourpoint et al., 2018). In contrast, an ambient noise tomography model based on the same GLISN seismic data, also based on the GLISN seismic data, does not image any lithosphere anomaly which can be interpreted as an evidence for the hotspot track (Darbyshire et al., 2018) (Fig. 3d), while a strong low-velocity anomaly across Greenland is imaged at a shallow depth (Fig. 3c).

Recent trends in global warming increased the awareness in the fate of the ice sheets in Antarctica and Greenland (De Conto and Pollard, 2016; Oppenheimer and Alley, 2016) which are eroded both from the surface and from the base. Basal melting is significantly controlled by geothermal heat flux (GHF). Knowledge on GHF and lithosphere temperature anomalies is crucial for understanding the lithosphere dynamics and estimating basal ice temperature which is an important parameter for scenarios of ice sheet melting. However, conventional heat flux measurements are nearly absent in Greenland (Sass et al., 1972) and other measurements such as in ice drill holes and water basins (Dahl-Jensen et al., 1998; Rysgaard et al., 2018) also remain very scarce. GHF estimates from geophysical models, supplemented by models of ice sheet dynamics (Fahnestock et al., 2001; Rogozhina et al., 2016; van der Veen et al., 2007) are inconsistent and poorly correlate with models of the Greenland lithosphere thermal structure constrained...
by seismic (Lebedev et al., 2018; Mordret, 2018) and magnetic data (Fox Maule et al., 2005, 2009; Martos et al., 2018). The controversy on the lithosphere thermal structure in Greenland hampers not only modeling of the future ice sheet mass balance, but also geodynamic modeling of the cratonic lithosphere of Greenland and understanding mechanisms of its stability and preservation since the Neoarchean, partial reworking by the Iceland hotspot and by similar processes in the past.

2. Method in brief

The present study models temperature heterogeneity of the lithosphere of Greenland based on thermal isostasy analysis (described below with further technical details in Artemieva, 2018), given that the island is close to isostatic equilibrium with free air anomalies typically in the range from −20 mGal to +50 mGal (Fig. 4a). The Airy-type isostasy predicts a linear correlation between the depth to Moho and topography. The latter is replaced here by the equivalent topography to account for a ca. 3 km thick ice cap (Fig. 1a). The seismic Moho depth (Fig. 1b) is based on PRF (Dahl-Jensen et al., 2003b), complemented by P- and SRF results along the TopoGreenland profile (Kraft, 2016) (Fig. 1b). Since the present study builds on isostasy analysis, gravity constraints on the Moho depth (e.g. Steffen et al., 2017) cannot be used.

Under isostatic conditions, the Moho depth and (equivalent) topography are linearly correlated, and the slope of the correlation line constrains the density contrast across the Moho (Fowler, 2005). For Greenland the slope for the best fit line fits typical densities of continental crust (2.8 t/m³) and in situ upper mantle (3.35 t/m³) but predicts a smaller average Moho depth (37.5 km) than a global continental average of 41.1 km (Christensen and Mooney, 1995). Limited data on the seismic Moho in Greenland do not allow to calculate its area-weighted value and the regional value of 37.5 km can be biased by the uneven seismic station coverage (Fig. 1). Therefore the analysis is presented for two models with the same density contrast across the Moho, but with a different Moho depth at zero topography: Model 1 is based on the global continental average and Model 2 - on the regional average Moho depth in Greenland. These two models provide end-member constraints on the lithosphere thermal structure in Greenland.

First, local deviations from the equivalent topography predicted by isostasy (termed here anomalous equivalent topography, AET) are calculated at all seismic stations (see Appendix for details). These topography anomalies may be of thermal, compositional, or dynamic origin, but the existing data do not allow to separate these competing effects. Assuming that the AET anomalies are primarily of thermal origin (see Artemieva, 2018 for discussion of this assumption), I calculate temperature anomalies in the lithosphere, which next are converted to the anomalies in the lithosphere thermal thickness and geothermal heat flux (see Appendix). The latter are compared with observations, where available, and show a good agreement, thus providing support to the modeling results. The results show a highly heterogeneous lithosphere thermal structure in Greenland, with distinct lithosphere domains associated with the cratonic mantle of different ages and with a possible passage of the Iceland hotspot.
3. Results

3.1. Anomalous equivalent topography (AET)

Six lithosphere domains (labelled D1 to D6, Fig. 5) can be recognized in Greenland by the AET values. Domain 1 (cratonic) in south-western Greenland with AET < 0 includes the Archean block with the adjacent Proterozoic mobile belts. In Model 1, the transition from negative to positive AET values closely follows a traditionally adopted northern boundary of the Archean block (Fig. 5a), while in Model 2 it extends to the Summit station SUM and may include station TULE in north-west (Fig. 5b). Both models predict the maximum AET values around Kangerlussuaq (station SFJ), close to the Archean-Proterozoic boundary, suggesting that this block (which hosts the largest kimberlite province in Greenland) may be underlain by the Archean mantle. This conclusion is in general agreement with recent interpretations of a 3.7Ga subduction at the Isua Supracrustal belt (Kaczmarek et al., 2016).

Domain 2 (transitional) with slightly positive AET includes the east coast between stations ANG and SOE where the maximal uplift rates of ca. 30mm/y are observed (Bevis et al., 2012). In Model 1, this Domain includes the Summit station and station TULE.

Domain 3 (anomalous) in east-central Greenland has extreme positive values of AET (2–3km at near-coastal stations DBG and SCO). Its boundaries are sharp with a strong gradient of AET over distances of ca. 250 km, suggesting a shallow origin of the AET anomalies. Regions with the strongest positive AET correlate with some models for the Iceland hotspot track (O’Neil et al., 2005; Doubrovine et al., 2012; Torsvik et al., 2015; Martos et al., 2018) (Fig. 5b).

Domain 4 (modified cratonic) has small negative AET values and is constrained by data on the Moho depth for stations NGR and NEEM. Its northern boundary is unconstrained due to the absence of seismic data on the Moho depth.

Domain 5 (Paleozoic), constrained by only one station ASS in the North Greenland fold belt at the Arctic coast, has weakly negative AET, similar to the cratonic Domain 1. Importantly, recent geochemical data suggest that the region may contain the Archean terrane with ages older than 3.2Ga (Nutman et al., 2008).

Domain 6 (Caledonian) includes station DAG at the boundary with Domain 3 and station NOR further north, located in the region that was possibly affected by the High Arctic Large Igneous Province (130–70 Ma). The western boundary of the Caledonian fold belt is not seen in the AET and thermal anomalies. A heterogeneous AET pattern suggests the presence of Proterozoic lithosphere below parts of the Caledonian orogenic belt.

GPS measurements along the coasts of Greenland (Bevis et al., 2012) demonstrate heterogeneous vertical movement with high uplift rates (10–22 mm/y) along the west coast between stations SFJ and Tule (AET between −2.0 and −0.5 km in Model 1) with the highest uplift rate around station SFJ (Fig. 6a). At the east coast, the uplift rates are low (< 10 mm/y) along the Caledonian fold belt (AET between +2.0 and +3.0 km and −0.3 + 2.1 km for Models 1 and 2, respectively), and locally increase to ca. 30 mm/y between stations ANG and SOE, where the AET
values are around +0.4 + 0.8 km. Therefore there is no correlation between the present uplift rates and the anomalies in equivalent topography. The absence of correlation suggests that on-going differential vertical movements in the ice-free zone along the coasts reflect lithosphere flexure associated with glacial isostatic adjustments and ice melting, but not deep geodynamic processes.

3.2. Origin of anomalies in equivalent topography

AET anomalies can be caused by processes at the surface, in the underlying mantle, and within the lithospheric plate. Near-surface processes may reduce the equivalent topography by decreasing ice mass and erosion. However, high positive AET in Domain 3 are in the ice-free region (Fig. 5), suggesting that near-surface processes are not a major controlling factor to the AET heterogeneity. Dynamic effect from the mantle remains controversial (Cadek and Fleitout, 2003; Heine et al., 2008; Flament et al., 2014; Artemieva and Vinnik, 2016), although regional dynamic contribution to topography associated with the Iceland hotspot track is possible in limited parts of Greenland. However, it cannot be quantified, given the controversy in the proposed hotspot tracks (some of them are shown in Fig. 5b).

The lithosphere contribution to AET can be through three mechanisms: density heterogeneity of compositional origin, of thermal origin, and deformation related to glacial isostatic adjustment. The latter depends on mantle viscosity (Forte and Mitrovica, 1996; Mordret, 2018), which is poorly known for Greenland, where predictions for glacial isostatic adjustment do not match GPS data (Khan et al., 2016). Effects of compositional density heterogeneity, associated with crustal heterogeneity and depletion of the cratonic lithospheric mantle, are hard to quantify in the absence of high resolution seismic data on the inner crustal structure and mantle-derived xenoliths to constrain the composition of the lithospheric mantle. However, the effect from mantle compositional heterogeneity on lithosphere density and AET is much less significant than from temperature anomalies (Deschamps et al., 2001; Simmons et al., 2009; Artemieva, 2018). The maximum expected density difference between a highly depleted Archean and fertile lithosphere mantle is ca. 100 kg/m³ in shallow lithosphere mantle (Poudjom Djomani et al., 2001). This difference gradually decreases with depth and typically vanishes in the basal part of the lithosphere (Griffin et al., 1999, 2004). Therefore, the maximum expected compositional density difference averaged over the entire lithosphere mantle column is 50–70 kg/m³. Lithosphere delamination (Bird, 1979) and metamorphic reactions (Le Pichon et al., 1997) may also contribute to topography changes, but these mechanisms cannot be examined for Greenland in the absence of data.

Thermal heterogeneity is therefore the best candidate to explain the AET anomalies. A strong inverse correlation (Fig. 7) between the Bouguer (Fig. 4b) and the AET anomalies (Fig. 1a) in Greenland supports an interpretation that the AET anomalies are caused by lithosphere mantle density anomalies. I therefore test if thermal heterogeneities in the lithosphere and the upper mantle can explain most of the AET anomalies, assuming that all other factors play only a minor role and their effects on the AET variations can be neglected.

3.3. Lithosphere thermal thickness

Lithosphere thermal thickness zLAB is non-linearly related to AET (see Appendix), and it can be determined from it (Fig. 8a) if zLAB is known at reference locations where AET = 0 km (that is the equivalent topography is the same as predicted by the isostasy). The reference stations with AET = 0 km are NUK in Model 1 (Fig. 5a) and ANG, SOE in Model 2 (Fig. 5b), and I assume the reference zLAB value of 200 km and 150 km for...
Models 1 and 2, respectively. A choice of other values will lead to a systematic shift of the calculated lithosphere thickness by ca. 30 km on average, if the reference zLAB value is changed by 50 km (Fig. 8a).

Variations in lithosphere thermal thickness, calculated from AET, reveal the same six lithosphere domains in Greenland as the AET anomalies (Fig. 9). The results agree with recent regional tomography models (Antonijevic and Lees, 2018; Darbyshire et al., 2018; Lebedev et al., 2018; Levshin et al., 2017; Mordret, 2018; Pourpoint et al., 2018) only in general. Seismic models significantly differ in details, precluding a unique comparison between the thermal Moho depth and the equivalent topography for Greenland, which yields the same density contrast across the Moho as in Model 1 and the average Moho depth of 37.5 km. Dashed lines in (a) show boundaries of major tectonic provinces inferred from the near-coastal geology. D1-D6 are six lithosphere domains discussed in the text.

The Archean part has the lithosphere thickness of ca. 200–270 km, with a very cold and thick (ca. 270 km) lithospheric keel around station SFJ (Kangerlussuaq), located within the Proterozoic block ca. 250 km north from the early Archean Isua greenstone belt (station NUK). This region hosts the largest known kimberlite province in Greenland. The results indicate that in West Greenland the Archean block may extend further north than presently accepted. In general, the Archean-Proterozoic boundary, as known from geology in the ice-free coastal zones, is not reflected in the lithosphere thickness variations (Fig. 9).

Lithosphere thickness decreases northwards to 125–190 km in most of the Proterozoic terranes, with ca. 180–190 km beneath the Summit stand, dropping to 95–110 km further north (Domain 4) and increasing to 180–200 km along the Arctic coast, where geochemical data indicate a possible presence of the Archean lithosphere (Nutman et al., 2008) (Domain 5). The anomalous Domain 3 has the shallowest lithosphere (<100 km and possibly locally thinned to the Moho). This result is supported by a short wavelength (<200 km) of negative Bouguer anomalies (~150–200 mGal) in the east-central Greenland (Fig. 4b) which indicate the presence of a low-dense material at shallow depths. This conclusion agrees with some of the regional tomography models (e.g. Levshin et al., 2017; Mordret, 2018) which image slow shallow Vs anomalies in East Greenland (Fig. 3).

### 3.4. Thermal structure of the upper mantle

Lithosphere thickness zLAB is next used to constrain regional geotherms (Fig. 10) and to calculate temperature anomalies in the lithosphere with respect to the reference stations (see Appendix). While information on the lithosphere thickness is important for understanding lithosphere evolution and mantle dynamics, knowledge of temperature anomalies is pivotal for understanding processes at the rock-ice interface. As expected, the pattern of the temperature anomalies (Fig. 11) closely mimics the pattern of the zLAB anomalies, although the
correlation between the two parameters in non-linear (Fig. 8b). Temperature anomalies in the upper mantle are calculated with respect to the reference stations: NUK in Model 1 and ANG in Model 2.

Model calculation (see Appendix) determines geothermal heat flux as 44–48 mW/m² at station NUK and 54 mW/m² at station ANG (Fig. 12), which corresponds to a modified Archean – Proterozoic lithosphere (Artemieva, 2006). Negative temperature anomalies (50–270 °C colder than below the reference stations) correspond to the largely Archean Domain 1, while extreme positive temperature anomalies (> 800 °C) are in the anomalous Domain 3 (Fig. 10). Note that stations DBG and SCO with the most anomalous lithosphere structure also have the anomalous crustal structure and do not follow the regional correlation between the Moho depth and the bulk crustal Vp/Vs ratio (Fig. 2b). The Summit station is in the transitional zone with a temperature anomaly of ca. 360 °C in Model 1 and ca. 135 °C in Model 2, and it appears as a cold “island” surrounded by a hot mantle (Fig. 11).

The results show the belt with an anomalously high upper mantle temperature in north-central Greenland (Domain 4), which extends from the east coast to nearly the west coast. Its presence is confirmed by recent tomography models (Antonijevic and Lees, 2018; Lebedev et al., 2018; Pourpoint et al., 2018) and the Curie depth model derived from magnetic data (Martos et al., 2018), which all resolve a similar belt with an anomalously low-velocity upper mantle and a shallow Curie depth. However, the amplitude of the anomaly, its depth, and the location differ between the models. In particular, some ambient noise tomography models (Levshin et al., 2017; Mordret, 2018) image a localized anomaly rather than an anomalous belt (Fig. 3f).

Thermal structure of the Greenland upper mantle has been recently modeled based on regional seismic tomography (Lebedev et al., 2018; Mordret, 2018). Given a significant difference between these two shear-wave tomographic models (Fig. 3g,h), their temperature predictions are also significantly different. Lebedev et al. (2018) estimate a temperature difference of ca. 400 °C at depths 80–150 km between the anomalous belt (Domains 3 + 4) and a cold and thick lithosphere to the north and to the south. This estimate is consistent with the present results (Fig. 11), which however show a much broader transition zone (ca. 500 km) between the anomalous hot mantle belt (Domain 4) and the cold cratonic lithosphere to the south (Domain 2). According to the present results, a high temperature gradient zone exists only east of stations SUM and S17-S15.

The present results also show a huge temperature anomaly in the
Fig. 8. (a) Relationship between lithosphere thermal thickness and anomalous equivalent topography. Variations in density of lithosphere mantle, crust, and crustal thickness on the calculated lithosphere thickness have negligible effect compared to the effect of assumed $z_{\text{LAB}}$ at reference stations (see Eqs. (6)–(8) in Appendix). Three cases are shown for $z_{\text{LAB}}$ at reference stations of 150, 200, and 250 km. (b) Correlation between geothermal heat flux and lithosphere thermal thickness based on reference continental geotherms of Pollack et al. (1993).

Fig. 9. Lithosphere thermal thickness constrained by AET (Fig. 8a), assuming the lithosphere thickness at reference stations is 200 km (Model 1, NUK) and 150 km (Model 2, ANG and SOE). Models 1 and 2 are explained in caption to Fig. 5. Thin lithosphere (<70 km) cannot be determined from Fig. 8a, and stations with AET > 1.3 km are assigned the $z_{\text{LAB}}$ values of 50–80 km proportionally to the AET values.
upper mantle of central-east Greenland (Domain 3) with the amplitude of ca. 800–1000 °C with respect to the cratonic stations. Such a temperature anomaly cannot be explained by conductive nor radiogenic mechanisms. Several tomography models also resolve an anomalous thermal structure in Domain 3 (Lebedev et al., 2018; Mordret, 2018), but the location of the anomaly maximum varies. A low Vs anomaly centered at ca. 30°W, 73°N at a 36 km depth slice is also present in the model of Lebedev et al. (2018) and it disappears at a 80 km depth slice. In tomography model of Mordret (2018), the maximum low-Vs and high-temperature anomaly is around station SOE, with the temperature difference of 300–400 °C at a 180 km depth between the hot anomalous mantle at East Greenland and the cratonic mantle in central-southern Greenland. The amplitude of the temperature anomaly is similar to the present results, which place the center of the anomaly ca. 400–500 km further north in East Greenland. The model of Mordret (2018) resolves two other anomalies, one is centered close to station NGR and the other approximately at station NOR, where the present results also resolve a thin lithosphere and high temperatures in the upper mantle. A low Vs anomaly around station NGR, consistent with the present results, is also resolved by Antonijevic and Lees (2018) (Fig. 3a).

Thermal structure of Greenland's lithosphere has been recently modeled based on variations in the Earth's magnetic field (Martos et al., 2018). Curie depth variations constrain a strong thermal anomaly that extends along at the central-east Greenland coast from station ANG to station SCO, extends to stations SUM and NGR, and continues with a lower amplitude towards station NEEM and the north-western Greenland. The present results show in general a similar pattern, but with the near-coastal anomaly shifted northwards and with a much stronger anomaly amplitude at the East Greenland coast at the Fjordland region between stations SCO and DBG. Curie depth variations (Martos et al., 2018) also show a strong shallow anomaly in the Archean domain in the south, which is not observed neither in the present study nor in other geophysical models.

3.5. Geothermal heat flux (GHF)

Reference continental geotherms for different tectonic settings (Pollack et al., 1993; Artemieva, 2006) parameterized by surface heat flux constrain lithospheric thermal thickness as a function of GHF. Here the opposite approach is taken, and GHF is calculated from the modeled values of zLAB (Fig. 8b).

The predicted GHF (Fig. 12) shows a huge variability, ranging from ca. 37–44 mW/m² in the southern, largely Archean, Greenland (Domain 1) to 45–55 mW/m² in the Proterozoic and Paleozoic orogenic belts (Domains 1, 2, 3, 5) and in the central part of the Caledonian fold belt (Domain 5), exceeding > 70 mW/m² in the anomalous Domain 3 with the maximum values of > 100 mW/m² around stations DBG, HJO. Westwards extent of the high GHF region may either cover most of the north-central Greenland (Model 1), or have a southern-western extension along the TopoGreenland profile, with a more moderate anomaly in Domain 4 (Model 2). In both cases, GHF at the Summit station (46 mW/m²) is close to typical cratonic values and in agreement with inferred value of 51 mW/m² at the GRIP ice drill hole (Dahl-Jensen et al., 1998). Both models suggest that the Summit station is underlain by a rigid lithospheric block with a thick and cold lithosphere, which may be surrounded from nearly all sides by a thin and rheologically weak mantle. GHF increases north-westwards towards seismic station NGR (NorthGRIP) where borehole, ice-core and radio-echo sounder observations also suggest higher ice basal temperature (Dahl-Jensen et al., 2003a).

The results for stations PAA and NRS (41–46 mW/m²) in South Greenland agree with two conventional heat flux measurements of 38 and 42 mW/m² (Sass et al., 1972), while in the anomalous region at the central-east coast the calculated GHF at station DBG (110 mW/m²) agrees with values of 93 ± 21 mW/m² determined from temperature and salinity time series in the Sund-Tyrolerfjord (Rysgaard et al., 2018).

The extreme temperature anomalies in Domain 3 are inconsistent with conductive geotherms, and indicate the presence of a strong non-conductive component of GHF beneath the central-eastern coastal Greenland, such as short-wavelength advective heat flux associated with a hot fluid percolation above active magma chambers, typical for active rift systems. Furthermore, AET > 1.5 km at the central-east coast of Greenland cannot be explained by thermal anomalies alone and requires other mechanisms, such as mantle dynamic topography. A localized magmatic activity can also explain a sharp transition in thermal structure between stations DBG and DAG. Note that in case of advective heat transfer the term GHF as heat flux from the Earth's interior does not correspond to its definition in conventional geophysical measurements based on the assumption of a pure conductive heat transfer.

4. Discussion

A passage of the Iceland hotspot should have led to a significant thermo-chemical modification of the cratonic lithosphere with an expected lithosphere thinning along the hotspot track and possible densification and refertilization of cratonic lithospheric mantle by

Fig. 10. Lithosphere geotherms for four lithosphere domains in Greenland (see Fig. 5 for locations). Geotherms within the crust cannot be reliably constrained in the absence of data on crustal heat production. Symbols – xenolith P-T arrays from different continental locations. Thermal structure of Domain 1 is similar to the Archean Kaapvaal craton. Thermal structure of the anomalous Domain 3 is similar to Phanerozoic magmatic provinces of the Baikal region, south-eastern China, and south-eastern Australia. Aschepkov et al., 2003; De Conto and Pollard, 2016; Griffin et al., 1996; Ionov et al., 1999; Kaczmarek et al., 2016; Kopylova et al., 1999; Kukkonen and Peltonen, 1999; Malkoves et al., 2003; Oppenheimer and Alley, 2016; O'Reilly and Griffin, 1985; Rudnick and Nyblade, 1999; Velicogna et al., 2014; Xu et al., 1996.
magmatic intrusions. The present results are consistent with the hotspot trajectories that pass through the central-east coast (stations SCO and DBG) and possibly continue through Central Greenland towards station NGR (Fig. 12). This indicates that the hotspot could have caused >2 km of extra topography by an increase in mantle temperatures (Fig. 11) and lithosphere thinning to <100 km (Fig. 9).

Not all of topography anomaly can be explained by thermal heterogeneity alone, and the present study favors an interpretation of a superposition of the Iceland hotspot track and presently active tectonomagmatic processes at the central-eastern Greenland. The interpretation is consistent with the thermal structure of the mantle transition zone based on seismic RF analysis, which indicates a >20 km depression of the 410-km discontinuity around stations SCO and DBG, suggesting a temperature elevation of ~150 °C at a 400 km depth (Kraft et al., 2018). This temperature anomaly is consistent with a typical temperature anomaly estimated for mantle plumes (White and McKenzie, 1995). In agreement with the present results, a smaller amplitude mantle temperature anomalies are also seen in seismic receiver functions at stations SUM, HJO and SOE, and the thermal anomaly vanishes westwards.

High GHF promotes basal ice melting. The moderately high GHF anomaly (>70 mW/m² and possibly >90 mW/m²), where intensive ice melting may occur, extends inland below the ice sheet, and its western and northern boundaries cannot be established with the present data coverage on the Moho depth. The size, shape, and amplitude of the anomaly (Fig. 12) differ from an earlier estimate (Rogozhina et al., 2016): in Central Greenland the anomaly extends 200–300 km further south and corresponds to the major anomaly in the basal melt rate (MacGregor et al., 2016); at the east coast it sharply terminates north of station HJO. The anomaly may also extend far inland, by ca. 500 km, under the ice cap south of the Summit (Fig. 12b), which may have a strong effect on basal ice dynamics. However, the effect of basal melting on mantle dynamics through changes in the ice load (Jull and McKenzie, 1996; Carminati and Doglioni, 2010) is insignificant: a 50 mW/m² GHF anomaly, in case it is entirely spent on the ice sheet melting, will melt the base of the ice with a rate of only ca. 4.5 mm/year (for ice density of 0.934 t/m³ and heat of fusion of 334 kJ/kg).

The likely presence of active shallow magmatic processes in central-eastern Greenland, with a significant advective component of GHF such as is common in active rift systems, may explain the apparent discrepancy between the present ice dynamics and the reported GHF which are 10 times too low to explain basal melting (MacGregor et al., 2016). A high heat flux from the Earth’s interior enhanced by a hot fluid percolation above active magma chambers at the edge of the ice cap may have dramatic consequences for ice basal melting in the central-eastern Greenland, and may be an important contributor to the Northeast Greenland Ice Stream in Central Greenland (Fig. 6b).

5. Conclusions

The analysis of the Moho depth, equivalent topography, upper mantle temperature anomalies, lithosphere thermal thickness, and predicted geothermal heat flux in Greenland shows the following:

1. The absence of correlation between the present uplift rates and the anomalies in equivalent topography suggests that on-going vertical movements in the ice-free zone along the coasts reflect lithosphere flexure associated with glacial isostatic adjustments, but not with
2. The anomalies in equivalent topography can be explained by mantle density anomalies caused by thermal heterogeneity of the upper mantle.

3. Lithosphere thermal thickness varies from 200 to 270 km in the cratonic SW Greenland with the maximum thickness in the kimberlite province of West Greenland. Most of the Proterozoic lithosphere is 120–190 km thick with a gradual decrease in thickness from SW Greenland towards the Summit station in Central Greenland. The transition from the Archean to Proterozoic lithosphere is gradual and cannot be identified from the variations in lithosphere thermal thickness.

4. The anomalous belt with a 100–120 km thick lithosphere and increased GHF extends from the NW coast towards East Greenland and may result from the passage of the Iceland hotspot.

5. East Greenland has anomalous crustal structure, thin (50–100 km) lithosphere, high mantle temperatures and a strong GHF anomaly of >100 mW/m² centered in the Fjordland region. Thermal anomaly cannot be explained by conductive heat transfer and requires advective heat transfer such as above active magma chambers, that may significantly enhance basal ice melting. A possible extent of the anomaly ca. 500 km inland may have strong consequences for a contribution of ice melt to the ice-drainage system in central Greenland.

Acknowledgements

Greg Houseman and an anonymous referee are acknowledged for positive reviews. Research supported by grant DFF-1323-00053 of the Danish Fund for Independent Research (DFF-FNU).

Appendix A

A.1. Methods

A.1.1. Temperature anomalies in the lithosphere

Isostasy requires that deviation of equivalent topography from the isostasy predictions, AET, (assuming it is caused by temperature variations alone) is related to lithosphere thickness zLAB as:

$$AET = zLAB \frac{\rho_l - \rho(\Delta T)}{\rho_m}$$

where $\rho$ is in situ density of the lithosphere (crust and lithosphere mantle) density when $AET = 0$ km, $\rho_m$ is in situ density of sublithospheric mantle, and $\rho(\Delta T)$ is in situ lithosphere density at temperature anomaly $\Delta T$, which is defined with respect to an average lithosphere temperature at reference stations, where $AET = 0$ km. Mantle density at $T = 0$ °C is assumed to be $\rho_m(0) = 3.40 \text{ t/m}^3$, which gives in situ $\rho_m = 3.23 \text{ t/m}^3$ assuming mantle temperature of 1400 °C. Since

---

**Fig. 12.** Predicted geothermal heat flux in Greenland. Models 1 and 2 are explained in caption to Fig. 5. Values in black boxes – heat flux measured in boreholes and in hydrothermal complexes (see references in Rysgaard et al., 2018). Color-coded circles with numbers – calculated values for each station, used in interpolation. Dotted lines – various hypotheses for the Iceland hotspot. Black dashed line – proposed boundary of a region with a high geothermal heat flux (Rogozhina et al., 2016).
\[
\rho(\Delta T) = \rho_0(1-\alpha \Delta T)
\]
(2)
where \(\alpha = 3.5 \times 10^{-5}\) is coefficient of thermal expansion,

\[
\Delta T = \text{AET} \rho_0 / z_{LAB} \approx \rho_0
\]
(3)

Therefore, AET can be directly converted to thermal anomalies in the lithosphere, assuming all other contributions to AET are negligible and \(z_{LAB}\) is known. The potential effect of variation of crustal heat production by radioactivity is indirectly accounted for in Eq. (3) through its contribution to temperature anomalies \(\Delta T\). The AET values are calculated from the correlation between the Moho depth and the equivalent topography. The correlation can be constrained by global data on average continental Moho depth and density contrast across the Moho (Model 1), or by the best-fit to the regional data on the Moho depth and the equivalent topography (Model 2).

Average lithosphere density (crust plus lithosphere mantle) at \(T = 0^\circ C\) is

\[
\rho_0(0) = [2.85 \rho_{Moho} + 3.35(z_{LAB} - z_{Moho})] / z_{LAB}
\]
(4)

where \(z_{Moho}\) and \(z_{LAB}\) are thicknesses of the crust and the entire lithosphere, respectively, with the assumed average densities of 2.85 t/m\(^3\) for the bulk crust and 3.35 t/m\(^3\) for the bulk lithosphere mantle (which corresponds to typical densities of Proterozoic mantle, Poudjom Djomani et al., 2001). Assuming surface temperature of 0°C and temperature at the lithosphere base of 1400°C, the average lithosphere temperature at the reference stations is 700°C. Then in situ lithosphere density is

\[
\rho_0 = \rho_0(0)/(1-700 \alpha)
\]
(5)

For crustal thickness of 40 km and lithosphere thickness of 200 km \(\rho_0(0) = 3.25 \text{t/m}^3\) and \(\rho_0 = 3.17 \text{t/m}^3\). Substituting all parameters in (3) by values defined by (4) and (5) yields \(\Delta T\) as a function of the average densities and thicknesses of the crust and the lithospheric mantle:

\[
\Delta T = \text{constant} \text{AET}/(3.35 z_{LAB} - 5 z_{Moho})
\]
(6)

In case of a 200 km thick lithosphere with a 40 km thick crust and AET = 1 km, temperature anomaly in the upper 200 km is ca. 145°C with respect to the reference stations.

For Greenland, the Moho depth is known from the seismic receiver function interpretations (Dahl-Jensen et al., 2003b), and variations in this parameter have little effect on the calculated \(\Delta T\). Lithosphere thickness variations, however, have a crucial effect on \(\Delta T\). Since lithospheric thermal thickness is closely related to thermal structure of the lithosphere (Pollack et al., 1993; Artemieva, 2006), \(z_{LAB}\) can be determined in a self-consistent way from AET.

### A.1.2. Lithosphere thermal thickness

Temperature anomaly \(\Delta T\) defined by Eq. (1) is the difference in average temperature in the upper mantle (down to the lithosphere base at reference stations, \(z_{LAB\text{(ref)}}\)) with respect to reference stations. At stations where \(z_{LAB} < z_{LAB\text{(ref)}}\), the temperature profile down to reference LAB includes the lithosphere part with an average lithosphere temperature of 700°C and the asthenospheric part at mantle potential temperature of 1400°C between the local LAB (lithosphere base) and reference LAB. Likewise, at stations with \(z_{LAB} > z_{LAB\text{(ref)}}\), average lithosphere temperature is less than at reference stations. Therefore, for any station the average temperature in the upper mantle down to \(z_{LAB\text{(ref)}}\) is

\[
\text{Tave} = (700 z_{LAB} + 1400 (z_{LAB\text{(ref)}}-z_{LAB}))/z_{LAB\text{(ref)}}
\]
(7)

and

\[
\Delta T = \text{Tave} - 700 = 700 (1-\text{zLAB}/z_{LAB\text{(ref)}})
\]
(8)

A combination of Eqs. (6) and (8) allows for calculating lithosphere thickness \(z_{LAB}\) from the anomalous equivalent topography AET (Fig. 8a).

Importantly, lithosphere thickness cannot be determined when AET is extremely high such as at the east-central coast of Greenland. In such regions mechanism(s) other than purely thermal contribute to high equivalent topography.

\(z_{LAT}\) values calculated from AET can now be used in Eq. (6) to calculate temperature anomalies \(\Delta T\) in the upper mantle. Therefore, calculation of temperature anomalies and lithosphere thickness is based only on assumption on lithosphere thickness at reference stations. Clearly, the choice of another depth value for \(z_{LAT}\) will modify the results without changing the pattern. The sensitivity analysis for the method is in Artemieva (2018).

Reference geotherms for different tectonic settings (Pollack et al., 1993) constrain thickness of thermal lithosphere as a function of geothermal heat flux. This correlation allows for estimating geothermal heat flux GHF from known \(z_{LAB}\) (Fig. 8b).

### References


Fahnestock, M., Abdalati, W., Joughin, I., Brozena, J., Gogineni, P., 2001. High geo-