The crustal structure of central East Greenland—II: From the Precambrian shield to the recent mid-oceanic ridges

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SUMMARY
We present a 3-D crustal model of the East Greenland Fjord Region between 69°N and 74°N. The model covers the Precambrian shield and the Caledonian orogenic belt, the adjoining Devonian and Mesozoic basins, the continent–ocean transition and the Cenozoic oceanic areas as far as the Kolbeinsey and the Mohns mid-oceanic ridges. Existing seismic models of the crustal structure are extrapolated into adjacent areas using 3-D gravity modelling. For this purpose, we compile a new regional-scale Bouguer anomaly map. The Precambrian shield, west of the Caledonian orogen (approximately west of 32°W), shows a mean thickness of 35 km with only small-scale undulations. This thickness is at the lower limit of the global range in shield thickness. The Caledonian orogen exhibits a pronounced mountain root with overall crustal thicknesses up to 51 km. Beside the Urals, the East Greenland Caledonides are one of the two Palaeozoic mountain belts where a crustal root has preserved to the present day. Continuation of the crustal model to the east, beyond the continent–ocean transition, yielded thicknesses of the crystalline oceanic crust from 9 km near the Kolbeinsey Ridge to 3 km west of the Mohns Ridge. Differences in the thermal structures of the old continental and the young oceanic lithosphere are responsible for the low-density mantle beneath the oceanic crust, which is also demonstrated by 3-D gravity modelling.

Key words: continental crust, crustal structure, East Greenland, gravity, oceanic lithosphere.

1 INTRODUCTION
The conjugate continental margins of the North Atlantic are formed by the East Greenland Fjord Region between 69°N and 74°N, together with northeast and southeast Greenland, and the coasts of Scandinavia and the British Isles. The Caledonian orogenic belt, formed in Silurian times during the closure of the Iapetus Ocean by continent–continent collision (e.g. Escher & Pulvertaft 1995; Henriksen et al. 2000), is the dominant feature of the region. The western termination of the fold belt in Greenland is covered by a permanent ice sheet; only few tectonic windows expose the Caledonian foreland. During the ensuing Devonian extensional collapse and long-term rifting, several sedimentary basins developed which are separated from the Caledonides by multiple faults. Large amounts of flood basalts were generated during the opening of the North Atlantic in Tertiary times, they obscure the southern termination of the Caledonides at 70°N along the southern coast of Scoresby Sund.

The East Greenland side remained geophysically poorly investigated for a long time, whereas the European margins were explored at an early stage. The long fjords in East Greenland provided good opportunities for land–sea seismic experiments, hence several seismic refraction surveys were conducted in the region between 1988 and 1994 (Weigel et al. 1995; Jokat et al. 1995, 1996). These yielded detailed models of the continental crust in the area of the orogenic belt and the sedimentary basins (Fechner & Jokat 1996; Mandler & Jokat 1998; Schlindwein & Jokat 1999; Schmidt-Aursch & Jokat 2005, this issue). The seismic models were complemented by analyses of gravity and magnetic data (Schlindwein & Meyer 1999; Schlindwein & Jokat 2000). The East Greenland shelf was investigated mainly by seismic reflection surveys (see Larsen 1990, for a review). A few expanding spread profiles (Hinz et al. 1987; Mutter & Zehnder 1988) and a couple of wide-angle lines (Weigel et al. 1995; Kodaira et al. 1998) cover the continent–ocean transition and the oceanic crust east of the shelf edge. Apart from recent seismological studies (Dahl-Jensen et al. 2003), little is known up to now about the architecture of the continental crust to the west of the seismic transects (west of 30°W).

The objective of this study was to create a regional three-dimensional (3-D) crustal model for central East Greenland between 69°N and 74°N. We compiled a new map of Bouguer anomalies in order to perform 3-D gravity modelling. The modelling was based mainly on the results of previous seismic studies, introduced in Schmidt-Aursch & Jokat (2005), and led to a crustal model which spans the central Precambrian shield of Greenland, the Caledonian...
fold belt and the sedimentary basins, the continent–ocean transition and the deep sea areas as far as the nearest recent mid-oceanic ridges, the Kolbeinsey and the Mohns ridges. The results of these investigations are presented here.

2 GRAVITY MODELLING

Several existing wide-angle seismic models provide a detailed view of the crustal architecture in the East Greenland Fjord Region (e.g. Weigel et al. 1995; Schlindwein & Jokat 1999; Schmidt-Aursch & Jokat 2005). A regional, 3-D gravity model interpolates this knowledge between the profiles and extrapolates it into regions that are difficult to access. The density model covers the Precambrian shield of central Greenland, the Caledonian fold belt and the Devonian and Mesozoic basins, and the Cenozoic areas across the continent–ocean boundary to the east as far as to the Kolbeinsey and Mohns mid-oceanic ridges (Fig. 1).

2.1 Compilation of the Bouguer anomaly map

A new, regional-scale Bouguer anomaly map was compiled in order to build the gravity model. The majority of the Bouguer gravity data were provided as a grid by the National Survey and Cadastre, Denmark (Brozena et al. 1993; Forsberg & Kenyon 1994). This grid, with a resolution of 10° × 5', covers all of Greenland and adjacent areas as far as 10°E. Due to the lack of a high-quality digital terrain model, no terrain correction was applied to the grid, and hence errors of 30–40 mGal (1 mGal = 10−5 m s−2) might be expected in rough terrain (Forsberg 1991). East of 10°E, free air anomalies derived from satellite data were used (Laxon & McAdoo 1998) to calculate Bouguer anomalies, for consistency also without inclusion of a terrain correction, using a slab density of 2.67 × 103 kg m−3 and water depths taken from the IBCAO grid (Jakobsson et al. 2000). Errors at the boundary between the two grids were mostly less than 10 mGal. Since the Bouguer anomalies span more than 450 mGal, no attempt was made to correct these.

Fig. 1 shows the new Bouguer anomaly map. In central Greenland, in the area of the Greenland Shield, the anomaly ranges from −80 to −120 mGal and has small-scale variations of approximately ±25 mGal. Eastwards, the pronounced negative anomaly of the Caledonian mountain belt dominates the map. Values below −200 mGal were reached at the western ends of the deep seismic profiles of Gåsefjord (GF), Nordvestfjord (NV) and Dickson Fjord (DF). The minimum of −230 mGal is located between the two seismic profiles of Kejser Franz Joseph Fjord (KFJF) and Kong Oscar Fjord (KOF). The Bouguer anomaly rises continuously to 60 mGal across the East Greenland sedimentary basins (e.g. the Jameson Land Basin, JLB) and the adjoining shallow shelf areas. The highest values, of more than 250 mGal, can be found in the deep-sea area of the Greenland Basin. The mid-oceanic Kolbeinsey and Mohns ridges, with their pronounced relief, imprint the mapped anomalies as well as the Jan Mayen Ridge and the Jan Mayen Fracture Zone.

2.2 Modelling procedure

We used the interactive IGMAS software (Götze & Lahmeyer 1988; Schmidt 2000), which calculates the potential field due to triangulated polyhedrons. The 523 600 km2 density model is defined
by nearly 4900 vertices on 31 parallel west–east trending vertical planes with a spacing ranging between 3 and 25 km in the centre of the model and 20 and 46 km at the outer margins. The model was expanded to all sides to avoid edge effects. Triangulated surfaces were calculated between these cross-sections, which bound bodies with constant densities. The model consists of 14 layers representing continental and oceanic crust and mantle (Table 1). The water layer was modelled with a density of $2.67 \times 10^3$ kg m$^{-3}$ resulting from the Bouguer correction. Densities of $2.00–2.55 \times 10^3$ kg m$^{-3}$ were assigned to the continental and oceanic sediments (Schlindwein & Jokat 2000; Nafe & Drake 1957). The continental crystalline crust was divided into six units based on wide-angle seismic profiles (Schlindwein & Jokat 1999; Schmidt-Aursch & Jokat 2005). The mean seismic velocities of each layer were converted into densities after Christensen & Mooney (1995), this yielded a density range of $2.67–3.10 \times 10^3$ kg m$^{-3}$. For the oceanic crystalline crust a mean density of $2.9 \times 10^3$ kg m$^{-3}$ (Seibold 1996) was assumed.

Additional boundary conditions were incorporated in the model to reduce the inherent ambiguity of gravity modelling: The extent of the continental sedimentary basins and the location of the continent–ocean boundary have been picked from the geological map of Greenland (Esher & Pulvertaft 1995). Six deep seismic profiles (Schlindwein & Jokat 1999; Schmidt-Aursch & Jokat 2005) provide the structure of the continental crust (Fig. 1). Information about the sediment and crustal thicknesses of the oceanic crust was given by Weigel et al. (1995) for the area from the Scoresby Sund (ScS) to Kolbeinsey Ridge and Kodaira et al. (1998) for the Kolbeinsey Ridge itself (Fig. 1). The seismic investigations of the Mohns Ridge by Klingelhöfer et al. (2000), which were located east of the northeastern edge of the model, were also used. We modelled the continent–ocean boundary in a simple way without any transitional zone, the high horizontal velocity gradient of which would require several additional density bodies. These additional bodies could propagate into model uncertainty, because existing velocity models about the transitional zone in the east Greenland Fjord Region are diverse and not detailed enough (Hinz et al. 1987; Mutter & Zehnder 1988; Weigel et al. 1995).

The intention of this work was to explain the long-wavelength Bouguer anomaly, but not detailed modelling of small-scale structures. The long-wavelength anomaly depends, to first order, on crustal thickness, so an attempt was made to fit the Bouguer anomaly only by varying the thickness of the continental lower crust and the oceanic crust. Starting from the seismic profiles, we kept all crustal densities fixed and interpolated the layer boundaries of the upper and middle continental crust between the profiles. West of the seismic profiles, in the area of the ice shield, the boundaries were simply extrapolated because no crustal model exists there. Forward modelling of the continental lower crust and the oceanic crust resulted the Moho depth beyond the seismic models. Owing to the 3-D character of the modelling technique, absolute values for errors cannot be provided as they strongly depend on the shape and extension of the considered structures. For example, changes in layer boundary depths of a certain amount can result in very different variations in gravity. Additionally, the errors in density and boundary depths of the upper layers have to be added. Therefore, we only estimate the maximum Moho depth errors for the continental crust to be ±5 km and the mean error to ±3 km. For the oceanic crust, the errors are smaller, due to the simpler model architecture, the mean error of Moho depth takes a value of approximately ±2 km.

### Table 1. List of body parameters used for 3-D gravity modelling.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Mean velocity (km s$^{-1}$)</th>
<th>Density ($10^3$ kg m$^{-3}$)</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>1.50</td>
<td>2.67</td>
<td>1</td>
</tr>
<tr>
<td>Cont. sediment</td>
<td>4.20</td>
<td>2.30</td>
<td>2</td>
</tr>
<tr>
<td>Cont. sediment</td>
<td>5.70</td>
<td>2.55</td>
<td>2</td>
</tr>
<tr>
<td>Cont. upper crust</td>
<td>5.87</td>
<td>2.67</td>
<td>3</td>
</tr>
<tr>
<td>Cont. upper crust</td>
<td>6.15</td>
<td>2.77</td>
<td>3</td>
</tr>
<tr>
<td>Cont. middle crust</td>
<td>6.40</td>
<td>2.85</td>
<td>3</td>
</tr>
<tr>
<td>Cont. lower crust</td>
<td>6.60</td>
<td>2.93</td>
<td>3</td>
</tr>
<tr>
<td>Cont. lower crust</td>
<td>6.80</td>
<td>3.00</td>
<td>3</td>
</tr>
<tr>
<td>Cont. lower crust</td>
<td>7.15</td>
<td>3.10</td>
<td>3</td>
</tr>
<tr>
<td>Cont. mantle</td>
<td>7.97</td>
<td>3.30</td>
<td>4</td>
</tr>
<tr>
<td>Oceanic sediment</td>
<td>2.10</td>
<td>2.00</td>
<td>5</td>
</tr>
<tr>
<td>Oceanic crust</td>
<td></td>
<td>2.90</td>
<td>6</td>
</tr>
<tr>
<td>Oceanic mantle</td>
<td></td>
<td>3.20</td>
<td>7</td>
</tr>
<tr>
<td>Oceanic mantle</td>
<td></td>
<td>3.15</td>
<td>7</td>
</tr>
</tbody>
</table>

### 2.3 3-D density model

Fig. 2 shows a cross-section (A–A′) of the 3-D density model, which runs nearly parallel to the Fønfjord seismic profile (FF, Fig. 1). Between km 390 and 600, the model is constrained by this seismic fracture trend with its maximum Moho depth of 38 km (Schmidt-Aursch & Jokat 2005). At km 390, Dahl-Jensen et al. (2003) derived a Moho depth of only 32 km from receiver function analysis. But errors for this station might be large due to poor data quality, therefore we deliberately didn’t include it in our density model. West of km 350, the Moho rises up to 35 km depth and flattens there. In this case, the Moho depth errors for the continental crust to be ±5 km and the mean error to ±3 km. For the oceanic crust, the errors are smaller, due to the simpler model architecture, the mean error of Moho depth takes a value of approximately ±2 km.
area, the Bouguer anomaly shows short-wavelength undulations, which could not be modelled with Moho variations, giving rise to residual misfits of ±30 mGal. Changes in the surface geology, which is hidden by the ice shield, might be the reason for these misfits. East of the seismic profile and the Jameson Land Basin (km 600) there is another region of large misfits. Although a 7 km thick root with a Moho depth of 28 km (Dahl-Jensen et al. 2003) was modelled beneath the crystalline Liverpool Land (LL, Fig. 1), misfits in the region of −35 mGal remain. This residual could not be modelled by Moho variations; there might also be density variations in the upper crust. The crust thins significantly east of the continent–ocean transition to 9 km near the Kolbeinsey Ridge, where there is a sedimentary cover of approximately 1 km near the ridge (Kodaira et al. 1998). The first effort to model the oceanic realm with the standard mantle density of $3.3 \times 10^3$ kg m$^{-3}$ and typical seismic velocities, implying typical densities, and so we modelled a density anomaly in the mantle. A reasonable fit for the Bouguer anomaly was achieved with anomalously low densities of $(3.2 \pm 0.05) \times 10^3$ kg m$^{-3}$ below the oceanic crust and $(3.15 \pm 0.05) \times 10^3$ kg m$^{-3}$ beneath the mid-oceanic ridges.

A second cross-section (B–B′) is displayed in Fig. 3. The Kejser Franz Joseph Fjord seismic profile (KFFJ, Fig. 1) is parallel to the cross-section and constrains the model between km 490 and 765 (Schlindwein & Jokat 1999). Beneath the Caledonian fold belt, the Moho decreases to a maximum depth of 49 km and rises again in the western part of the cross-section to 35 km. Again, small-wavelength undulations in the Bouguer anomaly could not be modelled solely with Moho variations. The oceanic crust in this region is very thin, at about 4 km near the Mohrins Ridge with a sedimentary layer of less than 1 km (Klingelhöfer et al. 2000). Again, we modelled a mantle density somewhat lower than normal to account for the Bouguer anomaly of more than 200 mGal.

One main result of the 3-D gravity modelling is a regional model of crustal thickness. Fig. 4 shows a perspective view of the triangulated surface between the crust and mantle. The Moho depth and also the crustal thickness without the overlying inland ice is around 35 km with small-scale variations beneath the Precambrian shield in the western part of the model. Eastwards, in the realm of the Caledonian orogenic belt, the crust shows a distinct root structure. In the northern and southern parts of the model the root is very pronounced with Moho depths as deep as 49 km, whereas in the middle the root seems to flatten up to 40 km. This leads to crustal thicknesses of 40–51 km, including the Caledonian mountains where heights are up to 2.5 km. The continuous rise of the crust–mantle boundary below the sedimentary basins and the continent–ocean transition, to depths and thicknesses of approximately 20 km, terminates in the eastern areas of the oceanic crust near the mid-oceanic ridges. The Moho depth is about 10 km south of the Jan Mayen Fracture Zone, underneath the Kolbeinsey Ridge, which corresponds to a crustal thickness of 9 km. The minimum Moho depth of 6 km, and minimum crustal thickness of 4 km, is reached in the far northeast of the model close to the Mohrins Ridge.

3 TECTONIC CONSIDERATIONS

3.1 The Precambrian shield

The Precambrian shield of Central Greenland is covered by an ice sheet more than 3 km thick (Escher & Pulvertaft 1995), which makes any seismic investigations of the crustal fabric logistically difficult. Due to the absence of deep seismic sounding data in that area, we extrapolated our seismically constrained density model of the Caledonian fold belt into the shield by keeping the densities fixed and only varying the Moho depth. This first-order estimate yielded a mean crustal thickness of about 35 km with only small-scale undulations, and a smooth Moho topography. Worldwide data compilations for Precambrian shields show large variations in crustal thicknesses, as do determinations available in the literature. For example, Meissner (1986) quoted values of 40–50 km for all ages, Durrheim & Mooney (1994) divided shield thicknesses into Proterozoic (40–55 km) and Archaean (27–40 km) ones, Christensen & Mooney (1995) calculated a mean thickness of 41.5±5.8 km and

Figure 3. Density model for profile B–B′: MR, Mohrs Ridge. For additional explanations see Fig. 2.
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Figure 4. Perspective view of the surface between crust and mantle (Moho) in the 3-D density model: COB, continent–ocean boundary; K, Kolbeinsey. The surface was interpolated between 31 vertical planes with a spacing of 3–46 km. On top the coastline is marked. The continent–ocean boundary is drawn as a dashed line. Dotted lines mark the area in Fig. 6. Vertical exaggeration is ×10.

Zandt & Ammon (1995) estimated a value of 36.9 km for all shields. Compared with these numbers, the 35 km thick Proterozoic crust of central East Greenland is either rather thin (Meissner 1986; Durrheim & Mooney 1994; Christensen & Mooney 1995) or exhibits normal crustal thickness (Zandt & Ammon 1995).

However, recent results of a receiver-function analysis study provided the first reliable overall crustal thickness of 50 ± 2 km in the Precambrian area spanned by our density model (Fig. 1, station SUM; Dahl-Jensen et al. 2003). This value includes the 3 km thick ice sheet and hence corresponds to a Moho depth and crustal thickness without ice, like that used in this study, of about 47 km. Compared with the thickness of 35 km resulting from gravity modelling, the 12 km difference in Moho depth implies a significant density variation. A difference of at least 7 km remains even if taking all possible uncertainties into account. Fig. 5 shows, as an example, the former profile of Kejser Franz Joseph Fjord (top), which is located just 5 km away from station SUM, and a possible alternative model with a deeper Moho (bottom). To adjust the Bouguer anomaly, a lower crust more than 20 km thick with high densities ((3.10–3.20) × 10^3 kg m^{-3}) has to be assumed. This would endorse the model of Durrheim & Mooney (1994), which predicts high crustal thicknesses (40–55 km) and a thick, high-speed, high-density basal layer in the lower crust for Proterozoic shields. Dahl-Jensen et al. (2003) also found a mean Poisson’s ratio of 0.28 for the entire crust, which differs significantly from the much lower values (0.22–0.25) of the Caledonian orogenic belt and adjoining rifted areas. This leads to the possibility that laterally inhomogeneous densities between the Proterozoic shield and the younger realms of central East Greenland exist not only in the lower crust but also in the upper and middle crust. Hence, without further detailed knowledge about the crustal architecture of the Precambrian shield, no satisfying density model can be calculated for the inner part of Greenland.

3.2 The Caledonian orogen

Wide-angle seismic profiles over the Caledonian mountain belt (Schlindwein & Jokat 1999; Schmidt-Aursch & Jokat 2005) resulted seismic velocities which are well within the range typical for Palaeozoic orogens (e.g. Meissner 1986; Christensen & Mooney 1995). The westward extent of the Caledonides is, due to the Greenlandic ice sheet which hides the surface geology, highly speculative. Caledonian foreland is exposed mainly in two tectonic windows west of the seismic profiles of Nordvestfjord and Fønfjord (e.g. Escher & Pulvertaft 1995; Henriksen et al. 2000), but the orogen itself might very well continue westwards. There were no variations in Poisson’s ratio found within the Caledonides (Schmidt-Aursch & Jokat 2005), therefore we decided to expand the density model constrained by seismic refraction data further to the west. The Nordvestfjord seismic profile (NF, Fig. 1) gave clear evidence for a crustal root, and a complex root structure with two troughs and a relatively flat area in between arose from the 3-D density modelling. Fig. 6 shows a comparison of surface bedrock topography (Ekholm 1996) and Moho depth, which features a good negative correlation, i.e. areas with high surface mountains are located directly above areas with large Moho depths (see also Figs 2 and 3). The maximum overall crustal thickness for the orogen is herewith 51 km.

For a long time, all Palaeozoic mountain belts were considered to have no remaining crustal root (Meissner 1986). This is true for the Scandinavian and British Caledonides, the North American Appalachian orogenic belt and the Variscan mountains in western Europe. The Scandinavian Caledonides remain especially enigmatic as a pronounced negative Bouguer anomaly of −80 mGal suggests that a mountain root exists, but seismic measurements revealed a Moho depth of no more than 38 km beneath the fold belt and no evidence for a crustal root (Meissner 1986; Kinck et al. 1991).
Instead, the Scandinavian Bouguer anomaly is regarded as the consequence of a less dense mantle (e.g. Theilen & Meissner 1979; Bannister et al. 1991) or the combination of less dense mantle and crust (Dyrelius 1985). The reverse was found in the Urals. Here, a small Bouguer anomaly of no more than $-40$ mGal is found in combination with crustal thicknesses up to 65 km, including an approximately 15–20 km thick root (e.g. Carbonell et al. 1996; Juhlin et al. 1996; Knapp et al. 1998). The reduced Bouguer anomaly is attributed to an anomalously high-density upper crust (Thouvenot et al. 1995). The main Uralian crustal root and the highest surface elevations are offset, which is attributed to a remnant effect of Palaeozoic deformation (Berzin et al. 1996). The East Greenland Caledonides show no such peculiarities; global standard crustal and mantle seismic velocities and densities (e.g. Meissner 1986; Christiansen & Mooney 1995) affirm the seismically constrained crustal thicknesses and Bouguer anomalies. The high symmetry of surface topography and the root fabric suggests that both have their origin in a single orogeny and both were affected in a similar way by the Devonian extensional collapse.

### 3.3 Oceanic crust and mantle

The 3-D gravity modelling yielded oceanic crustal thicknesses between 9 km near the Kolbeinsey Ridge and 4 km close to the Mohns Ridge. The model includes a sedimentary layer approximately 1 km thick, that is thinning in the immediate vicinity of the mid-oceanic ridges. The thickness of the crystalline oceanic crust differs significantly from the mean global value of $7.1 \pm 0.8$ km (White et al. 1992) both north and south of the Jan Mayen Fracture Zone. North of the Jan Mayen Fracture Zone the oceanic crust is uniformly thin between Mohns Ridge and the continent–ocean transition zone. Very slow spreading at the Mohns Ridge (16 mm yr$^{-1}$ full spreading rate) is hypothesized to cause conductive cooling of the upper mantle (by approximately $20^\circ$C) that leads to reduced melt production and hence reduced crustal thickness (Klingelhöfer et al. 2000). The uniform crustal thickness indicates a constant temperature and spreading ambience over long times. Kodaira et al. (1998) explained the increased crustal thickness south of the Jan Mayen Fracture Zone between the slow spreading Kolbeinsey Ridge (15–20 mm yr$^{-1}$ full spreading rate) and the Jan Mayen Microcontinent as a consequence of slightly increased ($20–60^\circ$C) mantle temperatures caused by the Icelandic hotspot. This rises the question of why there is no influence of the Icelandic hotspot visible at the nearby Mohns Ridge and the Greenland Basin on the other side of the Jan Mayen Fracture Zone. The pronounced change of oceanic crustal thickness of 5 km within less than 500 km distance must therefore be the result of regional independent mantle processes. The poor deep seismic
coverage across the Jan Mayen Fracture Zone and in the Greenland Basin does not allow a more detailed interpretation of the gravity data.

While the continental part of the 3-D model shows normal mantle densities of $3.3 \times 10^3$ kg m$^{-3}$, the oceanic mantle had to be modelled with reduced densities of $(3.15–3.2) \times 10^3$ kg m$^{-3}$. This was necessary to fit the large positive Bouguer anomaly as the other modelling variables, crustal densities and thicknesses, were well constrained by wide-angle seismic data. Breivik et al. (1999) conducted 2-D gravity modelling across the passive margin of northern Norway in the Barents Sea and also faced the problem of an inconsistency between the Bouguer anomaly and known crustal thicknesses and densities. Modelling of the thermal structure of the margin yielded different upper mantle temperatures in the old, cold, continental lithosphere and the young, warm, oceanic lithosphere. Breivik et al. (1999) converted the temperatures to densities that were consistent with the Bouguer anomaly. We did not undertake thermal modelling, but our oceanic mantle density of $3.2 \times 10^3$ kg m$^{-3}$ is comparable with the mean density of the model of Breivik et al. (1999). The modelled temperature differences between the continental and the oceanic lithosphere are ubiquitous at young passive continental margins, and the passive rifted margins of central East Greenland and Svalbard (Ritzmann et al. 2004) support this theory.

4 CONCLUSIONS

We present a regional 3-D crustal model of central East Greenland between 69°N and 74°N. It spans the eastern parts of the Precambrian shield, the adjacent Caledonian fold belt, the rifted areas of the Fjord Region with the Devonian and Mesozoic basins, the continent–ocean transition and the adjoining oceanic areas as far as the Kolbeinsey and the Mohns mid-oceanic ridges. A new regional-scale Bouguer anomaly map was compiled from different sources and 3-D gravity modelling was conducted to expand existing crustal models, derived from wide-angle seismic data, into areas that are difficult to access. Comparisons of the crustal architecture of central East Greenland with other regions of a similar age highlights analogies and differences. The thickness of the Precambrian shield was derived by 3-D gravity modelling, and at 35 km with only small-scale variations it is situated at the lower limit of the global range. Seismic constraints are needed to confirm if this is really the case. Seismically modelled Moho depths of 48 km indicate a crustal root of the Caledonian orogen, whose dimensions are shown by gravity modelling. The existence of a mountain root with crustal thicknesses up to 51 km is in contrast to the European Caledonides.

Extension of the known crustal structure to the east of the continent–ocean transition by gravity modelling yielded thicknesses of the crystalline oceanic crust between 3 km near the Mohns Ridge and 9 km west of the Kolbeinsey Ridge. These thickness variations of the oceanic crust, which is thicker than average south of the Jan Mayen Fracture Zone and thinner than average north of it, are in our interpretation a consequence of independent mantle temperatures. Warmer mantle beneath the Kolbeinsey Ridge leads to a higher melt production and hence a thicker crust. In contrast, conductive cooling at the Mohns Ridge is reflected in smaller amounts of melt and a thin oceanic crust. Differences in the thermal structure of the old continental and young oceanic lithosphere are responsible for the low-density mantle beneath the oceanic crust, which is necessary in order to make our 3-D gravity model consistent with seismically determined crustal thickness. Finally, the 3-D gravity results are a first step towards the description of the crustal thickness of the interior of Greenland. Seismic refraction lines are needed to provide better constraints on the crustal fabric there.

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