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Nature of the crust-mantle transition zone and the thermal state of the upper mantle beneath Iceland from gravity modelling

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SUMMARY

Gravity data from Iceland and its surroundings are analysed and modelled with respect to seismic data. A Bouguer gravity map of Iceland is recomputed based on admittance between the topography and the gravity and with corrections for glacial ice sheets. From seismic data and with the help of relations between the residual topography and the depth to seismic boundaries we construct maps of the main seismic boundaries, including the Moho. By inversion calculations we recomputed these maps, assuming different density values for Seismic Layer 4 to fit the observed gravity field. We found that the average density of Layer 4 has to be in the range 3050–3150 kg m^{-3} in order to fit both seismic and gravity data. Thus we conclude that Layer 4 is a transition zone between the mantle and the oceanic crust in Iceland. Furthermore by assuming that the upper-mantle density variations necessary to compensate for the gravity effect of crustal layers, are caused by thermal variations in the upper mantle, we calculate the depth to the 1200 °C isotherm to be at 30–50 km depth below Iceland but rising up to less than 20 km below parts of the volcanic zone in Northern Iceland. We conclude that the temperature within the Seismic Layer 4 is close to 600 °C at its top, increasing to approximately 950 °C at its bottom (Moho), which makes a widespread layer of partially molten material within Layer 4 unlikely. By use of cross spectral analysis of the gravity field and the external topographic load at short wavelengths, we conclude that the elastic plate thickness in Iceland can hardly exceed 6 km. In addition we point out that the residual isostatic anomalies have circular forms east of the eastern volcanic zone but are near parallel to the ridge axis on the western side. This form of the anomalies may be caused by pressure from the eastward moving mantle plume below the eastern volcanic zone.

Key words: crust, gravity, mantle, seismic boundaries, temperature.

1 INTRODUCTION

Iceland is a unique geological and geophysical phenomenon, located where the active Mid-Atlantic spreading ridge is superimposed on a hotspot. Thus the tectonic and physical parameters of the crust and the underlying mantle are controlled by interaction of these two processes: ocean floor spreading and an upwelling mantle plume. The character of the ocean floor spreading is reflected in the typical age-distance and heat flow relationships which fit well with normal spreading models (Pálmason 1973, 1986). The hotspot character is shown, for example, by the fact that Iceland and its surroundings comprise a huge topographic anomaly on the ocean floor forming the transverse Iceland–Faeroe–Greenland Rise (e.g. White 1997).

Since the early sixties, the nature and properties of the crust and upper mantle beneath Iceland have been explored and discussed intensively. Different authors have come to quite different conclusions, especially regarding the thickness and the thermal state of the crust. In principle two different models of the Icelandic crust have been put forward, the thin and hot crust model (Pálmason 1971; Gebrande *et al.* 1980; Beblo & Björnsson 1978, 1980; Flóvenz 1992, 1993) and the thick crust model (Båth 1960; Zverev *et al.* 1976; Pavlenkova & Zverev 1981; Bjarnason *et al.* 1993; Menke & Levin 1994; White *et al.* 1996; Darbyshire *et al.* 1998, 2000a,b; Weir *et al.* 2001). These models are based on different interpretations of seismic data, with or without the aid of magnetotelluric, heat flow and gravity data.

Existing seismic velocity models of the Icelandic lithosphere are similar but their interpretation is quite different. The upper crust, which is mainly made of sequences of flood basalts and hyaloclastites, is very inhomogeneous. Near surface *P*-wave velocities vary from less than 2.0 km s^{-1} in fresh new lava flows on the surface to more than 5.0 km s^{-1} in dense and altered flood basalts (Pálmason 1971; Flóvenz 1980; Flóvenz & Gunnarsson 1991). In general, the velocity increases more or less continuously with depth with a gradient of 0.57 s^{-1} until it reaches a value of about 6.5 km s^{-1}

at depths of 3–10 km. The 6.5 km s^{-1} isovelocity surface is usually defined as the top of Oceanic Layer 3. It appears to have almost constant velocity or slightly positive velocity gradient. At a depth range of 10–35 km velocities of 7.0–7.4 km s^{-1} are normally reported (Båth 1960; Pálmason 1971; Bjarnason *et al.* 1993; White *et al.* 1996; Staples *et al.* 1997; Brandsdóttir *et al.* 1997; Darbyshire *et al.* 1998), though the values in the lower part are likely to be uncertain due to insufficient length of the profiles. Pálmason (1971) refers to this region as Layer 4 and we keep his terminology here.

Based on the velocity structure, the nature of Layer 4 has been explained in two different ways; either it consists of hot and partially molten mantle material (Pálmason 1971; Gebrande *et al.* 1980) or subsolidus crustal material (Zverev *et al.* 1976; Pavlenkova & Zverev 1981; Menke & Levin 1994). The first explanation corresponds to a thin crust model while the latter can be referred to as the thick crust model. The thin crust model was supported by the presence of high electrical conductivity at Layer 4 depth, observed in magnetotelluric soundings (Beblo & Björnsson 1978, 1980; Hersir *et al.* 1984; Eysteinsson & Hermance 1985) and high heat flow (Pálmason 1973; Flóvenz & Sæmundsson 1993). Pálmason (1971) estimated on the basis of heat flow data that the temperature at the upper boundary of Layer 4 is close to the melting range of basalts beneath SW-Iceland, but 100–200 °C lower beneath N- and SE-Iceland.

Båth (1960) and Bjarnason *et al.* (1993) discovered strong wide angle reflections from 20–30 km depth in long range refraction surveys in SW Iceland. Bjarnason *et al.* (1993) interpreted these as Moho with sub-Moho velocities of 7.7 km s^{-1}. The presence of this reflector has later been confirmed at 15–40 km depth elsewhere below Iceland (White *et al.* 1996; Menke *et al.* 1996; Darbyshire *et al.* 1998; Weir *et al.* 2001).

Receiver functions have also been used to estimate the deep reflector below Iceland (Du & Foulger 1999, 2001; Schlindwein 2001 pers. comm.). In general the receiver functions are sensitive to large velocity contrasts as could be expected at Moho but due to the heterogeneity of the crustal material in Iceland signals from Moho are difficult to detect. However by careful analysis of the data the authors are able to determine the proposed Moho at several stations. The results show, in general some scattering of the data and probably slightly more depths than could be expected from the refraction data.

Flóvenz (1992) used the laboratory results of Sato *et al.* (1989) and Takahashi & Kushiro (1983) to conclude that in the thin crust model the temperature in Layer 4 has to be close to 1200 °C and that it might contain 1–4 per cent partial melt. This is based on the assumption that Layer 4 is made of mantle peridotite with *P*-wave velocity of 7.2 km s^{-1}. A consequence of this is that there could be favourable conditions for the formation of large volumes of molten crustal material at the interface between peridotitic Layer 4 and gabbroic Layer 3. He furthermore showed that this model is in good agreement with magnetotelluric and heat flow data. In this case the reflector at 20–35 km depth could be interpreted as the bottom of a partially molten zone in the topmost mantle.

Menke & Levin (1994) and Menke *et al.* (1995) have shown that the path-averaged quality factor for waves turning in the depth range of 12–20 km is typically 200–300 for *P* waves and 400–600 for *S* waves, the lowest values being 50 per cent lower. Based on 1-D layered inversion for the quality factor and assuming gabbroic lithology they conclude that the upper limit for temperature in this depth range (Layer 4) is low: 700–775 °C, thus without melt. These results are in disagreement with the temperature estimations of White & McKenzie (1989), White *et al.* (1995) and White (1997) who argued that the temperature near the bottom of Layer 4 is even 50–150 °C greater than under the normal mid-ocean ridge.

Thus, a key problem for understanding the nature and origin of the Icelandic lithosphere is to determine the thermal state and composition of Layer 4. The existing hypotheses about the composition and the thermal state of this layer are in obvious contradiction to each other and no single model, consistent with the results of all geophysical data, has yet been created. The thin crust model has difficulties to explain the high Q factors and the presence of relatively deep earthquakes (Einarssson 1989; Stefánsson *et al.* 1993; Flóvenz & Sæmundsson 1993) while the thick crust model has difficulties explaining the high surface heat flow and the low resistivity at Layer 4 depths. Seismic data hardly help to solve this problem. First, there are almost no data from refraction profiles for the depth interval 25–40 km, while reflection and receiver function studies do not provide reliable velocities. Due to obvious interference between crustal and mantle structures a vertical resolution of the tomography data based on teleseismic studies is also insufficient to discriminate between the lower crust and the uppermost mantle. Second, seismic velocities only do not provide direct indications for a composition of the material composing the Layer 4. Seismic properties of mafic and ultramafic rocks could be completely different at room conditions but the influence of temperature could make them more similar (Sato *et al.* 1989). Another factor, that has been largely ignored, is the presence of water which significantly reduces seismic wave velocities through anelastic relaxation (Karato & Jung 1998) but hardly affects densities of mantle material. Thus, due to the water content it is possible to reduce velocities in the mantle material even without significant melting. Gravity data might make a significant contribution to the solution of this problem since the densities of crustal material like gabbro differ very significantly from the density of mantle peridotites or dunites.

The seismic properties below Layer 4 to depths of several hundred kilometres have been studied by several authors, mainly based on teleseismic data (Tryggvason 1962, 1964; Long & Mitchell 1970; Tryggvason *et al.* 1983; Bjarnason *et al.* 1996; Shen *et al.* 1996; Wolfe *et al.* 1997). The results show a low velocity anomaly in *P* and *S* waves that is an expression of the Iceland mantle plume. According to Wolfe *et al.* (1997), the mantle plume anomaly is a cylindrical zone that extends from 100 km depth to at least 1000 km with a radius of 150 km, with the centre beneath the eastern volcanic zone in central Iceland. The magnitudes of the velocity anomalies are 2 per cent and 4 per cent for *P*- and *S*-waves, respectively.

The first Bouguer gravity map of Iceland was published by Einarsson (1954) who reported a gravity minimum of amplitude 75 mGal relative to the coastal areas. By applying the Airy–Heiskanen isostatic reduction he showed that the gravity bowl can be explained on the basis of a thin sialic crust of thickness 20 km or even less. In 1980 a new Bouguer gravity map of Iceland and its surrounding shelf was published by Thorbergsson *et al.* (1990) and a still new version including satellite data by Eysteinsson and Gunnarsson in 1995. There are several papers dealing with 2-D gravity modelling along seismic profiles in Iceland and surrounding area (e.g. Staples *et al.* 1997; Richardson *et al.* 1998). Darbyshire *et al.* (2000b) extrapolated the 2-D gravity models from FIRE and ICEMELT profiles for the whole Iceland. A significant progress in understanding the density structure of the crust and upper mantle in Iceland is achieved in the above papers, though several important problems are still unresolved.

First, there exists a significant trade-off between the lower crust and upper-mantle densities due to an obvious non-uniqueness of the

inverse gravity problem solution. The relationship between different models which equally fit the observed gravity has not been sufficiently analysed. Second, the obtained gravity model which fits the observed gravity may be not stable from a geodynamic point of view, in isostatic equilibrium in the simplest case. Third, the relation between crustal density structure and thermal state of the lithosphere needs to be studied in more detail. Large scale crustal structures (about 50 km or more) must be compensated by mantle density anomalies of the opposite sign. These anomalies are likely to represent variations in the thermal state of the upper mantle. Thus, by modelling the density structure of crustal and subcrustal layers it is possible to estimate temperature in the upper mantle at still greater depths. Fourth, it is necessary to check if 2-D models are accurate enough in the case of Iceland. Darbyshire *et al.* (2000b) started to analyse this problem by considering the effect of near surface bodies (volcanoes) and concluded that it is insignificant. However, a 3-D effect of deep non-linear structures like the Icelandic mantle plume could be much more important. We analyse these problems in the present study.

Another way to estimate the thermal state of the upper mantle through analysis of gravity data is to determine the flexural rigidity (effective elastic plate thickness) of the Icelandic lithosphere. The effective elastic thickness (T_e) of the oceanic lithosphere depends on its thermal state. We are going to use the well-known cross-spectral analysis of the gravity data and the external lithosphere load (Dorman & Lewis 1970) to determine the effective elastic thickness of the Icelandic lithosphere.

This paper deals with 3-D modelling of the available gravity data of Iceland and its surroundings, constrained by the seismic velocity structure, to try to differentiate between thin and thick crust models and to determine density and thermal state of the lower crust and upper mantle.

2 STRATEGY OF THE GRAVITY MODELLING

The main idea of the method used in this paper is to make several different initial models of density distribution in the crust and the upper mantle, based on topographic data and seismic data interpretation and to modify these models to fit the observed gravity field. Then the geological and geophysical implications of the resulting models are discussed and preference given to the most likely model.

The modelling process is done in the following steps:

(i) Creation of an initial structural model from seismic and topographic data, including maps of the top and bottom of seismic Layers 3 and 4.

(ii) Different density values of Layer 4 give different models of the Icelandic lithosphere. Minimum density would correspond to the hypothesis of a thick and relatively cold gabbroic crust under Iceland; maximum value corresponds to an anomalous mantle in Layer 4; and intermediate values relate to more complex cases. Density estimates for layers 1–3 are obtained from different sources.

(iii) Determination of the effective elastic thickness of the lithosphere (flexural rigidity) by analysing its gravity response to small scale external load. Comparison of the obtained value with the values for 'normal' mid-oceanic ridge crest provides a possibility to compare their thermal regimes.

(iv) Relatively large scale features of the lithosphere models (how large depends on the effective elastic thickness) should be compensated locally by mantle density anomalies. We expect that these anomalies are related to a temperature profile in the upper mantle. Thus different density models of the crust correspond to different temperature profiles in the upper mantle.

(v) Calculation of the gravity field for various lithosphere-asthenosphere models. Differences between the observed and the calculated gravity fields give the residual isostatic anomalies. These anomalies are indicators of the reliability of the individual density models.

(vi) Adjustment of main density boundaries (or densities) by application of inversion technique to minimize the residual isostatic anomalies.

(vii) Analysis of the obtained results.

3 THE INITIAL GRAVITY DATA PROCESSING

Eysteinsson & Gunnarsson (1995) have presented the most complete Bouguer gravity map which is based on approximately 5400 measurements of gravity values in Iceland and the surrounding sea area (Thorbergsson *et al.* 1990) and satellite data offshore. These data are corrected for the ice thickness on glaciers using data of Björnsson (1988, pers. comm. 1993), a terrain correction has also been applied beforehand. The resulting Bouguer gravity map is shown in Fig. 1. The density used to compute the Bouguer anomalies on land was 2520 kg m^{-3} (see below).

The Icelandic relief consists of structures with different densities: water, ice, topography. For gravity modelling we need to operate with homogeneous lithosphere loads. We use a so-called 'adjusted' topography, obtained after numerical densification of water, ice sheets and Icelandic relief to the density 2670 kg m^{-3}. Thus, the adjusted topography represents the homogeneous external lithospheric load and may be used together with Bouguer anomalies, for example, for admittance calculations.

When applying the Bouguer correction Eysteinsson & Gunnarsson (1995) used a density value of 2600 kg m^{-3}. This value is important for the gravity modelling that follows. Therefore we checked this value using the admittance technique (e.g. Dorman & Lewis 1970; McKenzie & Bowin 1976). The transfer function (admittance) represents a relation between Bouguer gravity anomalies and topography (external load) in the Fourier domain. At short wavelengths ($<$50 km) the admittance depends only on the average value of the topographic density, because the influence of isostatic compensation at these wavelengths is negligible. If the topographic density value used for calculation of the Bouguer anomalies is correct, the transfer function at wavelengths less than 50 km should be close to zero.

The admittance computed from the Bouguer gravity anomalies (2600 kg m^{-3}) and the adjusted topography is shown in Fig. 2. Despite some scattering all the values are negative which means that true topographic density must be on the average less than the one used. The average position of the admittance values gives the most appropriate value of the mean crustal rock density: 2520 $\pm$ 15 kg m^{-3}. We used this density to re-compute the Bouguer gravity anomaly onshore. Offshore, a standard value of 2670 kg m^{-3} was used. The topography and gravity are interpolated on a regular grid with a resolution of 2.5′ $\times$ 6′ (4.63 $\times$ 4.7 km) with the frames 27°–12°W and 62.5°–67.5°N.

4 MODELS OF THE LITHOSPHERE

For our gravity modelling, reliable maps of seismic velocities are necessary. These include a map of the depth to the top of the Layer 3,

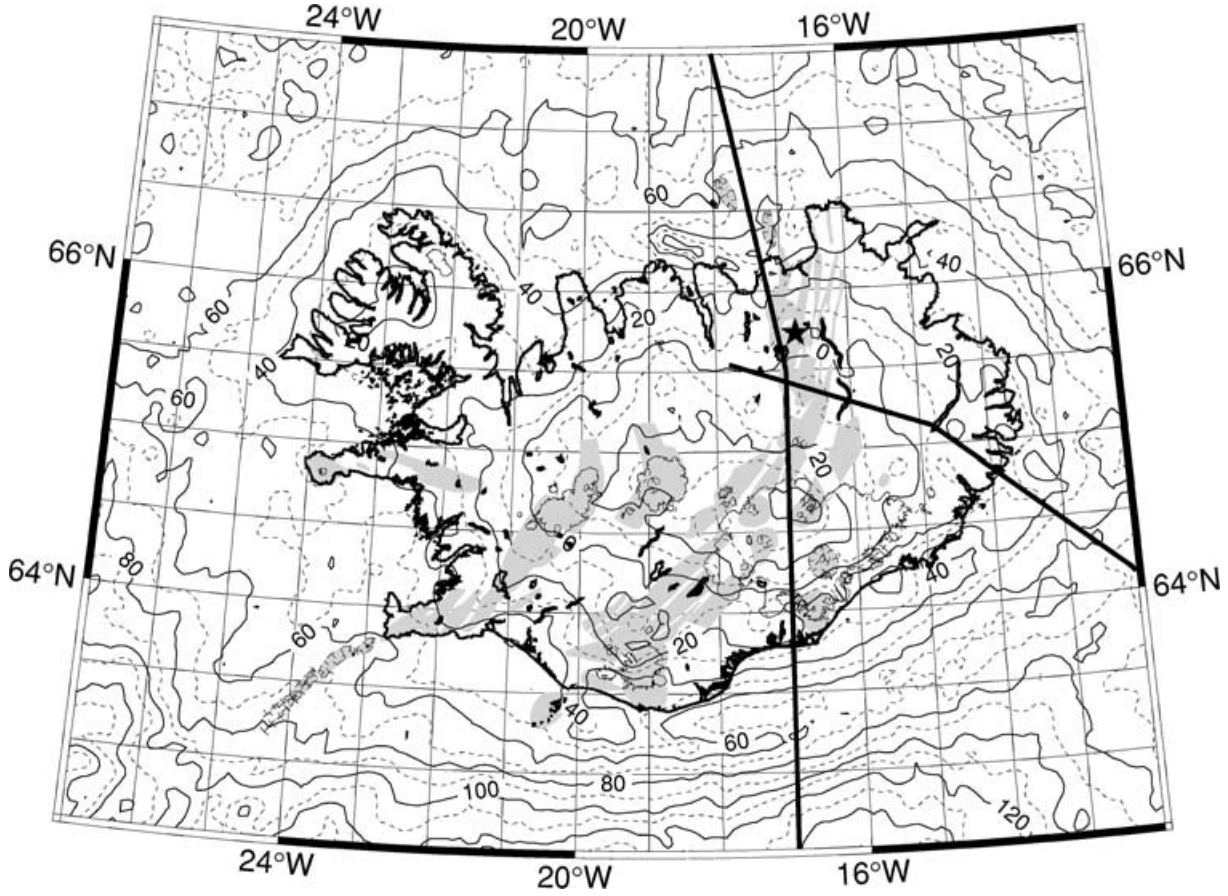

Figure 1. The Bouguer gravity map of Iceland. Reduction density is 2520 kg m^{-3} onshore and 2670 kg m^{-3} offshore. The location of the profiles discussed in chapter 6 is also shown and the location of Krafla is marked with a star.

defined as the 6.5 km s^{-1} isovelocity surface, and a map of the depth to the top of Layer 4. Such maps were published by Pálmason in 1971 but they can be improved considerably by adding results from successive surveys and by defining the top of Layer 4 as the 7.0 km s^{-1} isovelocity surface (Flóvenz 1980; Gebrande *et al.* 1980; Bjarnason *et al.* 1993; White *et al.* 1996; Darbyshire *et al.* 1998). The bottom of Layer 4 is defined by the strong wide-angle reflections observed in long range refraction surveys (Båth 1960; Bjarnason *et al.* 1993, 1996; White *et al.* 1996; Staples *et al.* 1997; Darbyshire *et al.* 1998, 2000a). The data coverage is, however, not sufficient to create a map by interpolation only. Darbyshire *et al.* (2000b) extrapolated the results obtained on the FIRE, ICEMELT and some other profiles using gravity data. This approach is not acceptable to our study because we are going to use the gravity data together with the seismic boundaries for density modelling. Thus, for preparing our Moho map we will only use data that are not based on the gravity field.

To produce the initial map of this boundary we are first going to find a relation between topography and the observed depth to the boundary where seismic profiles exist and then to use this relationship for the areas without seismic data. Similar approach is commonly used in the geophysical literature. For example, ocean topography is used to put some limitations on the well-known cooling lithosphere model, and the Airy model of isostatic compensation is sometimes used to determine the position of the Moho in mountainous areas of the continents. However, we cannot assume such a general dependence for the whole study area because it contains different tectonic units where different relations are likely to exist between topography and the relevant boundaries. Menke (1999) found an unusual relationship between the depth to the bottom of Layer 4 and topography which is essentially different from the normal continental relationships and from the one for the Mid-Atlantic Ridge near Iceland (White *et al.* 1995). Two main trends exist for the relationship between depths to Moho and topography in the study area. The 'normal' ocean is characterized by a near constant crustal thickness i.e. topographic highs are accompanied by a decrease of Moho depth. In Iceland, on the contrary, topographic highs are mostly connected with crustal thickening. These two trends result in a non-uniform relationship between the observed topography and the depth to Moho for the whole study area.

Thus, the first step in the construction of an initial lithosphere model is a separation of the superimposed structures. This is done by the construction of a model of 'normal' oceanic lithosphere and by regarding Iceland and the Rises as 'disturbances' to it. Then we make a map of the residual topography by subtracting the normal oceanic topography from the true topography. This is somewhat similar approach to that of White *et al.* (1995) and White (1997) who computed the residual topography by removing the topography of 'normal' ocean model. We expect to find a close relationship between this parameter and the residual Moho obtained in the same way by subtracting the 'normal' ocean Moho from the observed values and then to use this relationship for areas where seismic data are missing.

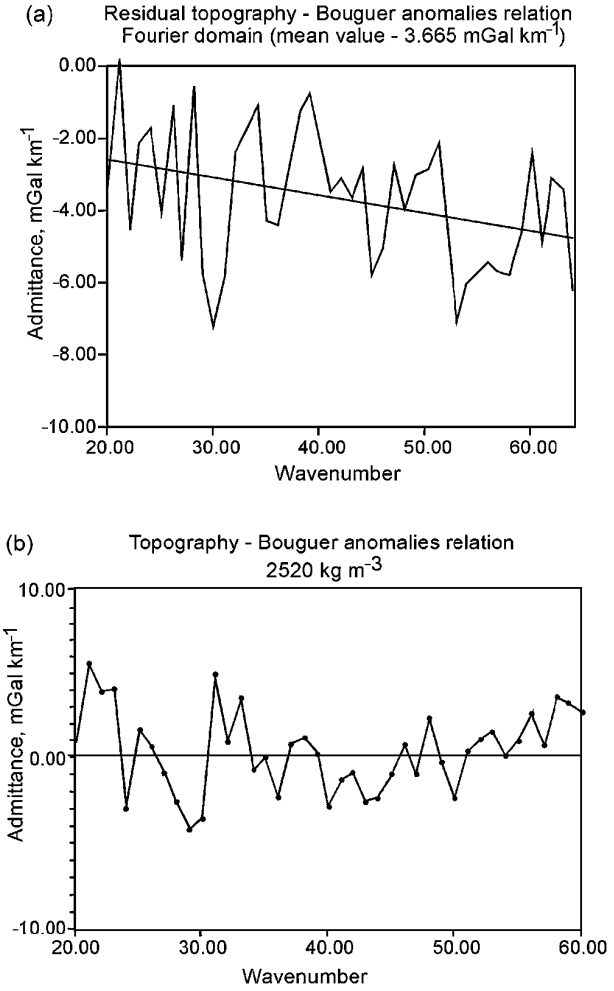

Figure 2. Admittance values estimated from (a) Bouguer anomalies (2600 kg m^{-3}) and (b) topography. The average position of the admittance values gives the most appropriate value of the mean crustal rock density, 2520 ± 15 kg m^{-3}.

4.1 'Normal' ocean lithosphere model

The properties of a 'normal' ocean lithosphere are well defined and depend only on the distance from the spreading axis, assuming constant spreading rate. Usually a square root relationship which corresponds to the cooling lithosphere model is used (McKenzie 1967; Davis & Lister 1974). We can determine this relationship for topography of the Mid-Atlantic ridge outside the Icelandic block and then use this relationship to remove the 'normal' ocean floor topography from the real topography in the whole area. The residual topography will then be caused by anomalous structures and processes in the Icelandic block. After the removal of the normal oceanic topography we can develop relationships between the seismic boundaries by defining top and bottom of Layer 4 and the 'residual' topography and use them to construct the initial maps of the seismic boundaries defining Layer 4 in the areas without seismic data.

We have determined an empirical relationship between the distance from the spreading axis and the depth for the Reykjanes Ridge to the south from Iceland:

$$d = 0.6 + 0.086\sqrt{x}\,(\mathrm{km}) \qquad (1)$$

where d is the depth and x is the distance from the spreading axis. The first term of this relation corresponding to the ocean depth at zero age is about 2 km less than that usually attributed to the 'normal' ocean floor. This difference is caused by a high in oceanic topography (also in gravity and geoid) with the centre in the northern part of the Atlantic Ocean. The depth of the ocean within this area fits a square root relation but the average level is shifted upward and depends on the distance from Iceland (e.g. White *et al.* 1995). The nature of this maximum is still under discussion. Probably its sources are located at depths of the order of 200 km or greater. It may also be caused by a global mantle flow (e.g. Lister 1982). Since we use this relationship only for the relatively narrow area of Greenland–Iceland–Faeroe Rise, while the total size of the North Atlantic gravity and topography maximum is several thousand km, the average level of normal topography is considered as a constant value. It is also worth noting that we use this relation only for a rough separation of different structures but not for precise modelling, thus it does not affect the final result notably.

Using this relationship and the position of the main spreading axes shown in Fig. 3 we estimated a 'normal' oceanic topography. The spreading axes system includes the Reykjanes and the Kolbeinsey ridges south and north of Iceland, their connections through two near-parallel spreading axes in Iceland, and an extinct spreading axis segment in W-Iceland (Saemundsson 1979; Jóhannesson 1980). By removing the effect of the normal ocean topography according to eq. (1) from the adjusted topography we obtain the residual topography shown in Fig. 3. A low pass filter is used to remove the small scale residual topography features ($L <$ 70 km) which may be supported by stresses within the lithosphere and are not reflected in Moho variations.

4.2 Main lithosphere boundaries

Standard seismic layered model for the North Atlantic crust is shown in Fig. 4 (e.g. Müller & Smith 1993). The residual component of

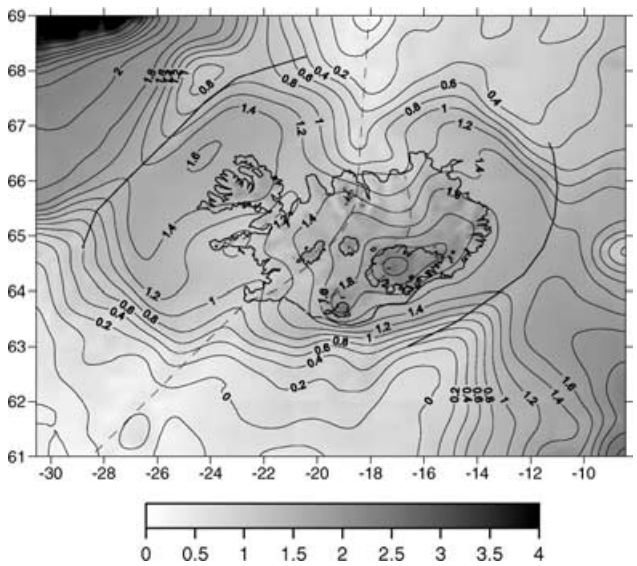

Figure 3. The low-passed residual adjusted topography after removal of the effect of normal oceanic topography and numerical densification of ice, water and surface rocks to density value of 2670 kg m^{-3}. It is obtained by adding the correction for the normal oceanic topography to the adjusted topography. The low pass filter reduces wavelengths less than 70 km. Solid lines indicate boundaries between the Icelandic plate and the Iceland–Faeroe and the Iceland–Greenland Rises. Dashed lines indicate the position of the ridge axes.

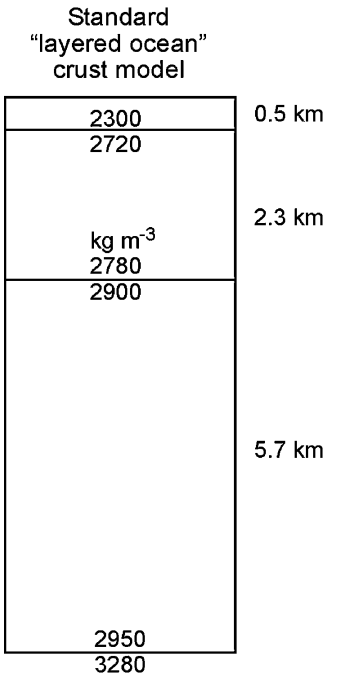

Figure 4. Standard seismic layered model for North Atlantic oceanic crust showing density and thickness of individual layers. It is used for reference in calculating residual Moho depth.

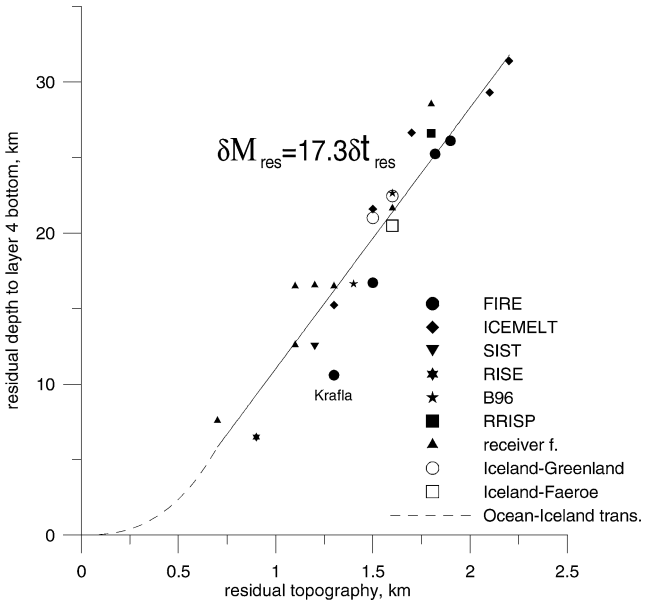

Figure 5. Residual Moho, i.e. the residual depth to the bottom of Layer 4, determined from seismic data minus the standard ocean column shown in Fig. 4 plotted against the residual adjusted topography as in Fig. 3. This relationship characterizes the average density contrast at the base of Layer 4 and is used to make a contour map of this boundary.

the adjusted topography (Fig. 3), which varies from 0 to +2.2 km within the study area, manifests deflections from 'ideal' oceanic topography. We assume variations within the range ±0.5 km to be attributed to the local irregularities of the normal ocean crust and within this range the total thickness and structure of the crust is equal to the standard ocean column shown in Fig. 4. We furthermore define the value of +0.5 km (Fig. 4) as the transition from the Atlantic Ocean to the Icelandic block and the rises.

For the residual topography greater than 0.5 km we establish a statistical relationship between the residual adjusted topography and the residual depth to the bottom of Layer 4 as it is defined by the available seismic data (Staples *et al.* 1997; Menke *et al.* 1998; Darbyshire *et al.* 1998, 2000a; Menke *et al.* 1996; Bjarnason *et al.* 1993; Schlindwein 2001; Du & Foulger 1999, 2001; Weir *et al.* 2001; Reid *et al.* 1997).

The relationship obtained is shown in Fig. 5. It shows a robust linear relationship between the residual values of the topography and the residual depth to the bottom of Layer 4. The difference between the best fit line and the refraction/reflection points are in most cases within 2 km in residual depth. Variations of 2 km are comparable to the accuracy of the seismic results. A notable exception is the area around Krafla volcano where the crust is about 5 km thinner than expected from the best fit line. The difference in the receiver function data is up to 3 km, thus, it also corresponds to the initial data quality.

The slope of the best fit line in Fig. 5 is 17.4 which is more than twice as large as normally found for continents (e.g. Artemjev *et al.* 1994). This means that the average density contrast at the bottom of Layer 4 is small, about 154 kg m^{-3}. A still lower value, 89 kg m^{-3} was found by Menke (1999) by analysing the relationship between the observed topography and the bottom of Layer 4 (Moho). Both indicate that either the lower crustal density is very high or the upper mantle material is very light; a joint effect is also possible. The difference between these two results is probably due to the fact that Menke (1999) used the observed values instead of the residual ones and did not take into account near surface density variations.

We use the linear relation shown in Fig. 5 to introduce the deviations to the initial 'normal' ocean model and to calculate the Moho depth (bottom of Layer 4) in areas without seismic data. This procedure was performed in the following way. The initial seismic determinations were put on the grid, the same as for other data sets. For the points located 50 km or more from any point containing the data we assign values according to the relationship in Fig. 5 and adding back the parameters of the 'standard' oceanic crust model. Then, we interpolate the grid by applying a standard kriging method.

The final map of the Layer 4 bottom is shown in Fig. 6 together with the values from seismic surveys. The discrepancies with the seismic data don't exceed 1 km and they are due to smoothing in the course of interpolation. According to this model, the depth to the bottom of the intermediate 4th layer varies from 16 km to 40 km beneath Iceland. Beneath the Iceland–Faeroe Rise it is about 30 km in accordance with the data from the FIRE profile. In the Atlantic Ocean the corresponding depth to the Moho is equal to 10–11 km.

The top of Layer 4 is here taken to be about the 7.0 km s^{-1} isovelocity surface. Since it is not well defined as a reflective boundary we decided not to introduce small scale details in the existing map (Pálmason 1971). According to this map the depth to the top of Layer 4 varies within 12–15 km beneath Iceland. For the Iceland–Faeroe Rise we use the data from FIRE profile (Richardson *et al.* 1998), where this depth is almost constant within the study area (about 15 km). The variations of this boundary are relatively small and are not as important for the gravity modelling as the bottom. They are shown in the following figures presenting cross-sections of the lithosphere.

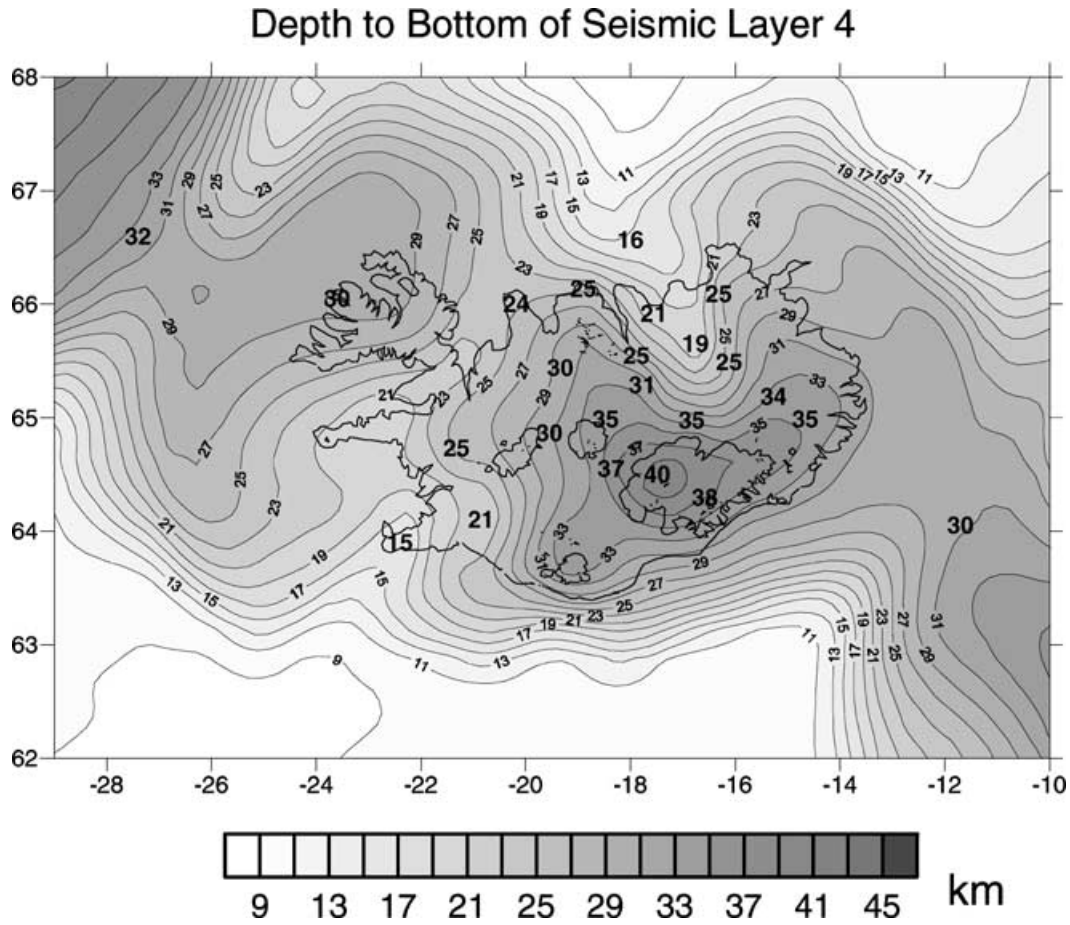

Figure 6. Map of the Moho beneath Iceland. The contours show the depth (km) to the bottom of the 4th seismic layer (Moho) based on seismic data and the relation in Fig. 5. The isolated numbers show the actual seismic results. The seismic results are from Staples *et al.* (1997) (E and NE Iceland), Menke *et al.* (1998) (NE-Iceland), Darbyshire *et al.* (1998) (North and Central Iceland), Menke *et al.* (1996) (S-Iceland), Bjarnason *et al.* (1993) (SW-Iceland), Weir *et al.* (2001) (Reykjanes peninsula) and Reid *et al.* (1997) (NW Iceland).

4.3 Densities of the lithosphere layers

In Table 1 we introduce three different models for the Icelandic crust and upper mantle. All these models have the same seismic properties but they differ only in the density of Layer 4. Model 1 (the thin crust model) assumes mantle-like density in Layer 4, Model 2 (the thick crust model) assumes typical lower crustal densities in Layer 4 and Model 3 (the transition model) assumes density value in-between crustal and mantle values. In principle the three models only differ in that for Model 1 the top of Layer 4 is assumed to be Moho, the main density boundary, while for Model 2 the reflecting bottom of Layer 4 is taken to be 'normal' Moho (Fig. 6). In the third model Layer 4 is a transition zone between crust and mantle.

Table 1. Seismic velocities and estimated densities of the various layers of crust and upper mantle for the three different gravity models.

Layer	Seismic model	Average density value		
		MODEL-1	MODEL-2	MODEL-3
	P-wave velocity km s^{-1}	Thin crust kg m^{-3}	Thick crust kg m^{-3}	Transition model kg m^{-3}
Upper crust (topography)	2.0–4.0	2520	2520	2520
Upper crust (layers 1–2)	3.0–6.5	2662	2662	2662
Lower crust (Layer 3)	6.5–6.8	2930	2930	2930
Seismic Layer 4 (crust or upper mantle)	7.0–7.2	3150	2970	3050
Seismic layer 5 (mantle)	7.5–7.7	3280	3280	3280
Asthenosphere	?	3180–3280	3180–3280	3180–3280

Average density of topographic features has been determined to be 2520 kg m^{-3} as described above. For estimation of average density values for the upper crustal layers below the topography as well as for Layer 3 (Genshaft *et al.* 1993), we have mainly used the results published by Carlson & Herrick (1990) who presented a comprehensive review of the density properties of the Atlantic crust. Although sediments are negligible within the crust in Iceland their thickness can locally be quite large offshore. Seismic data indicate a few hundred metres of sediments on the southern shelf while they can exceed 2 km on the shelf north of Iceland (Flóvenz & Gunnarsson 1991). The effect of sediments is taken into account while constructing the final model.

Carlson & Herrick (1990) determined the following velocity-to-density relationship for the normal oceanic crust in the North Atlantic:

$$\rho = 1000[(3.81 \pm 0.02) - (6 \pm 0.1)/V_p]\,(\mathrm{kg\,m^{-3}}) \quad (2)$$

By applying this relationship we obtain an average density 2970 kg m^{-3} for Layer 4. This value agrees with the results published by Christensen & Mooney (1995).

If we assume that the intermediate Layer 4 represents an anomalous mantle, its density might be about 3150 kg m^{-3} taking into account the velocity decrease (e.g. Sobolev *et al.* 1996), though the last value crucially depends on the melt content and may be changed within some interval. The value for the transition Model 3 is taken as 3050 kg m^{-3}. Its geological significance will be discussed later. For sub-Moho density (Layer 5) we use the value of 3280 kg m^{-3} for the density keeping in mind that mantle density varies from 3180 to 3280 kg m^{-3} according to the 'cooling' lithosphere model (e.g. Hager 1983).

5 EFFECTIVE ELASTIC PLATE THICKNESS OF THE LITHOSPHERE

Elastic properties of the lithosphere have been studied extensively during the last three decades since Dorman & Lewis (1970) introduced the cross-spectral technique for investigations of isostasy. A calculated value of the effective elastic plate thickness may be influenced by different properties of the lithosphere (Burov & Diament 1995) but the general tendency is that this parameter is related to its thermal state (e.g. Lowry & Smith 1995). In the ocean regions the base of the 'elastic' lithosphere is generally attributed to a specific isotherm: from about 450 °C (e.g. Watts 1992; Wessel 1992) to 600–750 °C (Morgan & Forsyth 1988; Chen & Morgan 1990; Neuman & Forsyth 1993; Burov & Diament 1995). Thus, determination of T_e for the Icelandic lithosphere may provide some indications of its thermal state.

The main problem with the determination of T_e using cross-spectral analysis of the gravity field and subsurface load (both admittance and coherence methods) is that this load is often internal, which always leads to big systematic errors of the T_e values obtained (Artemjev & Kaban 1991; Macario *et al.* 1995). To overcome this problem we use a method described by Kaban & Mooney (2001) where the admittances are calculated from the gravity model instead of using theoretical formulae.

We assume that small scale variations of the near surface load in Iceland (e.g. recent volcanic edifices deposited outside the main rift zone and the main ice caps) are supported by elastic deformation of the lithosphere. The large scale variations are, on the contrary, assumed to be supported by the upper-mantle density inhomogeneities or were originally compensated more locally during the formation of the lithosphere and then 'frozen' into it. We analyse the relationship between the gravity field and the near surface load for wavelengths less than 150 km. However, this value is not well defined, which should be considered in interpreting the results.

We take the initial lithosphere model including all boundaries as described above and filter out variations with wavelengths less than 200 km. This model is assumed to be locally compensated by corresponding density inhomogeneities in the upper mantle. Then we add the small-scale near-surface load and calculate the 3-D deformations of all boundaries assuming different T_e values (flexural rigidity of the lithosphere). For the calculation of elastic deformations we use the methods described by Kuang *et al.* (1989) and Kaban & Yunga (2000). We calculate the gravity effect of these deformations on the surface using the same algorithm as Artemjev & Kaban (1994), based on the formulae of Strakhov *et al.* (1989). The programme computes the sum of the gravity influence of elementary volumes corresponding to the initial grids. Then the transfer functions (admittances) are calculated for the fields estimated for different elastic and density models, exactly in the same way as for the observed gravity field and the adjusted topography. The advantage of this approach is that it provides a possibility to take into account huge variations of the depth to the main compensation boundaries instead of assuming some average value while using the theoretical formulae.

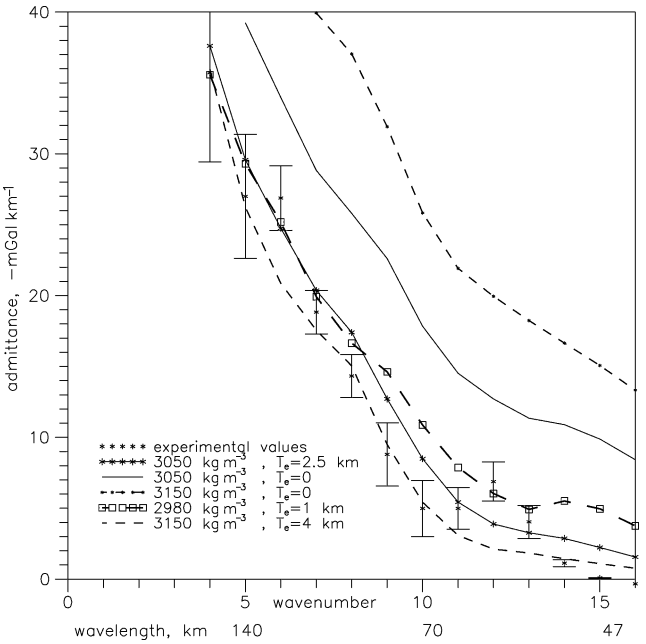

Figure 7. Model admittance and values based on observed gravity used for determination of the effective elastic thickness of the Icelandic block. Two parameters are varied: average density of Layer 4 and effective elastic plate thickness (T_e). See text for more details.

The model admittances together with the values obtained based on the observed gravity field are shown in Fig. 7. We varied two parameters of the models: the average density of Layer 4 and the effective elastic plate thickness, trying to fit the observed values. Both of these parameters affect the admittance similarly. Thus, a stable unique solution is possible only if we fix one and change the other. In this way we determined an effective elastic plate thickness for each of the base values of density of Layer 4: 2970, 3050 and 3150 kg m^{-3}. An increase of the average density of Layer 4 leads to increase of absolute values of the admittance because the principal density contrast moves upward. Vice versa, an increase of the effective elastic plate thickness causes a decrease of the modulus of the admittance.

For a density value of 2970 kg m^{-3} for Layer 4, the most appropriate effective elastic plate thickness is equal to 1 km, for 3050 kg m^{-3} $T_e = 2.5$ km and for 3150 kg m^{-3} $T_e = 4$ km, (Fig. 7). For comparison, in this figure we also show the admittance obtained for the same density but zero elastic plate thickness except for 2970 kg m^{-3} when it is almost the same because of the small T_e value. It is also noteworthy that the fit between the model and the observed curves remains over the whole wavelength range from 150 to 40 km. If there is some influence of the internal load on the lowest considered wavelength the fit would be remarkably different, since that influence must decrease with decreasing of the wavelength.

This analysis has shown that possible values of the lithosphere effective elastic plate thickness in Iceland are similar to values in the range of 2–6 km determined for the Reykjanes Ridge and the East Pacific Rise (Bowin & Milligan 1985; Cochran 1979). It is important to note that if we decrease the density of Layer 4 to its absolute limit and practically join Layers 3 and 4 into one thick lower crustal Layer 3, thus assuming cold thick lithosphere, we obtain near zero effective elastic plate thickness. This combination of the parameters is hardly possible. We cannot distinguish between the two models with the higher density values in Layer 4: both of them have an elastic thickness within permissible limits.

Another factor which may lead to an error in the determination of the effective elastic plate thickness is if the plate is broken along the rift zone so in fact we have a contact of two plates. Previous estimates show that in this case the real effective elastic plate thickness could be twice as great but only in case of deformation under large scale topography. For short wavelength deformations this factor is less important but however, we tried to test this possibility. For one profile across Iceland we estimated elastic deformations assuming contact of two plates in the area of the volcanic zone using the method of Sheffels & McNutt (1986). It turns out that even in the extreme case (when we assume plate contact just under the maximum feature of the external load) the elastic plate thickness is not more than 1.5 times greater than the effective one. Thus, for Iceland this value can still not be greater than 6 km for the dense case of Layer 4. It is also important to note that in the previous studies of mid-ocean ridges mentioned above the authors used the same hypothesis about continuous elastic plates, thus the results are directly comparable.

The main conclusion of this Section is that the effective elastic plate thickness of the Icelandic lithosphere is small, hardly more than 6 km, approximately the same as for the Reykjanes Ridge, and that the lithosphere cannot support elastically significant long term stresses. This conclusion is important for further analysis. Our results indicate also that the thermal state of the Icelandic block and the nearest mid-ocean ridge is similar on the average. This disagrees with the results of Menke & Levin (1994) and Menke *et al.* (1995) who argued for the cold lithosphere under Iceland but agrees with the results of their opponents (e.g. White & McKenzie 1989; Flóvenz 1992; White *et al.* 1995).

6 EVALUATION OF DIFFERENT MODELS

6.1 Isostatic compensation of the crustal structures and the thermal state of the upper mantle

By assigning different density values to seismic Layer 4, we create different lithosphere models based on the same seismic velocity structure. We have shown that the lithosphere in the study area is not rigid enough to support significant stresses caused by external and internal loads. Thus, structures with horizontal dimensions greater then 35–50 km are almost fully compensated in an isostatic sense and the total sum of crustal density variations for each crustal model should be equal with an opposite sign to the total sum of anomalous masses within upper mantle. Actually the support from the upper-mantle density inhomogeneities is not only due to isostatic effects but also due to dynamic effect of upwelling material. The former is proportional to the total sum of anomalous masses but the latter is proportional to the vertical velocity gradient in the upwelling mantle flow multiplied by viscosity (e.g. Pari 2001). However, the last component may be neglected for the uppermost mantle which is confirmed, for example, by the fact that kernels for a dynamic topography are close to -1 near the surface (e.g. Corrieu *et al.* 1995). Sources that are located deeper cause near constant level of the dynamic topography within the study are a and we do not need to consider them.

The integrated anomalous density of the upper mantle can be expressed through the following equation (Hager 1983).

$$\int_{topo}^{L_4} \Delta\rho(h)\left(\frac{R-h}{R}\right)^2 dh = -\int_{L_4}^{L_{\max}} \Delta\rho(h)\left(\frac{R-h}{R}\right)^2 dh, \tag{3}$$

$\Delta\rho(h)$ is the anomalous density at depth h (relative to any standard model), L_4 is the depth to the bottom of seismic Layer 4 and $L_{\max}$ is the maximum depth to the ocean lithosphere bottom. This equation is written for a spherical earth, R is the earth's radius.

We assume that the density variations in the right part of the above equation are due to temperature variations in the upper mantle. For determination of the lithosphere–asthenosphere temperature profiles we use the following equation (Hager 1983):

$$T(h) = T_0\left\{\frac{(R-L_{\max})h}{(R-h)L_{\max}} + \frac{2R}{(R-h)\pi}\sum_{n=1}^{\infty}\frac{1}{n}\sin\left(\frac{n\pi h}{L_{\max}}\right) \times \exp\left(-n^2\pi^2 k_t t / L_{\max}^2\right)\right\} \tag{4}$$

where $T_0 = 1300\ ^\circ$C is temperature at the depth $L_{\max} = 100$ km, $k_t = 10^{-6}$ m^2 s^{-1} is the thermal diffusivity and t is the age. The equation is written for the spherical earth.

Eqs (3) and (4) are used together in the way described below to calculate temperature distribution in each point for all evaluated models. The depth to 1200 °C isotherm along two profiles as well as other parameters are shown on the following figures.

(1) For each point, the left part of eq. (3) gives us the sum of the total anomalous masses in the crust, down to the bottom of Layer 4.

(2) By application of eq. (4) a temperature–depth profile is calculated at 0.1 My intervals.

(3) By using a value $3.3 \times 10^{-5}\ ^\circ\text{C}^{-1}$ for the coefficient of thermal expansion (Hager 1983) the temperature-depth profile in the mantle can be converted into density-depth relationship. Thereafter the integral of the anomalous density, caused by each temperature depth profile is calculated. This integral corresponds to the right hand side of eq. (3). Then we find the effective age t which makes the value of the integral equal to the left hand side of eq. (3).

The parameter t is artificial, and means that the thermal state of the upper mantle at a given point corresponds on the average to the thermal state of 'normal' ocean lithosphere with age t. It is only used to calculate a series of possible temperature profiles for comparison with the gravity data. The 1200 °C isotherm marks approximately the bottom of the lithosphere, the most informative parameter of the models. At greater depths the temperature distribution is near adiabatic and temperature changes insignificantly (e.g. Anderson 1989). Above the 1200 °C isotherm, the temperature–depth relation is essentially non-linear according to the above equation for a cooling lithosphere.

The cross-sections in Figs 8 and 9 show the main gravity boundaries together with the position of the 1200 °C isotherm. One cross-section coincides approximately with the 'FIRE' profile from

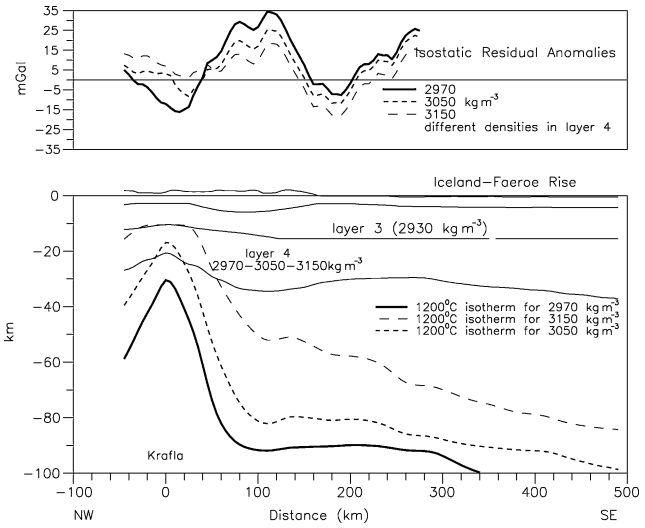

Figure 8. A profile showing a cross-section from the NE volcanic zone in Iceland to the Iceland–Faeroe Rise, approximately along the FIRE profile. The location of the profile is shown in Fig. 1. The lower section shows the main boundaries and the calculated 1200 °C isotherm for different densities of Layer 4. The corresponding residual isostatic anomalies are shown in the upper section. The higher density values in Layer 4 give clearly less isostatic anomalies over Iceland.

Krafla to the Faeroe Islands (Fig. 8) but the other one crosses Iceland from north to south through the area near Krafla (Fig. 9). For both profiles zero distance indicated on the horizontal axis corresponds to the position of Krafla. The location of these profiles is shown in Fig. 1. From the profiles the following results emerge:

(i) For Model 2 with the value of 2970 kg m^{-3} in Layer 4 (modifications of Layer 3) the minimum depth to the 1200 °C isotherm is under Krafla (about 30 km), but the maximum is under the eastern and south-eastern parts of Iceland (up to 90 km).

(ii) For Model 1 with the value of 3150 kg m^{-3} in Layer 4 (anomalous mantle), the 1200 °C isotherm appears to be directly under the crust near the volcanic zones. Moreover, under the central part of the eastern volcanic zone (EVZ) we can't explain isostatic balance of the lithosphere only by its thermal state if we limit the uppermost position of the 1200 °C isotherm to 10 km. It is necessary to assume some density variations or/and dynamic support. In other parts of Iceland this depth varies from 25–50 km for Model 1.

(iii) According to the Model 3 with intermediate density values for Layer 4 (3050 kg m^{-3}) the position of the 1200 °C isotherm is at a depth of about 20 km under Krafla, about 80 km in the eastern and south-eastern parts of Iceland and 25–50 km in other places.

In general, we may conclude that for all models the position of the 1200 °C isotherm under the EVZ around and south of Krafla is relatively close to the surface: just under the crust for the 'thin' crust model and close to the bottom of Layer 4 for the models with lesser values of the density in this layer. In contrast, in the eastern and south-eastern parts of the Icelandic block this position is below the bottom of the Layer 4 for all models.

6.2 Calculation of the gravity effect of the individual models

Having constructed the initial maps for the main lithospheric boundaries and assigned a density value to each seismic layer, it is possible to estimate the gravity effect of each model and to remove it from the observed gravity field to produce residual anomalies. These anomalies might be also named 'isostatic' anomalies since the models are in isostatic equilibrium according to eq. (3).

In the initial Bouguer gravity anomalies the terrain correction has been applied within a radius of 167 km. We estimate the gravity effect of different lithospheric layers within the same radius. The influence of the structures outside the calculation area is taken from the global gravity model of Kaban *et al.* (1999). The gravity anomaly of any layer or boundary within the Earth's crust and mantle is calculated as a sum of the gravity influences of elementary volumes

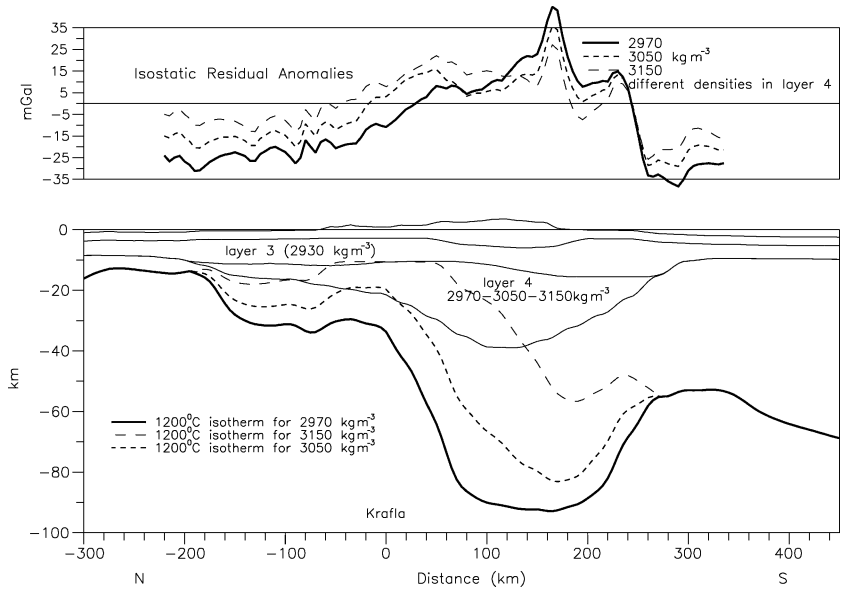

Figure 9. A profile showing a cross-section from the normal oceanic crust south of Iceland, into the Icelandic block and along the NE volcanic zone of Iceland close to Krafla. The location of the profile is shown in Fig. 1. The lower section shows the main boundaries and the calculated 1200 °C isotherm for different densities of Layer 4. The corresponding residual isostatic anomalies are shown in the upper section. Like in Fig. 8, the higher density values in Layer 4 give clearly less isostatic anomalies over Iceland and the junction between Iceland and the normal oceanic crust off the south coast.

corresponding to the initial grids within a radius of 167 km using formulae provided by Strakhov *et al.* (1989). Boundaries of the volumes correspond to the seismic layers. Vertical density gradient in the mantle is modelled by division of an initial column into a few parts with different densities to attain the required accuracy. The accuracy of the direct gravity calculations is within 1 mGal. The initial model is defined within the area of 61°–69°N and 30°–9°W and the calculated gravity fields cover the areas between 62.5°–67.5°N and 27°–12°W. All model parameters, as well as the initial gravity and topography data, were prepared on the same 2.5′ × 6′ (4.63 × 4.7 km) grid, but the parameter grids have additional 167 km frame corresponding to the gravity grid.

The gravity field produced by each density model has been computed and removed from the Bouguer anomalies together with the effect of the distant zone beyond 167 km. The residual isostatic anomalies (RIGA), obtained for different density models are shown in Figs 8 and 9 along the analysed profiles and in the maps in Figs 10–12: for the thin crust model in Fig. 10, for the thick and light crust model in Fig. 11 and for the intermediate model in Fig. 12. Constant value is removed from the resulting anomalies. It is caused by the influence of regional gravity maximum and is not a subject of the present study.

To compare our results with the results of 2-D modelling along seismic profiles it is useful to check if there exists a significant disagreement between 2-D and 3-D models in the case of Iceland. Darbyshire *et al.* (2000b) started to analyse this problem by considering the effect of near-surface bodies (volcanoes) and concluded that it is insignificant. A 3-D effect of deep and essentially non-linear structures like the Icelandic mantle plume could be much more important because the amplitude of a gravity anomaly caused by some feature at the depth Z is proportional to $e^{-2\pi Z/L}$, where L is the wavelength. In Fig. 13 we show the north–south profile with the gravity effects of the bottom of Layer 4 and of the upper mantle temperature anomaly estimated for the intermediate density model (3050 kg m^{-3}) using 3-D and 2-D methods. The difference between the curves showing the influence of the crust-mantle boundary reaches 20 mGal on the 200 km interval in central Iceland which value is essential for evaluation of the results. An even more pronounced systematic discrepancy of up to 60 mGal is found for the gravity field of the mantle density anomaly under Iceland. Thus, we conclude that for the gravity modelling of the Icelandic lithosphere the use of a 3-D calculation is necessary.

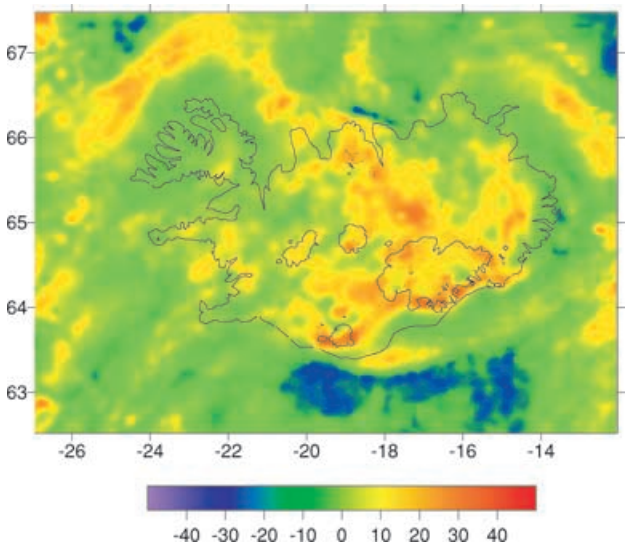

Figure 10. The residual isostatic gravity anomalies (RIGA) for the thin crust model, density of the Layer 4 is 3150 kg m^{-3}. The values over Iceland are around zero but considerable negative anomalies are still observed just south of the coast. The negative anomaly just off the central northern coast of Iceland is due to a few kilometres thick sediment basin associated with the Tjörnes fracture zone (Pálmason 1974).

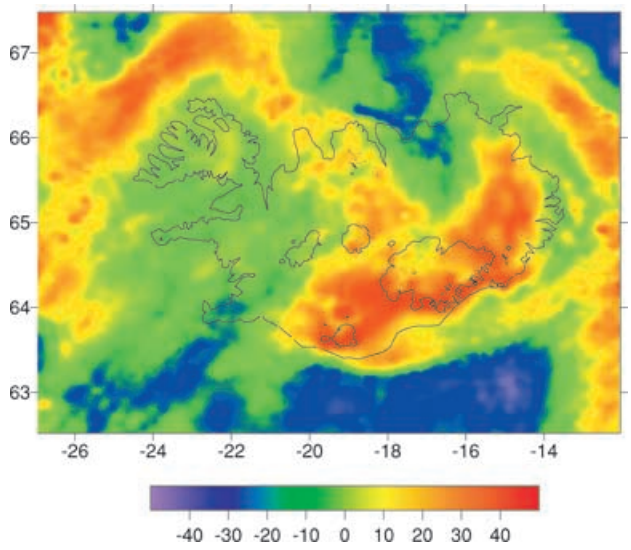

Figure 11. The residual isostatic gravity anomalies (RIGA) for the thick and light crust model, density of Layer 4 is 2970 kg m^{-3}. The amplitudes of the isostatic anomalies are much higher than for the thin crust model in Fig. 10.

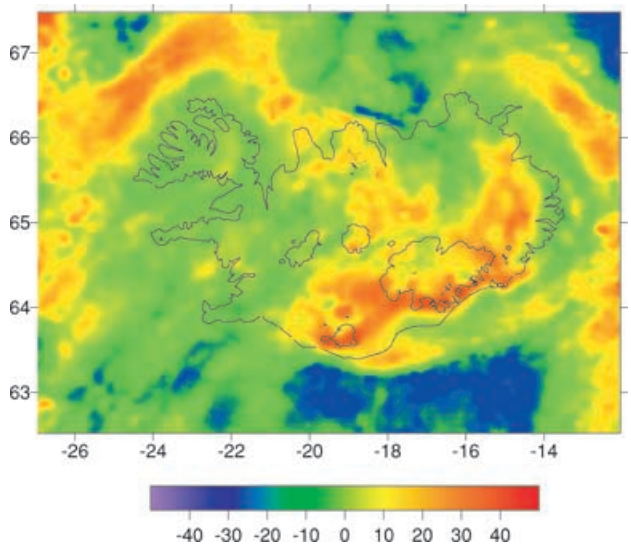

Figure 12. The residual isostatic gravity anomalies (RIGA) for the thick and dense crust model, density of Layer 4 is 3050 kg m^{-3}. The amplitudes of the isostatic anomalies are somewhat higher than for the thin crust model in Fig. 10 but lower than for the thick crust model in Fig. 11.

The most significant RIGA have been obtained for a thick and relatively light crust model (2970 kg m^{-3} in the 4th layer, Figs 8, 9 and 11). There is a significant difference in gravity values along the ridge axis for Iceland compared to the surrounding ocean. The deepest minimum in the residual anomalies which exceeds −45 mGal exists off the southern shelf of Iceland to the east of the Reykjanes ridge. In the southern part of Icelandic block the amplitude of RIGA

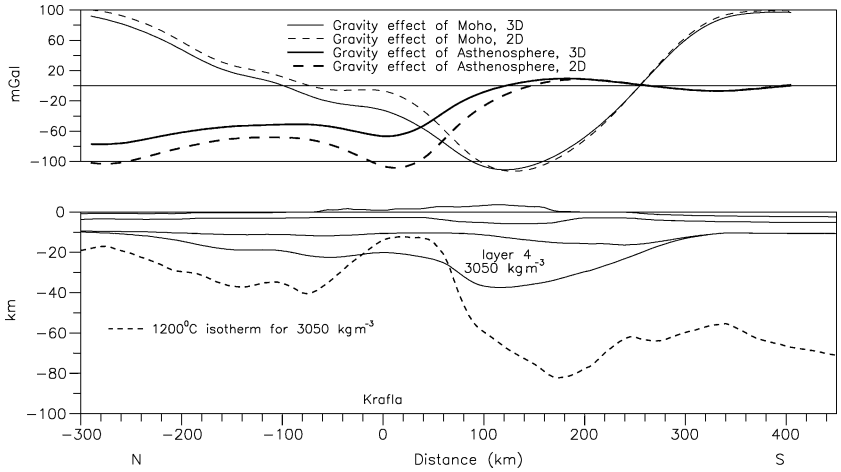

Figure 13. A north–south profile (location in Fig. 1) showing comparison of 2-D and 3-D calculation of the gravity effects of the bottom of Layer 4 and of the upper mantle temperature anomaly estimated for the intermediate density model (3.05 g cm^{-3}). The figure clearly shows the importance of using 3-D calculations.

is about ±45 mGal (Fig. 11). Normally, such amplitudes of isostatic anomalies are observed in continental collision zones. For areas near mid-oceanic ridges such big disturbances can't be caused by disturbances of the isostatic equilibrium itself because the lithosphere is weak and cannot support such stresses. These results, together with the results of the admittance analysis give us strong reason to reject the model with thick continental-type and relatively light crust.

For the thin and the thick-but-dense intermediate crust models the variations of isostatic anomalies are within −30 to +20 mGal (Figs 10 and 12 correspondingly). The dense crust model (3150 kg m^{-3} in Layer 4) works better in general, especially in the eastern and south-eastern parts of Iceland, but in the area around Krafla the intermediate model provides lesser isostatic anomalies. This result gives us a good reason to construct iteratively a model with density variations within Layer 4. We will discuss this possibility in the following Section.

6.3 Inversion of the residual anomalies

6.3.1 Theory

Residual isostatic gravity anomalies (RIGA) mostly indicate deviations of the lithosphere–asthenosphere model from the reality (errors in the initial gravity model). By an inversion we are going to change the initial models to provide an acceptable fit to the observed gravity (near zero residual anomaly). A gravity inversion problem is not unique so the source of the anomaly should be defined as precisely as possible. Our approach to the inversion is based on the method suggested by Müller & Smith (1993) for interpretation of the gravity anomalies in the central Atlantic with some modifications. We assume that the residual anomalies are due to deflections of the main lithospheric boundaries from the initial positions. Then, we assume, based on the existing seismic profiles, the variations of the main crustal boundaries to be conformal, i.e. a deflection in one boundary is accompanied by corresponding deflections in the other boundaries probably with different amplitude. This is visible, for example, in the area around the eastern volcanic zone where the top and the bottom of Layer 4 domes up, as well as in the transition zone from the Icelandic block to the Iceland–Faeroe Rise. Later we will discuss possible effects if our assumptions are not completely valid. These variations are assumed to be the result of a folding process or/and sublithospheric load of the mantle plume.

The local negative RIGA may be also caused by low density bodies (presumably magma chambers) at depths 0–10 km. This effect can be significantly reduced by low-pass filtering of RIGA which also removes the influence of isostatic disturbances caused by elastic support of small scale features of external load. Variations of the crustal boundaries are to be accompanied by corresponding changes of the depth to the asthenosphere top to provide isostatic equilibrium of the crust-mantle column.

The main principle of the gravity inversion includes the following steps:

(i) Removal of non-significant small scale features with wavelength less than 50 km from the residual anomalies.

(ii) Use of linear inversion in the Fourier domain (Müller & Smith 1993) to determine the first approximation of the model corrections based on the RIGA.

(iii) Introduction of necessary corrections to the initial models and to estimate once again the residual anomalies. The reality is essentially non-linear, thus the result will be non-zero as could be expected.

(iv) Repetition of steps i–iii until the residual anomalies do not change significantly.

(v) Analysis of an alternative model in which the initial boundaries of Layer 4 are fixed while the same inversion technique is used to determine the horigontal density variations within this layer.

According to Parker (1972), the expression for the gravity effect of a density contrast with varying elevation $h(x, y)$ is

$$F[g(x, y)] = 2\pi G\rho_j e^{-|k|Z_j} \sum_{n=1}^{\infty} F[h(x, y)^n]|\vec{k}|^{n-1}/n! \qquad (5)$$

$|k| = 2\pi/L$, L-wavelength,

where F indicates the Fourier transform, $g(x, y)$ the gravity anomaly, G is the gravitational constant and Z_j is the mean depth to the density contrast. We use only the first linear term ($n = 1$) and in the case of several boundaries the input of their disturbances into the residual isostatic anomalies will be:

$$F[g(x, y)] = 2\pi GF[h(x, y)] \sum_{1}^{m} C_j\rho_j e^{-|k|z_j} \qquad (6)$$

where ρ_j are density contrasts for corresponding boundaries and C_j are the coefficients which control an 'input' of each boundary into the inversion or its 'weight'. In the case of isostatic compensation:

$$\sum_{1}^{m} C_j \rho_j = 0 \tag{7}$$

Upper-mantle density inhomogeneities are modelled during the inversion by an artificial boundary placed at a depth of 50 km. The coefficient C_m for this boundary is determined based on the previous equation. Thus the disturbances to the principal density boundary (bottom or top of the intermediate Layer 4) may be computed as:

$$h(x, y) = F^{-1}\left(F(g(x, y)) \Big/ \left(2\pi G \sum_{j=1}^{m} C_j \rho_i e^{-|k|z_j} \right) \right), \tag{8}$$

where F^{-1} is an inverse Fourier transform.

This expression represents an unstable inverse problem. Stable estimates of $h(x, y)$ may be obtained by bandpass filtering of RIGA before making inversion. The filtered field should not contain significant low or high wavelength components. We do this during the inversion by multiplying the Fourier representation of RIGA by:

$$F(g(x, y)) \exp\left(1 - (L_{g\min}/L)^2\right) \quad \text{for} \quad L < L_{g\min} \tag{9}$$

for short wavelengths and by

$$F(g(x, y)) \exp\left(1 - (L/L_{g\max})^2\right) \quad \text{for} \quad L > L_{g\max} \tag{10}$$

to remove the long wavelengths where L is the wavelength and $L_{g\min}$ and $L_{g\max}$ are the lower and higher wavelength limits respectively. It turns out after a few checks that $L_{g\min} = 100$ km is the most appropriate value, which does not lead to any high frequency modulation. This uncertainty can't significantly change the final result (amplitudes of computed boundary disturbances) but can slightly change the shape of the computed anomalies. $L_{g\max}$ does not change the solution because the RIGA for all models have no significant low wavelength component except of the constant value which corresponds to the regional gravity and geoid maximum.

The corrections obtained by the inversion are added to the initial models but with certain limitations: permissible variations of the layer 1–layer 2 boundary should not be outside the Earth surface. Seismic data indicate that variations of the layer 2–Layer 3 boundary are not large and we limited them to ±2.5 km.

At the next step of the process, the gravity influence of the corrected model is calculated which gives a new version of the residual isostatic anomalies. They are not equal to zero because we use a linear inversion and apply some limitations to the solution obtained, but they should be significantly less than the initial RIGA. New residual anomalies may be inverted again until corrections will be negligible.

By assigning different density values to seismic Layer 4 of the Icelandic lithosphere we obtain different solutions. By comparing these results with available data the validity of the different models may by estimated. The main criterium is that the final model has to be consistent with the initial seismic data, in other words, the deviation from the initial model should not be large. In addition, comparison of different parameters of the models obtained (depth to Moho or to the asthenosphere top etc.) with the results obtained by other methods might also be helpful to understand the structure and properties of Icelandic lithosphere.

The described algorithm of inversion is also applicable if we fix initial seismic boundaries and try to fit the observed gravity field by varying density within the intermediate Layer 4. We will try this possibility for the best models checked at the previous stage.

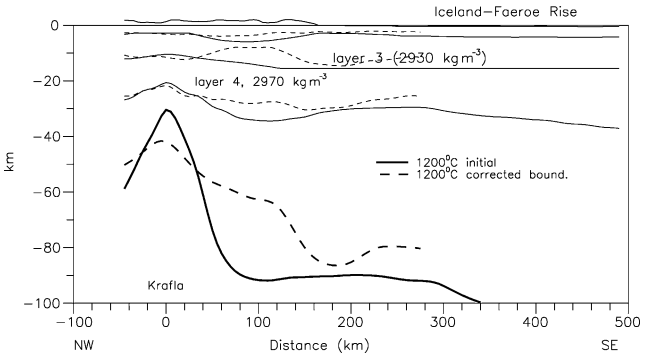

Figure 14. A cross-section from the Iceland–Faeroe Rise to the NE volcanic zone, showing the calculated 1200 °C isotherm and boundaries of seismic layers after the inversion of residual gravity anomalies, assuming the density of Layer 4 to be 2970 kg m^{-3}.

6.3.2 Results of inversion

The inversion technique described above was applied to the RIGA calculated for the different density models. In the inversion we modified the main density boundaries. Since we do not know exactly what is the input of each boundary to the residual gravity, we stabilize the solution by minimizing the corrections to the initial models by distributing them between several boundaries. The corrections can be about 1.4–2 times greater if they are attributed to one boundary only. Therefore, we can justify a rejection of a model because the necessary corrections would be much larger if applied to a single boundary.

For the thin and intermediate models we obtained a stable solution, but for the light thick crust (2970 kg m^{-3} in Layer 4) the residual anomalies are still significant if we apply the bandpass filtering (eqs 9 and 10): in the case of relatively deep main density contrast it is not possible to fit medium and short-wavelength features of the residual anomalies. Thus, for the thick light crustal model the results of the inversion are not successful. For the other models correspondence of the gravity fields at wavelengths greater than 50 km is satisfactory. The corrections for second step iteration for models 1 and 3 (according to Table 1) was at least 4 times less than in the first one and in the third iteration for they were negligible.

In Figs 14 and 15 the results of the inversion for the model 2 (2970 kg m^{-3} in Layer 4) along the above-mentioned profiles are shown. The necessary corrections of the main seismic boundaries reach 10 km in the central and south-eastern parts of the Icelandic block. The bottom of Layer 4 moves up for about 10 km under Vatnajökull in the case if we give approximately the same inversion weights to all boundaries and this uplift may exceed 17 km if we correct the position only of this boundary. Positive anomaly in the central part of Iceland still exists after the inversion. These results are in obvious contradiction with the existing seismic data. Taking into account also the above mentioned results we finally reject this model.

The results of the inversion for the 'dense crust' model (3150 kg m^{-3} in Layer 4) along the two reference profiles are shown in Figs 16 and 17. This model works well in the outer parts of Icelandic block with maximum depth (30–40 km) to the bottom of seismic Layer 4. In the eastern part of Iceland it is necessary to introduce only a small correction to the initial position of Boundaries 3 and 4 within possible errors of their determination. The position of the 1200 °C isotherm moves down here to about 45 km. The temperature at the bottom of the 4th layer is about 950 °C. At the southern edge of the Icelandic block the transition between the plate and

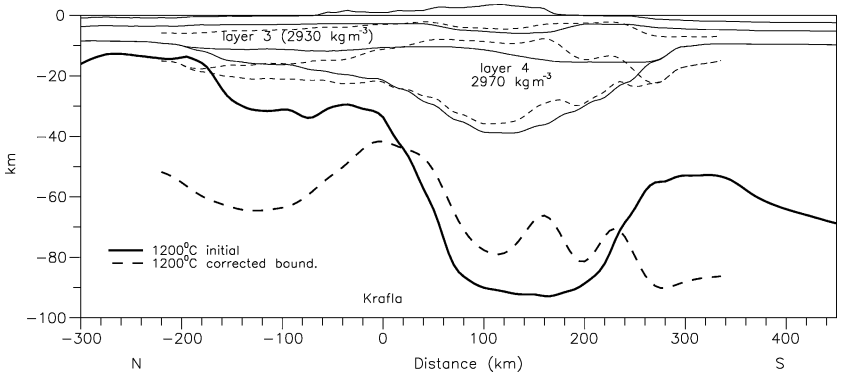

Figure 15. A cross-section from the normal oceanic crust south of Iceland, along the NE volcanic zone near Krafla. It shows the calculated 1200 °C isotherm and boundaries of seismic layers after the inversion of residual gravity anomalies, assuming the density of Layer 4 to be 2970 kg m^{-3}.

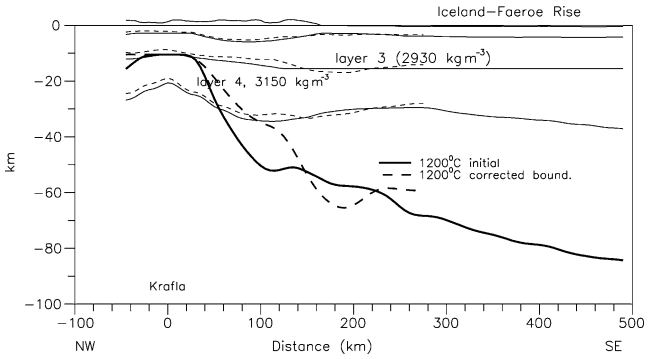

Figure 16. A cross-section from the Iceland–Faeroe Rise to the NE volcanic zone showing the calculated 1200 °C isotherm and boundaries of seismic layers after the inversion of residual gravity anomalies, assuming the density of Layer 4 to be 3150 kg m^{-3}.

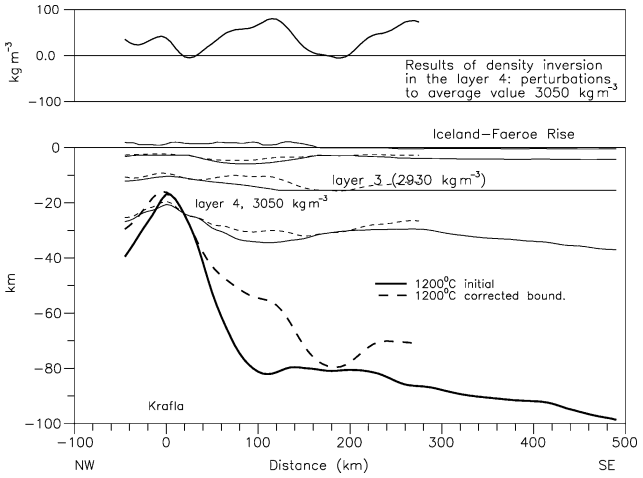

Figure 18. A cross-section from the Iceland–Faeroe Rise to the NE volcanic zone, showing the calculated 1200 °C isotherm and boundaries of seismic layers after the inversion of residual gravity anomalies, assuming the density of Layer 4 to be 3050 kg m^{-3}. See the main text for explanation of the upper part.

'normal' ocean should be more smooth than in the topography. In the central part of Iceland and especially under Krafla, the thermal regime of the upper mantle cannot provide isostatic compensation of the lithosphere structure. After the inversion this situation is changed for the worse. To correct it, it is necessary to decrease crustal density in the volcanic zones, thus to turn to the intermediate model 3.

The results of the inversion for the intermediate model (3050 kg m^{-3} in Layer 4) are shown in Figs 18 and 19. This model works well in the central part of Iceland while in its eastern and south-eastern outer parts necessary corrections of the base seismic boundaries are up to 7 km. If we correct only the position of the bottom of Layer 4, its depth would decrease under the eastern edge of Iceland from 35 to 26 km. Depth to the 1200° isotherm decreases relative to the initial model and is equal to about 60 km in this place. Thus, its position is

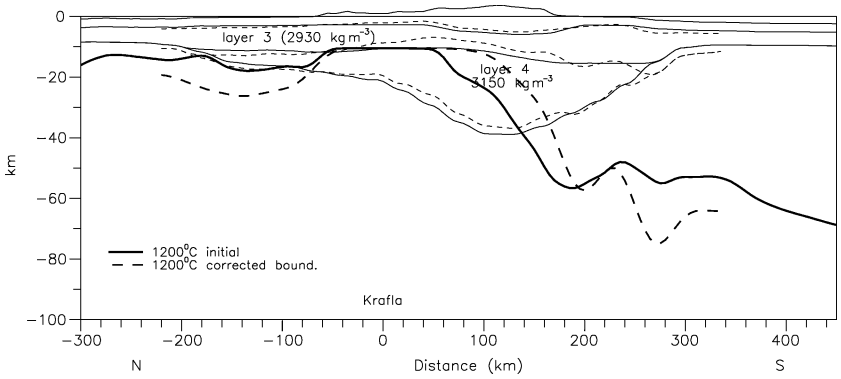

Figure 17. A cross-section from the normal oceanic crust south of Iceland, along the NE volcanic zone near Krafla. It shows the calculated 1200 °C isotherm and boundaries of seismic layers after the inversion of residual gravity anomalies, assuming the density of Layer 4 to be 3150 kg m^{-3}.

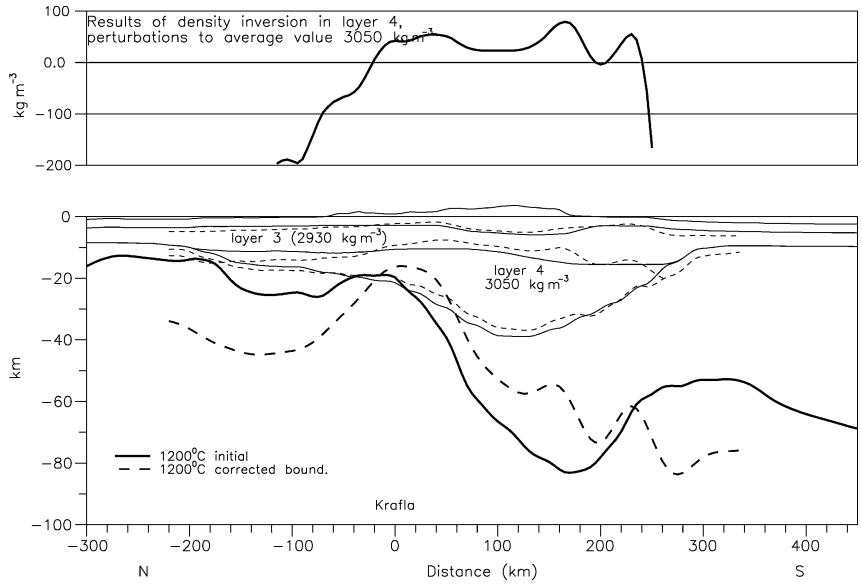

Figure 19. A cross-section from the normal oceanic crust south of Iceland, along the NE volcanic zone near Krafla. It shows the calculated 1200 °C isotherm and boundaries of seismic layers after the inversion of residual gravity anomalies, assuming the density of Layer 4 to be 3050 kg m^{-3}. See the main text for explanation of the upper part.

relatively independent of the initial density model. Under the Krafla area this isotherm penetrates into the Layer 4 and lies near its upper boundary.

The obtained results show that neither Model 1 (dense crust) nor Model 3 (intermediate) fit the observed gravity field in all places. The 'dense' crust model is more appropriate for the parts of Icelandic block with thick crust, while the intermediate model works better in the central area especially near Krafla. We have to keep in mind that the inversion has up to now assumed fixed density of the crustal layers. It is however reasonable to assume that the density within Layer 4 varies in horizontal direction. We tried therefore to perform a extended inversion process by fixing the initial position of the main boundaries except of the asthenosphere bottom (thermal state) and varying density within the intermediate Layer 4. This procedure works only where this thickness is large enough. In the southern part of Iceland where significant residual isostatic anomalies exist we have to change also the depth to the main boundaries to obtain an agreement between the observed and model gravity fields. The intermediate model was used as the starting point for the inversion. It turns out that the largest deviations of the density within the 4th layer is in the eastern and south-eastern parts of Iceland in the places where the bottom is deepest. There the average density here attains a value of 3140 kg m^{-3}, (Figs 18 and 19), but in other parts of Iceland where the thickness of the Layer 4 is large enough the inversion gives no significant density changes. These results confirm the idea that density within Layer 4 varies in horizontal direction. These variations are obviously connected with the thickness of this layer.

The southern shelf where the Icelandic block is in contact with the normal oceanic crust may be considered as a 'key' feature for understanding the lithosphere structure in the region. It turns out that the sharp jump in the gravity field which corresponds to the boundary of Icelandic block as inferred from the topography cannot be explained by the corresponding seismic boundaries variation. It is necessary in all cases to assume significant variations of the main density boundaries. In the case of the dense crust model these variations should be about 2–3 km, 5–7 km for intermediate crust, and 12–15 km for thick-and-relatively-light crust. The last interval seems to be unlikely.

7 DISCUSSION

7.1 Thermal state of Icelandic lithosphere

The thermal state of the crust in Iceland is mainly derived from temperature measurements from boreholes up to 1500 m deep, located outside known areas of hydrothermal activity. The general pattern of the temperature gradients from these boreholes is considered to give a fairly reliable estimate of temperatures in the upper crust (Bodvarsson 1982). The latest map of the heat flow was presented by Flóvenz & Sæmundsson (1993). They attempted to estimate the depth to the 1200 °C isotherm by extrapolating the borehole data assuming a slight increase in thermal conductivity with depth in the upper crust and typical values for the radioactive heat generation in basalts. The 1200 °C isotherm was found to be in the range 10–30 km. Since no undisturbed temperature gradient data exist from the active rift zone the depth could not be estimated there. The results obtained in this paper show that the temperature gradient must vary with depth in the lower crust i.e. in Layers 3 or 4. A similar conclusion was reached by Menke & Sparks (1995).

The map of the 1200 °C isotherm estimated for the corrected combined model of the Icelandic lithosphere with density changes in the intermediate Layer 4 in the eastern and south-eastern parts of Iceland is shown in Fig. 20. The figure shows that the depth to the 1200 °C isotherm is relatively shallow along the axis of the Reykjanes and Kolbeinsey oceanic ridges north and southwest of Iceland and their direct connection below Western Iceland, i.e. beneath the Western Volcanic Zone (WVZ). It deepens with distance from this axis. Shallow depths also appear beneath the northern part of the Eastern Volcanic Zone (EVZ), beneath the Krafla. It should be stressed here that shallow depth to the 1200 °C isotherm under Krafla is purely a consequence of the thinning of the crust reported by seismic studies in the area (Brandsdóttir *et al.* 1997). Beneath the southern part of the EVZ the depth to the 1200 °C isotherm

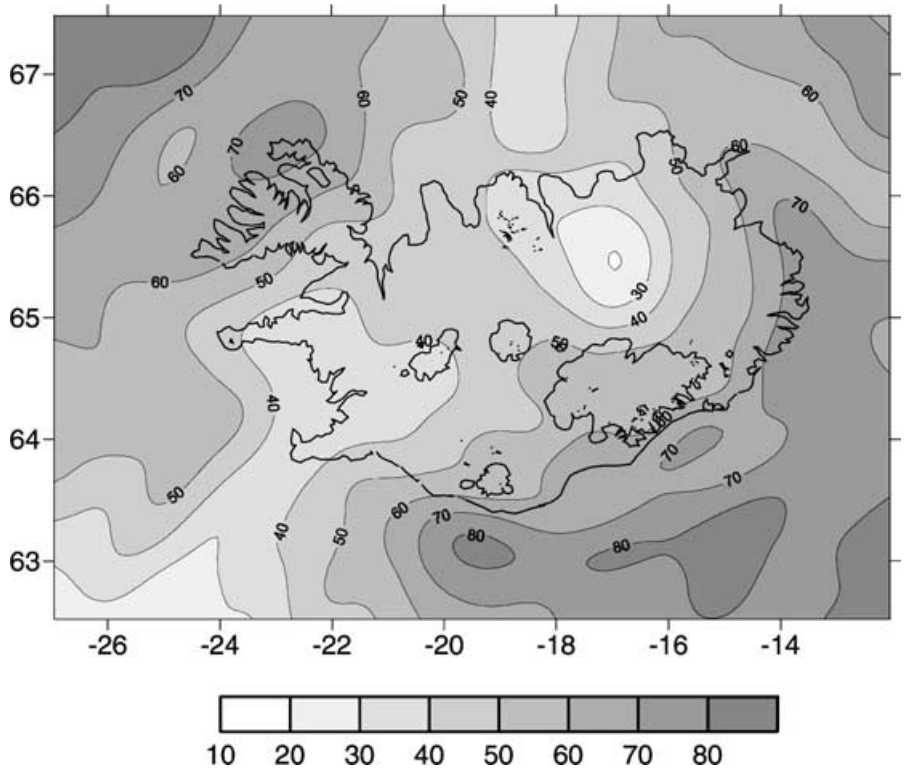

Figure 20. A map of the depth to the 1200 °C in km isotherm based on gravity data interpretation. Note that the anomaly beneath Krafla is a consequence of the seismic results of Brandsdóttir *et al.* (1997) showing shallow depth to the Moho.

increases to 60–70 km. The erupted material in this area is of alkalic composition which indicate relatively deep origin of the magma. It is also interesting to note relatively shallow depth west of the WVZ below the extinct spreading axis there.

7.2 Structure and composition of Layer 4

The nature and composition of Layer 4 is important for understanding the origin and tectonics of the crust in Iceland. The results obtained in this study show that the *average* density within seismic Layer 4 is between 3050 and 3150 kg m^{-3} depending on thickness and thermal state. Seismic data give a near linear V_p–depth relation between 6.9 and 7.4 km s^{-1}. It is reasonable to expect a similar relation for the density. In that case the '*in situ*' density within Layer 4 varies from 2980 kg m^{-3} in the upper part to 3250 kg m^{-3} in the deepest root under the eastern edge of Iceland. The latter value is near 'normal' mantle peridotite density. Thus Layer 4 is neither typical crust nor typical upper mantle, rather a transition zone between them. Another possibility is a frozen high-magnesium primary basaltic melt as suggested by White & McKenzie (1989).

It follows from the results of the admittance study that the seismic Layer 4 cannot support significant long-term stresses but at the same time it is characterized by normal V_p/V_s ratio equal to about 1.76–1.79 (Bjarnason *et al.* 1993; Menke *et al.* 1996; Staples *et al.* 1997; Darbyshire *et al.* 1998). A decrease of this ratio caused by partial melt has not been detected. These data are in agreement with our results of temperature modelling which give temperatures in Layer 4 in the range 650–950 °C. The only exception is in the NE volcanic zone where the 1200 °C isotherm moves up depth of 11 km.

Significant difficulties still exist in combining the results mentioned above with the results of magnetotelluric soundings, according to which a widespread zone of low resistivity exists within the seismic Layer 4. This low resistivity has been explained by partial melt (e.g. Beblo & Björnsson 1980) but from the seismic results of V_p/V_s ratio and Q-value estimates (Menke & Levin 1994), and our temperature estimates from the gravity data, the existence of a widely spread partial melt within seismic Layer 4 is very unlikely. However, it is possible that thin channels and lenses (thin and sparse enough not to be detected by seismic methods) of partially molten material exist within Layer 4.

From our point of view the parameters of the intermediate Layer 4 correspond well to the hypothesis formulated by Cannat (1996): 'The temperature field modelled in these thin crust ridge regions is such that some of the melts extracted from the asthenosphere should crystallize in the mantle, before they reach the crust. This prediction is consistent with observations made on gabbroic and ultramafic samples collected in thick lithosphere/thin crust regions of the Mid-Atlantic Ridge'. In a similar way it is suggested here that the huge productivity of the Icelandic mantle plume in combination with relatively low spreading velocity forms Layer 4, an enormously thick transition zone between the crust and the mantle.

7.3 Iceland plume

On the final map of the 1200 °C isotherm (Fig. 20) one may see an anomalous hot area in the northern and central part of Eastern Volcanic Zone. It is likely due to the influence of the mantle plume. This conclusion is also confirmed by the analysis of the isostatic anomalies. The pattern of these anomalies reflects isostatic disturbances and contains tectonic information. It is interesting that the general form of the local isostatic anomalies is different on the eastern and western sides relative to the main mid-ocean ridge axis. Useful information about the configuration of the main tectonic features may be obtained by computing horizontal gradients of

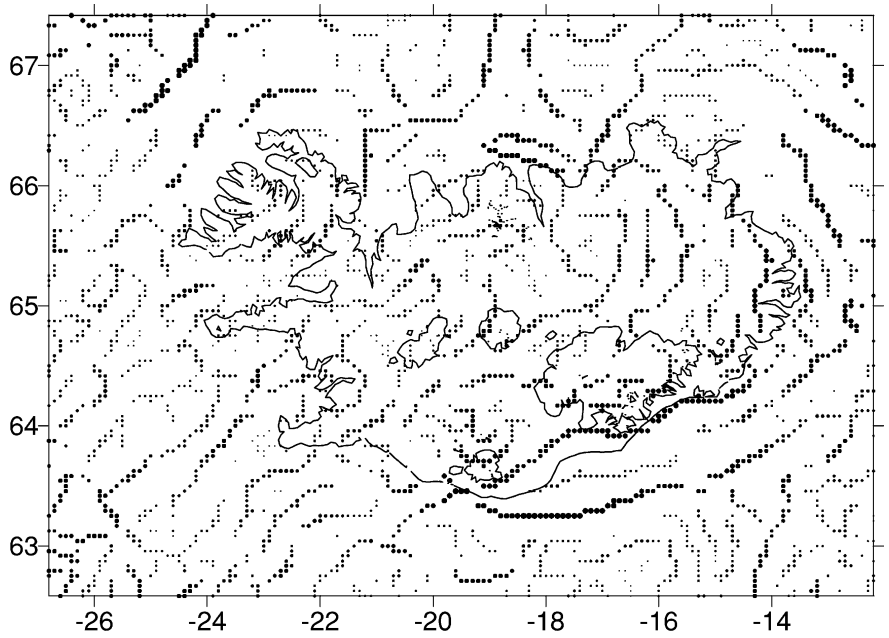

Figure 21. Location of the maximum values of the horizontal gradients of the isostatic anomalies in Fig. 12. This map shows a configuration of the main tectonic patterns in Iceland and surrounding area. Note the semi-circular pattern of the gradients in the eastern and southeastern parts of Iceland with centre somewhere in the NE volcanic zone. West of the eastern volcanic zone the gradients are more or less parallel to the rift axis.

the isostatic anomalies. These gradients show main tectonic boundaries, faults and folding zones. We calculated a map of horizontal gradients of the isostatic anomalies and selected then local maximum values of these gradient values using the method described by Kaban & Artemjev (1996). This map is shown in Fig. 21. Each value exceeds two opposite neighbour points in at least two of four possible directions. On the western side the small-scale features follow in general the ridge direction. On the eastern side short-wavelength features of the residual isostatic anomalies form slightly circular structures centred near the Krafla volcanic zone which coincides with the centre of anomalous hot zone. Circular zones of the horizontal gradient maxima likely reflect folding structures formed under the additional pressure of the plume and might be associated with eastward movement of the mantle plume relative to the plate boundary.

Anomalous 'cold' zones along the eastern and especially the southern edges of the Icelandic block could not be explained by the simple 'cooling' lithosphere model. The depth to the 1200 °C isotherm is up to 90 km in this area, which corresponds to a lithosphere age of about 80 My instead of 10–20 My. This thermal state of the lithosphere may also be explained by the influence of downgoing mantle flows which deform ideal temperature profiles and probably increase mantle density modifying its composition. These flows could be connected with upwelling flow in the centre of the plume. As was mentioned before, the very active Icelandic plume is situated at a relatively slow spreading ridge. Regular spreading is not fast enough to incorporate full production of the plume, therefore the upwelling flow partly turns to the horizontal direction, forming circular folding structures, when it is partly accumulated by the lithosphere and partly returns into the mantle.

8 CONCLUSIONS

In this paper we have used the gravity and topographic maps of Iceland, information on seismic boundaries and typical temperature–age relationships to conclude the following:

(i) A new Bouguer gravity map of Iceland is recomputed based on admittance between the topography and the gravity and with corrections for glacial ice sheets. The most appropriate value for the Bouguer correction is determined to be 2520 ± 15 kg m^{-3}.

(ii) By using available seismic data and finding the relationship between them and the near surface load we have constructed a new Moho map, a map of the depth to the bottom of Layer 4.

(iii) The *average* density of seismic Layer 4, which has been interpreted as either hot partially molten mantle or relatively cold basaltic lower crust, is most likely in the interval 3050–3150 kg m^{-3} and varies depending on the position. We found a general correspondence of the Layer 4 density and thickness: the thicker Layer 4 is also characterized by higher density. Assuming that there is some vertical density gradient within this layer we conclude that the densities near its bottom are close to typical mantle densities in the deepest parts. Values as low as 2970 kg m^{-3}, typical values for lower oceanic crustal materials, are almost excluded there. Based on these results we suggest that the seismic Layer 4 is a transitions zone from crustal to mantle material, a mixture of both.

(iv) By assuming that the upper-mantle density variations necessary to compensate the gravity effect of crustal layers are caused by thermal variations we estimate the average depth to the 1200 °C isotherm to be close 45 km. This isotherm seems to be everywhere well below the bottom of seismic Layer 4 except in a narrow zone around the Krafla volcanic centre. We conclude that the temperature within the seismic Layer 4 is close to 650 °C at its top, increasing to approximately 950 °C at its bottom (Moho), which makes a widespread layer of partially molten material within Layer 4 unlikely.

(v) Based on the analysis of horizontal gradients of the isostatic anomalies we conclude that the main tectonic patterns in Iceland have circular forms east of the eastern volcanic zone but are near parallel to the ridge axis on the western side. The eastern circular zones of the horizontal gradient maxima likely reflect folding

structures formed under the additional pressure of the plume, which centre is clearly marked in the obtained map.

(vi) By use of cross spectral analysis of the gravity field and the external topographic load at short wavelengths, we conclude that the elastic plate thickness in Iceland can hardly exceed 6 km and the Icelandic lithosphere cannot support significant long-term stresses elastically.

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