A 3D regional crustal model of the NE Atlantic based on seismic and gravity data

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Abstract: We present a 3D regional crustal model for the North Atlantic, which is based on the integration of seismic constraints and gravity data. The model addresses the crustal thickness geometry, and includes information on sedimentary thickness, the presence of high-velocity zones in the lower crust, and information on the crustal density distribution in the continental and oceanic domains. Using an iterative forward- and inverse-modelling approach, we adhere to the seismic constraints within their uncertainty, but manage to enhance the crustal geometry in areas where seismic data are sparse or absent. A number of basins are resolved with more detail. Recently released seismic reflection data beneath the NE Greenland Shelf allowed for a major improvement of the crustal thickness estimates. Estimated Moho depths beneath the basins there vary between 15 and 25 km, which is compatible with the conjugate Norwegian margin. A major lower-crustal seismic velocity anomaly in the vicinity of the Greenland–Iceland–Faroe Ridge complex is supported by density modelling. We discuss the validity and uncertainties of our model assumptions and discuss the correlation with the main structural elements of the North Atlantic.

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Knowledge about the location of the crust–mantle interface (Moho) and the structure of the crust is essential to understand the geological evolution and present-day tectonic setting. The definition of the Moho interface is strongly based on seismic interpretations because it usually shows a strong signal in the data. The same applies to the identification of the top of the basement and different sedimentary sequences. However, seismic data are spatially limited to profiles of individual surveys. In addition, imaging of these interfaces is not always possible because of signal loss (e.g. sub-basalt imaging: Martini & Bean 2002) or because of poor data quality. Data gaps must then be interpolated to facilitate a reconstruction of the geological setting.

Gravity data are often used to validate seismic interpretations as the main density contrast in the Earth is expected between the crust and mantle. High-resolution shipborne data are often acquired simultaneously with seismic surveys, while airborne data are collected on designated surveys. In addition, gravity models generated from satellite data (either satellite altimetry or by direct measurements) cover the entire Earth almost uniformly and are publicly available (e.g. Pail et al. 2010; Bouman et al. 2013; Andersen et al. 2014).

We present results of a 3D regional modelling study of the North Atlantic between Greenland and NW Europe. During the Northeast Atlantic Geoscience Tectonostratigraphic Atlas project (NAG-TEC), a joint project between the geological surveys of NW Europe, an unprecedented amount of geophysical data, including seismic and potential field data, were compiled in one comprehensive atlas (Hopper et al. 2014). This source of information was used for the compilation of a novel, seismic-based depth to Moho map as presented by Funck et al. (2016), followed by the construction of a lithospheric model in the NE Atlantic region using gravity data. On the scale of the North Atlantic, gravity measurements can be used to study the extent and depth of sedimentary basins, as well as variations in crustal thickness. In order to build the 3D regional model, forward- and inverse-modelling techniques were applied. We used gravity inversion to verify the seismic interpretations of the Moho and to improve the estimations in regions with no seismic coverage. The procedure is described in

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the following, together with a discussion of the resulting implications.

Data

The lithospheric 3D model is based on seismic interpretations, well information and potential field data, as well as on structural information. The study area is outlined by the yellow polygon in the elevation map in Figure 1a.

Seismic data

A wealth of seismic surveys covers the NE Atlantic Ocean and adjacent areas. These data include seismic refraction/wide-angle reflection surveys and receiver functions. During the NAG-TEC project, these data were compiled in one atlas and provided the basis for a new Moho and crustal structure compilation (Funck et al. 2014). Supplemented by elevation and gravity data, the seismic interpretations were interpolated and extended into regions without seismic coverage using kriging techniques (Funck et al. 2016).

An alternative model based on a similar, although less comprehensive, seismic database has been published by Artemieva & Thybo (2013). This model is, unfortunately, not yet available digitally, but differences to the compilation by Funck et al. (2016) are mostly subtle. To some degree, the differences are related to variations in the database but also depend on the inclusion of seismic reflection data, as done by Artemieva & Thybo (2013). The use of reflection data requires a proper velocity function for the time-to-depth conversion, which is not always available.

Gravity data

A variety of shipborne and airborne gravity data exist for the North Atlantic (see Olesen et al. 2010 and references therein). Available non-commercial data cover most of the onshore areas, except Greenland where coverage is very sparse. Offshore data are mainly available in the shelf regions in the eastern part of the NE Atlantic but sparser in the central part and lacking in the western part (Fig. 1a). As gravity data are measured from diverse platforms and can be rather old, the data differ in quality and resolution.

Therefore, in order to have a consistent and continuous set of gravity data, we use the satellite-derived DTU10 global gravity field model. A detailed description of the processing is given in Andersen (2010) and Andersen et al. (2010). Offshore, the DTU10 model is based on satellite altimetry, onshore it relies on the EGM2008 model (Pavlis et al. 2012), which incorporates most of the onshore terrestrial data shown in Figure 1a. In the North Atlantic, satellite altimetry-derived free-air data have an estimated accuracy and resolution of approximately 3 mGal over 10–15 km wavelength (Andersen et al. 2010), compared to approximately 1 mGal over 5–10 km wavelength for shipborne data (Dragoi-Stavar & Hall 2009). In the Arctic Ocean, the difference between satellite altimeter data and airborne surveys is around 2–3 mGal on a regional scale (Childers et al. 2001). Hence, instead of compiling a new gravity dataset, we use the existing DTU10 model as it provides an acceptable accuracy for our regional modelling study.

The DTU10 model provides free-air gravity (Fig. 1b), which is dominated by topography/bathymetry. Spreading ridges (extinct and active) show up clearly in the data (e.g. Aegir Ridge, Mohns Ridge, Knipovich Ridge, Kolbeinsey Ridge and Reykjanes Ridge), as do major tectonic features such as shelf edges (which may indicate the continent–ocean transition (COT)) and major fracture zones (e.g. Jan Mayen Fracture Zone, Senja Fracture Zone, Greenland Fracture Zone and Bight Fracture Zone).

From the free-air anomaly, the Bouguer anomaly was calculated (Fig. 1c). The Bouguer correction takes into account a standard rock density of 2670 kg m$^{-3}$ for the onshore areas. The Greenland ice sheet thickness was taken from the ETOPO1 Global Relief Model (Amante & Eakins 2009) and is corrected with an ice density of 940 kg m$^{-3}$. Offshore, the water column is filled with sediments with a density of 2200 kg m$^{-3}$. This approach eliminates the strong density contrast between water and sediments at the seafloor. The Bouguer anomaly highlights crustal thickness variations across the study area. Positive values correlate with thin but denser oceanic crust, whereas negative values appear in areas of thick, less dense continental crust.

Finally, the tilt derivative (Miller & Singh 1994) of the isostatic anomaly was calculated. The isostatic anomaly is obtained by removing the gravimetric effect of an isostatic model of topographical relief from the Bouguer anomaly. Here, an Airy–Heiskanen root was used, with a normal reference crustal thickness of 32 km and a density contrast of 400 kg m$^{-3}$ at the crust–mantle interface. The gravity signal from the base of the crust has a long wavelength character. Correcting for these effects enhances crustal anomalies and intracrustal density variations. The tilt derivative is then calculated from the vertical and horizontal gradients of the anomaly, and enhances the edges of these intra-crustal anomalies and highlights linear features (Fig. 1d). It also delineates major structural elements more clearly than the free-air anomaly and...
Fig. 1. Elevation and gravity maps based on the DTU10 global gravity field model (Andersen 2010; Andersen et al. 2010). (a) Bathymetry and coverage with airborne, seaborne and terrestrial gravity data (red dots and lines). The yellow polygon outlines the study area. (b) Free-air anomaly. (c) Bouguer gravity anomaly, including a correction for ice; (d) Tilt derivative of the isostatic anomaly considering an Airy–Heiskanen root, with a reference crustal thickness of 32 km and a density contrast of 400 kg m$^{-3}$ at the crust–mantle interface. Red lines indicate the interpretation of the continent-ocean boundary (COB) after Funck et al. (2014); blue lines show COB interpretations after Escher & Pulvertaft (1995), Breivik et al. (1999), Scott (2000), Mosar et al. (2002a, b), Sigmond (2002), Hamann et al. (2005), Kimbell et al. (2005), Tsikalas et al. (2005), Gaina et al. (2009), Voss et al. (2009) and Peron-Pinvidic et al. (2012). BFZ, Bight Fracture Zone; AR, Aegir Ridge; GFZ, Greenland Fracture Zone; JMFZ, Jan Mayen Fracture Zone; KnR, Knipovich Ridge; MR, Mohns Ridge; RR, Reykjanes Ridge; SFZ, Senja Fracture Zone.
can, for example, be used to interpret the COT. Recent interpretations of the continent–ocean boundary (COB) are shown by polygons in Figure 1d. The isostatic anomaly was not directly used during the inversion procedure.

**Methodology**

A 3D density model covering an area of 2600 × 6400 km was built, with a vertical extent of 250 km. The initial model was set up using the GM-SYS 3D modelling software (Geosoft 2014). The model is defined by stacked grids that represent major geological boundaries where density contrasts occur. The resolution of the grids is equal for all horizons and, for this regional study, a grid spacing of 10 km was chosen in order to preserve small-wavelength structures in the upper crust.

The model was later transferred to the forward-modelling software IGMAS+ (Schmidt et al. 2010), which facilitates easier access to the model geometry than provided by GM-SYS 3D for the grid-based model. Main components in IGMAS+ models are vertical sections that span the entire model and form a 3D model via triangulation. The model geometry can be modified interactively along these sections. For the structural inversion (Moho depth) and lateral density inversion, the 3D model was transferred back to GM-SYS 3D. Although the two software packages have a different definition of model geometry, the calculated responses are identical, except for some numerical noise.

The model is based mainly on the information described above and some external information for supplementary modelling. Initial key horizons are topography, top basement and Moho. Below, a description of all model layers is given, together with the assigned layer densities (see also Table 1).

**Initial model**

The topography/bathymetry is taken from the DTU10 model and is based on satellite altimeter measurements. Higher-resolution data are available, for example, from the ETOPO1 (Amante & Eakins 2009) or the IBCAO (Jakobsson et al. 2012) compilations. However, the DTU10 model is consistent with the gravity model used and provides the necessary resolution. For the modelling, the Bouguer-corrected gravity data were used and, therefore, the corresponding density of 2200 kg m$^{-3}$ was assigned to the water column. This is equivalent to replacing the water column with sediments to eliminate the effect of bathymetry and topography.

We did not differentiate between individual sedimentary layers but use the total sediment sequence thickness from Funck et al. (2014) and Hopper et al. (this volume, in review) to define the depth to the top of the basement. We assigned a density–depth function with an exponential trend that increases from 2200 kg m$^{-3}$ at the seabed to a maximum of 2700 kg m$^{-3}$ at a depth of 8 km below the seafloor and beyond. This accounts for higher densities due to increased compaction with depth. The bottom of the total sediment sequence is defined as the top basement horizon. Interbedded basalts landwards of the COB are generally considered part of the sedimentary column if a deeper basement is imaged. However, given the diversity of data and their quality, it cannot be excluded that top basalt was erroneously interpreted as top basement. In this case, sub-basalt sedimentary rock sequences are missing in the thickness compilation (Funck et al. 2014).

The continental crystalline crust is divided by an intra-crustal horizon at a general depth of 20 km, dividing the layer into an upper and lower crust. In the shelf areas, the distribution of upper and lower crust goes over to a constant ratio to simulate a shear model. Densities of 2750 and 2950 kg m$^{-3}$ are assigned to the upper and lower continental crust, respectively, accounting for increasing densities with depth. The oceanic crust differs in composition and is modelled with a density of 2850 kg m$^{-3}$. The COB follows the interpretation of Funck et al. (2014), as shown in Figure 1d, and primarily follows

<table>
<thead>
<tr>
<th>Layer</th>
<th>Density (kg m$^{-3}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water column</td>
<td>2200</td>
</tr>
<tr>
<td>(replacement density*)</td>
<td>2200–2700</td>
</tr>
<tr>
<td>Sediments</td>
<td>2850</td>
</tr>
<tr>
<td>Oceanic crust</td>
<td></td>
</tr>
<tr>
<td>• upper</td>
<td>2750</td>
</tr>
<tr>
<td>• lower</td>
<td>2950</td>
</tr>
<tr>
<td>Lower crustal bodies</td>
<td>3100</td>
</tr>
<tr>
<td>Lithospheric mantle</td>
<td></td>
</tr>
<tr>
<td>• down to 700°C isotherm</td>
<td>3300</td>
</tr>
<tr>
<td>• down to 900°C isotherm</td>
<td>3260</td>
</tr>
<tr>
<td>Asthenospheric mantle</td>
<td>3200</td>
</tr>
</tbody>
</table>

*The replacement density for water refers to a sediment-filled ocean and is in accordance with the Bouguer anomaly correction. Densities for the lithospheric mantle are calculated after Sandwell (2001) with $\rho(T_l) = \rho_m[1 - \alpha(T_l - T_m)]$, where $T_l$ is the temperature at the thermal boundary layer, $T_m$ is the mantle temperature, $\rho_m$ is the mantle density (3200 kg m$^{-3}$) and $\alpha = 3.1 \times 10^{-5}$. $C^{-1}$ is the thermal expansion coefficient.
the landwards-most points of seismically defined oceanic crust. In addition, gravity and magnetic data were used for the compilation of the COB, and final adjustments were made using break-up reconstructions (Funck et al. 2014).

A new depth to Moho map for the study area, based on seismic refraction data and receiver functions, was compiled by Funck et al. (2014) (Fig. 2a). Seismic depth information is only available along the profiles and was then interpolated for regions without data coverage. The interpolation process and the seismic database are described in more detail in Funck et al. (2016).

In addition to the crust–mantle interface, regions displaying a high-velocity lower crust (HVLC) were mapped based on the seismic data. These regions occur in the COT zone, and velocities in the lower crust reach values of up to 7800 m s\(^{-1}\). The information on these zones is sparse and limited to the seismic lines. In the 3D density model, these areas are modelled as lower crustal bodies (LCB) with a higher than typical lower crustal density of 3100 kg m\(^{-3}\). The LCB density lies within the typical range of densities for this lower crustal feature (3000–3300 kg m\(^{-3}\); e.g. Mjelde et al. 2009; Kvarven et al. 2014; Nirrengarten et al. 2014).

Gravity data in the oceanic domain contain a thermal effect that is related to the deep lithosphere. The effect is expressed by a long-wavelength gravity field that originates from density changes related to lithospheric cooling with age. In the continental domain, where the lithosphere is thicker, this thermal effect is also present. This gravity field can be modelled and approximated by estimating the depths to isotherms and assigning matching densities. In the oceanic domain, the isotherm calculations are based on the age of the crust as presented in Gaina (2014) and Gaina et al. (this volume, in review). Here, the approach from Sandwell (2001 and references therein) was used for the calculation of the isotherms and densities. In the continental domain, the definition of the lithospheric thickness and the depth to the isotherms are based on Artemieva & Mooney (2001) and Artemieva (2006). As the isotherms for the entire study area are based on these two different approaches, a rather sharp transition from the oceanic to the continental domain occurs. This transition has been smoothed by forward modelling of the ultra-long-wavelength gravity field (here we used the IGMAS\(^+\) software). The density values for the lithospheric and asthenospheric mantle are relative. This means that the decrease in density with depth is caused by the thermal field, while the effect of pressure and composition, which would lead to an increase of densities with depth, is neglected. The use of absolute
pressure- and temperature-dependent densities does not affect the gravity field by more than 1%, and would require significantly more complicated computations. Compared to errors in the data and the regional extent of the model, it is acceptable to use the simpler approach and avoid the extensive calculations. The uppermost part of the lithospheric mantle has a density of 3300 kg m\(^{-3}\). The 700°C and 900°C isotherms were then used with densities of 3260 and 3240 kg m\(^{-3}\), respectively. The 1300°C isotherm is regarded as the base of the lithosphere, and a density of 3200 kg m\(^{-3}\) was assigned to the asthenospheric mantle underneath.

Inversion

After the initial model set-up and preceding forward modelling, a structural gravity inversion was calculated to verify the Moho depths derived from seismic interpretations and to obtain estimates on the Moho depth in regions with no seismic coverage. The software package GM-SYS 3D, which allows for a structural inversion with the aid of a constraining grid, was used. In order to limit the changes applied to the initial Moho interface, a constraining grid that is based on the confidence in estimates of applied to the initial Moho interface, a constraining grid, was used. In order to limit the changes for a structural inversion with the aid of a constraining grid in all areas with no observed LCB, preventing upwards movement of the Moho. In oceanic areas, the intra-crustal layer boundary is often coincident or very close to the grids defining the Moho and the top of the LCB. Along the mid-Atlantic ridges, downwards movement of the Moho grid is limited by the isotherms that were added to model the thermal effect of the lithospheric mantle. In order to bypass these limitations, the named grids were removed from the model geometry. This simplifies the density structure in a way that the model has too much density in the mantle and too little in the crust. In turn, the calculated gravity will be affected and the measured Bouger gravity needs to be adjusted to remain comparable. The gravity effects from the residual densities were calculated and the measured gravity corrected accordingly. With this, the inversion can be run with only the main model horizons and the adjusted gravity data to which a 60 km low-pass filter was applied that removes small wavelength signals that are not considered to originate from the Moho.

The inversion reached its convergence criterion within 15 iterations, providing a gravity misfit of 16 mGal standard deviation. Owing to the 10 km resolution of the Moho, in combination with the constraining grid based on the seismic uncertainties, the inverted Moho interface contains some small wavelength features that are considered as artefacts. In order to remove these features, the resulting Moho grid was smoothed with a 100 km-wavelength low-pass Butterworth filter. Finally, the removed layers are put back into the model and the final fit to the measured gravity is calculated.

Results and discussion

The final fit between measured Bouguer anomaly (Fig. 1c) and calculated anomaly from the 3D model has a standard deviation of 22 mGal. The residual gravity map is shown in Figure 3a. The misfit is acceptable for this regional model, where many local intra-crustal sources had to be neglected. An additional source of error and distortion in the model are inaccuracies in the cover sequence, as the compilation includes a significant amount of high-density volcanic rocks (e.g. on the Norwegian margin or around the Faroe Islands) that are not accounted for in the density distribution. Considering the differences between measured and calculated anomalies, the remaining residuals are mainly of short wavelengths, which usually are not associated with the crust–mantle interface. In some areas, large misfits coincide with the tight constraints on the initial data. Here, further adjustment of the model was prevented. This is especially apparent to the south of Iceland along the Reykjanes Ridge, in the Hatton Basin, along the Kolbeinsey Ridge and in the vicinity of the Jan Mayen microcontinent. Also noticeable are the jumps in residuals along the COB: for example, along the Hatton High, the Faroe Platform, the Vøring Basin and the Kangerlussuaq Basin, to name just a few. The density distribution at the COT is more complex than implemented in our model and also contains volcanic sequences, which partly explains the misfit. The correlation of residuals and constraints may suggest that the seismic constraints for Moho and COB need revisiting in some areas.

The residuals in the centre of the study area (Fig. 3a) correlate well with an upper-mantle seismic velocity anomaly that runs from south of Iceland to the Norwegian coast (Bijwaard & Spakman 1999; Weidle & Maupin 2008). The residuals
become smaller towards Norway but this correlation suggests that the definition of the lithospheric mantle might be too simple. The issue could be followed up by combining and improving the model with satellite-derived gravity gradients that are more sensitive to density anomalies at larger depths (Bouman et al. 2015) and by focusing more on the mantle composition, as done by Afonso et al. (2007). However, the focus of this work was not on a detailed modelling of the mantle but on the crustal structure. The central residuals also correlate with the Greenland–Iceland–Faroe Ridge complex and will be further addressed in the following discussion.

**Depth to Moho**

The final depth to Moho after gravity inversion is given in Figure 4. Compared to the initial Moho depth (Fig. 2a), the new map provides more detail in areas that lack seismic coverage and previously only showed smooth interpolations. Along the seismic lines, the gravity inversion indicates similar Moho depths as in the initial map. This is related to the fact that changes are limited to the assigned uncertainty in the seismic data (Fig. 2b). However, structural features between the seismic lines become better defined compared to the initial interpolations and a few examples are listed here. The northwards prolongation of the Rockall Basin is clearer, as are the outlines of the Hatton and Møre basins. Off SE Greenland, the Ammassalik Basin (Gerlings et al. this volume, in review) is characterized by a shallowing of the Moho that was not seen previously. The NE part of the Norway Basin is now more clearly outlined; the gravity inversion indicates that the shallow Moho extends up to the COB. Very distinct improvements are seen in the Danmarkshavn and Thetis basins. Seismic data in this area are sparse due to the ice cover. However, recently released seismic reflection data allowed for a significantly improved sediment thickness map (Geissler et al., this volume, in review) covering this area. These data are included in our 3D model, and allowed for an improved and more detailed image of the Moho depth (Fig. 4) when compared to the rather smooth Moho map based on extrapolation of the seismic refraction data (Fig. 2a). Moho depths beneath the NE Greenland Shelf vary between 15 and 25 km, and are consistent with the Norwegian margin on the conjugate side.

![Fig. 3.](image-url)
The crustal thickness is defined as the thickness of the crystalline crust from top basement to the Moho (Fig. 5). Underneath the sedimentary basins, crust is usually thin and major basins, such as the Rockall, Hatton, Møre and North Danmarkshavn basins, are characterized by a crustal thickness of around 10 km, but can locally be less than 6 km (Møre and Rockall basins). The igneous complex comprising the Greenland–Iceland–Faroe Ridge is characterized by a crustal thickness of around...
30 km. Similar values are observed beneath the Rockall and Porcupine highs. The central part of the study area overlaps with previous studies incorporating gravity data (e.g. Greenhalgh & Kusznir 2007; Torsvik et al. 2015). The crustal thickness of the model here is slightly thicker than in these studies. However, the results still produce gravity residuals here and, therefore, hold a degree of uncertainty. Unfortunately, neither of the previous two studies shows how well the models fit the data.

Fig. 5. Crustal thickness based on the structural gravity inversion. The white lines show the interpretation of the COB after Funck et al. (2014) and the black lines mark the locations of the 2D cross-sections shown in Figure 6. DB, Danmarkshavn Basin; GIFR, Greenland–Iceland–Faroe Ridge complex; GFZ, Greenland Fracture Zone; JMFZ, Jan Mayen Fracture Zone; SFZ, Senja Fracture Zone; TB, Thetis Basin.
Larger discrepancies occur at the northern part of the Jan Mayen microcontinent, where the crustal thickness is, in most parts, shallower in comparison with that reported in Torsvik et al. (2015). This part, however, is well constrained by seismic data (Fig. 2) and the results here agree, for example, with Kodaira et al. (1998). Along the Reykjanes Ridge, the results are strongly influenced by the seismic constraints, as mentioned above (Fig. 3a). Similarly, the seismic constraints are responsible for the thinner crust around the Faroe and British islands compared to that discussed in Torsvik et al. (2015).

2D cross-sections

In Figure 6, three vertical Trans-Atlantic cross-sections through the final lithospheric model are shown. The location of the cross-sections is shown in Figures 4 and 5. The profiles show a generally good fit between observed and calculated gravity, but major residuals are present: (1) around most of the COBs; (2) in areas where volcanics are being neglected in the sediment sequence; (3) close to onshore areas; and (4) partly where seismic information is sparse or does not allow for further adjustments within the seismic error bounds. These four cases of increased misfit will now be examined in more detail.

(1) In the density model, the COB of Funck et al. (2014) is used to distinguish between oceanic and continental upper crust, as well as oceanic and continental lithospheric mantle. In recent years, alternative interpretations of the COB have been published (see Funck et al. 2014 for a summary) differing by up to approximately 80 km off East Greenland (Fig. 1d). A number of the published COBs cross-cut gravity anomalies on the tilt-derivative map (Fig. 1d), which is unusual as the tilt derivatives generally correlate with crustal structures. However, the COB on glaciated margins is complicated by the thick wedge of glacial sediments, the strong lateral density contrasts of which might interfere with the crustal gravity signal. Uncertainties in the location of the COB are reflected in the 2D sections. Even though we implemented a slightly inclined COB (due to the size of the grid cells and minimum-curvature gridding), the changes in crustal densities at this boundary are sharp. In reality, the setting is much more complex and sharp changes are unlikely. Furthermore, the density distribution of the glacial sediments may not be well accounted for. These simplifications are then reflected in the results.

(2) The model uses the total sediment thickness for the estimation of the top basement depth. This sequence, as described earlier, also contains basalts and volcanic rocks, the higher densities of which are now missing in the model. Depending on the volume, this effect can be of the order of tens of mGal. This most probably explains part of the errors in the eastern portions of the lines (e.g. parts of the Lofoten, More and Hatton basins). No volcanic rocks were mapped in the Danmarkshavn Basin (Geissler et al., this volume, in review) and the mass deficit here might, instead, be explained by too low sediment densities in this very deep (c. 18 km) basin (Line 1).

(3) The onshore parts of Norway and Greenland mark the borders of the study area, and discrepancies here are to be expected. Owing to the lack of seismic information, these areas were only roughly modelled to reduce edge effects. Detailed modelling of the intra-crustal geometry and even of upper-mantle sources (due to varying composition), as in more dedicated modelling studies, was not carried out. Examples for such studies (3D and 2D) can be found for Scandinavia (Ebbing et al. 2012; Gradmann et al. 2014) and Greenland (Medvedev et al. 2013; Schiffer et al. 2015).

(4) The western part of Line 1 crosses the COB and the Danmarkshavn and Thetis basins. This is an area where the initial Moho depths were interpolated based on very sparse seismic refraction coverage. Owing to the absence of detailed seismic information and the probability that local, unmodelled sources are present (e.g. high-density LCB as mapped on the conjugate Norwegian side: Funck et al. 2014), the gravity does not match perfectly.

Implications from density inversion

Part of the shorter wavelength gravity residuals (Fig. 3a) can be attributed to variations in crustal density. A density inversion, allowing for laterally varying densities in the upper crustal layer, confirms

Fig. 6. Three cross-sections through the final 3D lithospheric model. (Top panels) Observed and calculated gravity. (Middle panels) Model cross-section over the entire model depth. (Bottom panels) Enlargement of the cross-sections showing only the top 40 km. Locations of the sections are indicated by black lines in Figures 4 and 5. The blue arrows indicate the locations of the COB. Densities are listed in Table 1; the upper-crustal density is given in Figure 3b. DB, Danmarkshavn Basin; TB, Thetis Basin.
this (Fig. 3b). The initial densities of 2850 and 2750 kg m\(^{-3}\) for oceanic and upper crust, respectively, were allowed to vary between 2700 and 2950 kg m\(^{-3}\).

The resulting density distribution (Fig. 3b) is close to the initial values in most parts of the study area. Larger changes are mainly observed in the oceanic domain, where large misfits after structural inversion occurred due to (1) the seismic constraints and (2) along the Greenland–Iceland–Faroe Ridge complex. Along this ridge, the velocity of the lower oceanic crust deviates from standard values and exceeds 7 km s\(^{-1}\) (Richardson et al. 1998; Holbrook et al. 2001), which consequently implies higher densities than the 2850 kg m\(^{-3}\) used here. The increased crustal densities suggested by the inversion can be attributed to the lower-crustal mass deficit. Trials have shown that an increase in lower-crustal densities by up to 100 kg m\(^{-3}\) provides a noticeable improvement in the gravity fit. Integration of such density changes in the model requires more detailed modelling of the respective area than done in this study. Furthermore, uncertainties in the COB are reflected in the density results, in particular by increased densities around and south of the Faroe Islands and at the conjugate side offshore SE Greenland. The previously discussed neglect of volcanic rocks in the sediment cover also contributes to density changes in the upper crust where the inversion tries to compensate for the mass deficit.

Conclusions

Seismic interpretations and a global gravity model were used to construct a regional 3D density model for the NE Atlantic. While the initial geometry of the density model is defined by the seismic information, the gravity data helped to verify and improve Moho depth estimations, especially in areas of sparse seismic coverage. The Moho depth model enhances local details based on a reasonable model set-up. Owing to the regional character of the model, intra-crustal sources (LCB) are only modelled to a certain extent and distortions caused by neglected basalts in the sediment cover are not taken into account. Considering the ambiguity of potential field data, the uncertainties and applied simplifications, many possible interpretations for the North Atlantic crustal structure exist. The model here presents one of the possible solutions, the one that fits the input constraints best. It is well suited as a starting point for further, more local and detailed studies.

The crustal thickness model outlines the main structural elements of the North Atlantic region. The crustal model is well constrained and some of the remaining gravity residuals probably originate from deeper sources. The model is therefore an important contribution to identify the reason for observed upper-mantle velocity anomalies (Bijwaard & Spakman 1999; Weidle & Maupin 2008), and to understand the evolution of the North Atlantic topography and its surroundings. Furthermore, the inverse- and forward-modelling procedure can be applied to test the reliability of other seismically derived crustal thickness models (e.g. Artemieva & Thybo 2013).

The modelling results, together with the tilt-derivative gravity map, can be used to better understand the location of the COB. The residuals in the gravity (Fig. 3a) and results from the crustal density inversion (Fig. 3b), as well as a more detailed look along 2D sections (Fig. 5), show the additional value of gravity modelling. However, for a full discussion and better resolution of the boundary, higher-resolution modelling is required. In this context, the integration of magnetic data is essential.

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