Icelandic-type crust enigma solved

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Introductory paragraph

Although on the mid-Atlantic extensional plate boundary, the crust beneath Iceland is unlike any produced by sea-floor spreading ^{1,2}. It is up to 40 km thick – 6-7 times thicker than typical oceanic crust ^{3,4}, its seismic velocity gradients are inconsistent with a solely mafic petrology, its lower part has a density approaching that of the mantle, and lithosphere thickness is up to 100 km. Furthermore, downward continuation of surface high thermal gradients predict partial melt in the lower crust that is at odds with the subsolidus state detected by seismic measurements. For the last ~ 50 years it has been assumed that its unusually large thickness results from excess magma production in a hot source ⁵ but it cannot explain these observations. We model jointly the seismic and thermal data and show that they are consistent with stretched continental crust ⁶ overlain by a few kilometres of young flood basalts. This resembles a volcanic passive margin under construction. It is mechanically feasible ⁷ and compatible with the composition of Icelandic lavas. This solution is the first proposed that can explain the geophysical properties of Icelandic-type crust. It has major implications for Atlantic Ocean opening, the structure of passive margins, extent of continental crust in the oceans, natural resources and limits of national seabords.

Geological background

Iceland is a 350 x 400-km volcanic island in the central northeast Atlantic Ocean through which the currently active mid-Atlantic extensional plate boundary passes ⁸⁻¹⁰. It forms the subaerial part of a shallow bathymetric ridge that stretches from Greenland to the Faroe Islands (the GIF ridge) and stands ~ 2 km shallower than the surrounding sea floor (Fig. 1).

Figure 1. Map of the Iceland region showing bathymetric contours and tectonic and geological features. Thick black lines: axes of Reykjanes and Kolbeinsey ridges (RR, KR), grey: spreading segments of the main neovolcanic zones, white: glaciers. Black dots: locations of rocks dated at 16 ± 0.3 Myr and 12.92 ± 0.14 Myr in NW and E Iceland ¹¹⁻¹³. Colour (bar at top right) shows crustal thickness. Red contours on land: measured surface conductive heat flow in mW/m^2 ^{14, 15}. Blue dashed lines: limit of Miocene and Lower Pliocene rock outcropping at the surface (i.e. rocks older than 3.3 Myr).

Iceland and the flanking GIF ridge have been extensively studied using many seismic methods. A gravity map is also available ¹⁶. Crustal thickness is generally ~ 30 km and up to a maximum of 42 km in central Iceland ², greatly exceeding normal oceanic crustal thickness of ~ 6 km. The extraordinarily large thickness of Icelandic crust is commonly attributed simply to unusually thick oceanic crust resulting from enhanced melt volumes produced by an exceptionally hot source. However, this model cannot explain multiple features in particular four primary observations, as follows.

First, the positive P-wave velocity gradient throughout the thick crust cannot be explained with mafic oceanic lithologies alone. Second, lithosphere thickness beneath ~ 15 Myr parts of Iceland

is up to 100 ± 20 km¹⁷ in stark contrast to the ~ 37 km expected for oceanic lithosphere of this age. Third, the density of the lower crust, determined from seismic and hypsometry data^{18,19}, is $\sim 3,150$ kg/m³, close to that of the mantle, while its seismic velocity is only 7.2-7.3 km/s⁴. No basaltic petrology can yield this combination of density and seismic velocity^{1,18}. Fourth, shallow geothermal gradients in Iceland predict partially molten material at 10-30 km depth¹⁴. However, this is ruled out by seismic measurements that show the lower crust is subsolidus^{20,21}. Any model for the Icelandic crust must thus feature a lower geothermal gradient at depth.

A new continental model

We test a model for Icelandic crust similar to a volcanic passive margin under construction⁶. In this model, much of the lower crust is stretched, continental mid- and lower crust inflated by magmatism and capped by young eruptives and intrusions that form most of the upper crust.

We assume the continental material was sourced originally from the Greenland and Faroe margins where the pre-breakup (~ 54 Myr) crust was originally 50 km thick and the lithosphere 120 km. Original surface heat flow was 64 mWm^{-2} and mantle heat flow 23 mWm^{-2} , values typical for the continents. We assume that significant thinning and mafic magmatism in the local Iceland area started at about 15 Myr, the age of the oldest surface rocks there¹¹⁻¹³. The crust is heated by magma from the uplifted asthenosphere and stretching-induced shear of the lower, continental part of the crust.

Method

We model the seismic- and thermal data, combining both extension and magmatism. From this we derive candidate thermal and petrological structures. For the volcanically extinct (1-15 Myr) parts of the Icelandic crust, we construct 1D models in which the original continental crust and subcrustal mantle lithosphere are thinned by stretching and the upper crust is inflated by emplacement of sills and dykes (see Supplement 'Methods and Modelling' for further details). Extension occurs as pure shear such that temperatures are determined by geometric stretching and the geothermal gradient increases according to the applied extension factor. Thinning of the lithosphere increases basal heat flow as we keep the lithosphere-asthenosphere boundary at a constant temperature of 1353°C . Once emplaced, the magmatic layers are also extended.

In this model, the original continental crust thins from 50 km to ~ 10 km over 15 Myr ($\beta = 5$ for the continental part). At the same time the mafic upper crust thickens to 10 km. The intrusions and eruptions displace the continental rocks downwards and simultaneously the lithosphere-asthenosphere boundary shallows as a result of the stretching. The resulting total crustal thickness after 15 Myr is 20 km, which agrees with crustal thicknesses in W Iceland. Magma alone is not responsible for the elevated geothermal gradients in the older parts of Iceland. There, our modelling suggests that extension is the most important process maintaining warm conditions. The modelling predicts Moho temperatures of $\sim 700^\circ\text{C}$ beneath 35 km thick crust and $900\text{-}1000^\circ\text{C}$ beneath the thin, 20-25 km thick crust.

For the currently active volcanic extensional zones we use a 2D model. Shallow magmatism is predominantly dyking. In our model we approximate magmatism to vertical dykes that gradually

replace upper crustal material (see Supplement 'Methods and Modelling'). We emplace 1 km wide dykes at 1200°C in the upper crust at 0.2 - 7.7 km. Emplacement occurs every 50 kyr. The dyke material widens and builds the mafic upper crust. Crustal thickness after extension is 36 km, and lithosphere thickness is 86 km. This compares with measurements of crustal thickness of 25 - 40 km beneath the volcanic zones and lithosphere thickness of 80 km in central Iceland ²² (Fig. 1).

The solidus curves of dehydration melting of muscovite, biotite and hornblende are clearly above the model geotherm in practically the whole crust (Fig. 2). The thermal lithosphere-asthenosphere boundary is at 85 km depth where the geotherm meets the solidus curves of H₂O-bearing peridotite. The model suggests that partial melt in adiabatically rising mantle peridotite would not exist shallower than 100-80 km depth, depending on water content. This is in good agreement with the lithosphere thickness estimates from surface waves in central and eastern Iceland ^{17, 22}.

Our modelling produces very high temperature gradients in the upper crust immediately after dyke emplacements, exceeding 200 K/km in the youngest dyke emplaced just 50 kyr ago. With time, heat diffuses out from the dykes and the effects of subsequent dykes are superposed. The shallow (<2 km) temperature gradient in the oldest dykes emplaced at 0.7 Myr is about 60 K/km. In the active volcanic zones magmatic heat is dominant and responsible for the elevated geothermal gradients. These results accord with observations ^{7,14}.

Our modelling shows, critically, that most of the crust beneath ~10 km is not affected thermally by the magmatism that builds the upper crust. This is because magma is transported rapidly up from the asthenosphere along narrow pathways and either stored in shallow magma chambers that are small compared with the whole crust, or erupted. It does not permeate the continental lower crust, which retains inherited low temperatures. This result immediately explains the enigma of the cool, subsolidus Icelandic lower crust detected by seismic observations ^{6, 20, 21}. There, temperatures are controlled by stretching, shearing, and mantle conductive heat flow.

Single rock types cannot explain the positive seismic velocity gradients observed with depth because the seismic velocities of single rock types are predicted to *decrease* with depth under normal geothermal and compaction conditions (Fig 3a) ²⁶. Similarly, in the very thick Icelandic crust the velocity gradients observed can only be explained by variation in lithological composition with depth, particularly decreasing SiO₂ content. Assuming the lower Icelandic crust to be largely continental, we model the observed P-wave velocities with typical continental rock type suites and calculate corresponding density profiles (Figs. 3b and 4) (see also Supplement 'Methods and Modelling').

Figure 2. Model geotherm (thick pink line) applied in lithological modelling of Icelandic crust, assuming extension of continental crust by $\beta = 1.4$ and emplacement of mafic dykes in the upper crust. Geotherm represents temperatures at a dyke emplaced 400 kyr BP in a 2D model. Mantle adiabat ($T_P = 1353^\circ\text{C}$) and selection of solidi curves are for dry peridotite²³ (PER Dry) and wet peridotite²⁴ (PER $<0.4\%$ H_2O). Intermediate values for 0.05 and 0.1% H_2O were interpolated. MS, BT, HB, indicate the dehydration melting temperatures of muscovite, biotite, hornblende. WET or-ab-qz is solidus in the wet orthoclase-albite-quartz system²⁵.

Figure 3. a) Thin lines: P-wave velocities vs. depth calculated for various rock types using the geotherm from Fig. 2. Blue dashed line: generic P-wave velocity vs. depth profile ⁴, b) P-wave velocities modelled using continental rock type mixtures (red line and squares). Boxes give modelled compositions at 5 km depth intervals. Subscripts indicate vol% proportions of each rock type. BGN, biotite (tonalite) gneiss; AMP, amphibolite; MGR, mafic granulite; GGR, mafic garnet granulite; BAS, basalt; DIA, diabase; GAB, gabbro-norite-troctolite; ECL, mafic eclogite; HBL, hornblende; PYX, pyroxenite; HRZ, harzburgite; LHZ, lherzolite.

We ignore the uppermost crustal layers (<5 km) in our modelling, because of their extreme heterogeneity. For the depth range 5-30 km we use felsic and intermediate continental rocks types (BGN, AMP), mafic dykes, and intrusions (DIA and GAB) (Fig. 3b). The mafic component gradually increases downward as we add lower-crustal mafic granulites, mafic garnet granulites and hornblende (MGR, GGR, HBL). At 25-30 km we add small amounts of eclogite (ECL) and the ultramafic rock types pyroxenite, lherzolite and harzburgite (PYX, LHZ, HZB) to reproduce the highest velocities. The uppermost mantle is modeled with lherzolite and harzburgite. Average seismic velocities are calculated assuming mean values weighted by the vol% of each rock type at appropriate pressures and temperatures ^{26, 27}. We estimated *in situ* densities from the P-wave velocity vs. depth profile ⁴ and density-velocity relationships derived from gravity modelling ¹⁹ (see Supplement 'Methods and Modelling').

Lithological modelling (Fig. 3) of the crust is ambiguous, and the assemblage we describe is only an example. The rock types used are, however, consistent with evidence from lower crustal xenoliths and seismic modelling in continental areas ²⁸⁻³⁰. We find that continental rock types can satisfy observed seismic velocities, densities and geothermal gradients observed in Iceland, whereas oceanic petrologies with melt volumes boosted by elevated source temperatures, cannot.

Figure 4. Crustal densities estimated from generic P-wave velocity vs. depth profile for Icelandic crust ^{4, 19} (blue dashed line) and vertical variation of density of the lithological model in Fig. 3b (green line).

Summary & conclusions

Our model of the structure and physical properties of Icelandic crust using stretched continental crust overlain by mafic intrusions and eruptives, easily accounts for the four primary enigmas described above. These enigmas resulted from using a model of unusually thick oceanic crust created by elevated source temperatures ².

The petrological variation required to explain the generally positive velocity gradients is easily reproduced using typical continental petrologies. A lithosphere up to 100 km thick, characteristic of continental lithosphere, is readily accounted for. The densities of the lower crust, uncharacteristic of oceanic layer 3, are easily explained by our continental petrology. Finally, the sub-solidus temperature observed in the lower crust, in the face of downward extrapolation of shallow thermal gradients predicting partial melt as shallow as 10 km, is accounted for because delivery of melt to shallow levels is achieved with minimal heating of the lower crust.

A largely continental composition for Icelandic crust can also account for numerous structural and tectonic anomalies in the Iceland region ⁶. If applied to the entire GIF ridge, which has similar structure, it implies that the Northeast Atlantic ocean at this latitude has not yet progressed to

full sea-floor spreading but continues to extend in the style of passive-margin formation. These findings may also be applied to passive margins around the north Atlantic and worldwide, and have implications for their lithologies and thermal states below drilling depth ³¹.

Our modelling is preliminary and more work needs to be done, in particular to study regional variability in structure. It is particularly urgent to model the petrology and geochemistry of Icelandic basalts to test the idea that they are consistent with MORB melt that has risen through stretched continental crust. Petrological support already exists from Iceland in the form of Precambrian and Mesozoic zircons ³² and Sr and Pb isotope ratios with continental affinity ³³.

The extent of continental crust in the oceans is important to nations for reasons of national sovereignty and natural resources. The present study makes a significant contribution to mapping continental crust in the oceans and is thus important to these wider societal issues ³⁴.

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Supplement: Methods and modelling

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Half-space cooling model. We calculated temperatures, temperature gradients and heat flow values for a spreading ridge (and hot plume) environment using an analytical solution of a cooling half-space^{35, 36}. Heat flow (Q) and temperature gradient on the half-space surface are obtained from:

$$Q(t) = -\lambda \frac{\partial T}{\partial z} = -\lambda \frac{d}{dz} [T_a \operatorname{erf}\left(\frac{z}{2\sqrt{st}}\right)] = -\frac{\lambda T_a}{\sqrt{\pi st}} \quad (1)$$

where t is time, λ is thermal conductivity, $\frac{\partial T}{\partial z}$ is temperature gradient, T_a is asthenosphere temperature, and s is thermal diffusivity of the lithosphere. For normal (present) oceanic lithosphere T_a is 1350°C, and λ is 2 Wm⁻¹K⁻¹. At 100 km depth T_a 1350°C corresponds to a mantle potential temperature of 1300°C.

Thickness of new crust produced by mid-ocean ridge spreading and plume productivity. We estimated the production of new crust in mid-ocean ridge spreading and plume activity by determining the thickness of a melt layer generated by partial melting of upper mantle peridotite³⁷. The results concerning melt-layer thickness and mantle potential temperature were applied. Adiabatic uplift of mantle material with potential temperature of 1300°C produces an ~ 8 km thick layer of basalt which normally forms the ocean crust. A potential temperature of 1530°C would be required to produce 30-km-thick crust, a value comparable to crustal thickness observed in Iceland³⁷. Such a high mantle potential temperature would yield a more ultramafic composition of the produced crust than observed due to the high degree of melting³⁷⁻³⁹. Assuming a normal potential temperature (~1350°C) would require very active upwelling of the mantle to produce such a crustal thickness. Nevertheless, the lower crustal velocity would then be only ~7 km/s³⁹.

Temperatures in the lithosphere during extension and magmatism. The transient thermal conditions of continental lithosphere with magmatic intrusions, dykes and eruptive layers on top were modeled using a numerical finite-difference heat transfer modelling program solving the conductive heat transfer equations⁴⁰. A continental steady-state geotherm was used as the initial condition with 120 km thick lithosphere and 50 km thick crust. The initial geotherm was calculated using the following boundary conditions: A surface constant temperature (5°C) was used, and constant heat flow at the lower boundary (23 mWm⁻²). Thermal parameters and layer thicknesses assumed are given in Table 1.

We modelled heating by lithospheric extension with geometric stretching and elevating the lithosphere-asthenosphere boundary (LAB) correspondingly⁴¹. Magmatic and volcanic inflation of the upper crust was modelled by gradually adding layers and dykes at 1200°C to the continental upper crust during extension.

Using this method we obtained high geothermal gradients in the uppermost crust and lower gradients at deeper levels within the extending continental lithosphere beneath. Present-day crustal thickness in eastern Greenland is about 50 km⁴² and we used this for the pre-breakup crustal thickness. Model parameters are given in Table 1. Geotherms in the neovolcanic zones (<0.7 Myr) were modelled using 2D models, and those in the Tertiary-Pliocene areas with 1D models.

Table 1. Thermal parameters of the initial continental crust, extension models and mafic magmas

Depth	Layer	Conductivity	Heat production	Steady-state surface heat flow	Basal heat flow
km		Wm ⁻¹ K ⁻¹	μWm ⁻³	mWm ⁻²	mWm ⁻²
0-10	Felsic UC	2.5	3	64	23
10-50	Mafic & granulitic lower crust	2.5	0.4		
50 - 120	Upper mantle	2.5	0.001		
0 - 10	Mafic dykes, sills and volcanics	2	0.3		

Geotherms of the Neovolcanic Zone. Our initial steady-state 2D heat transfer model is 120 km deep and 80 km wide. Vertical and horizontal mesh sizes are 100-1000 m and 1000 m, respectively. The uppermost and lowermost 1 km layers have refined 100 m mesh sizes. Extensional thinning of the lithosphere occurs instantaneously at model time 0 Myr. Thinning of the lithosphere is simulated by decreasing the vertical mesh size from the initial value of 1000 m (initial steady-state) to 700 m, corresponding to a beta factor of 1.4. We approximate magmatism by the following simplified model. Vertical 1 km wide dykes with depth extent of 0.2 – 7.7 km are emplaced side by side at 50 kyr intervals so that after 700 kyr a 14 km wide bundle of mafic dykes is produced. This corresponds to a full spreading rate of 2 cm/yr, simulating the production of new crust. After emplacement of each dyke, temperatures equilibrate for 50 kyr before emplacement of the next dyke. Time stepping applied in the transient modelling is 0.02 yr, and the temperatures were stored at intervals of 10 kyr. In the 2D model the dyke material replaces continental rocks and builds a mafic upper crust. Crustal thickness after extension is 36 km, and lithosphere thickness 86 km. Result are shown in Fig S1 and in manuscript Fig. 2.

Figure S1. Modeled geotherms for the neovolcanic zones. Continental lithosphere is thinned instantaneously as per a beta factor of 1.4 from original crust and lithosphere thicknesses of 50 and 120 km, respectively. Subsequently dykes 1 km wide are emplaced at 50 kyr intervals producing a mafic upper crustal layer 14 km wide and 7.7 km thick in 0.7 Myr. Curves represent temperatures at the locations of dykes emplaced in the 2D model. Curve parameter is the age of a dyke in Myr. The insert shows the results on the scale of the whole lithosphere. UC, upper crust; LC, lower crust; M, Moho; LAB, lithosphere-asthenosphere boundary.

Geotherms of the Tertiary – Plio-Pleistocene (0.7-15 Myr) areas. Geotherms in these areas were modeled with 1D heat transfer models. In the 1D models the original continental crustal thickness of 50 km is gradually reduced to 10 km over 15 Myr (Fig. S2). Thinning is partly compensated by mafic sill emplacement in the upper crust at rate of 1100 m/Myr. The emplacement rate was adjusted to match the present thickness of the uppermost (mafic) crust. Once emplaced, the sills also extend in the model. The intrusions displace the continental rocks deeper and the LAB shallows.

Figure S2. Thicknesses of the crust, continental layer and mafic upper crust used in the 1D extension modelling.

The initial steady-state model is 120 km deep, and the mesh is 100 m in the vertical direction (refined to 10 m in the uppermost and lowermost 100 m of the model). During extension the mesh size is reduced according to the beta value at 1 Myr steps. In the final model at 15 Myr (beta 5.0) the model is 24 km deep with mesh size 25.2 m (refined to 2 m in the uppermost and lowermost 20 m). Temperatures equilibrate for 1 Myr between extensional steps and the heat transfer equations are solved at time steps of 0.1 yr. Data are stored at intervals of 100 kyr. The boundary conditions used for the transient models are constant temperature at the surface (5°C) and the bottom boundary (LAB, 1353°C). After each extension step, a 1100-m thick layer of mafic sill is added to the model at an emplacement temperature of 1200°C. Sills are added starting from the surface to increasingly greater depths so that a continuous 10-km thick layer of mafic rock is gradually produced over 15 Myr. Extension occurs as pure shear in the 1D models, i.e., the whole lithosphere is extended.

Ultimately, at 15 Myr the crust comprises 10 km of mafic material underlain by 10 km of extended continental crust (Fig S2). The resulting crustal layer thicknesses correspond to present crustal thickness in the thin-crust areas of Iceland ^{4,17}. Correspondingly, the LAB shallows to reach a final value of 24 km (beta = 5). LAB depths are in agreement with surface-wave measurements ¹⁷ for thin crust and thin lithosphere areas. Final crustal thickness in the model is 20 km, in reasonable agreement with values measured in W Iceland. The thinnest crust in S Iceland is about 20 km.

Modelled temperatures are shown in Fig. S3. Generally, the geothermal gradient increase on the scale of the lithosphere results from extension and shallowing of the asthenosphere depth, while magmatic heat produces high gradients in the upper crust (Fig. S3). The results in Fig S3 represent situations when the emplaced sill layers have cooled for 0.1-0.5 Myr. Magmatic heat diffuses away quickly, and the deeper the sill emplacement is, the smaller becomes the thermal effect at the surface. The surface geothermal gradient is generally 50-80 K/km in >10 Myr crust, but very high (>125 K/km) if sills have been recently emplaced at shallow depth.

Figure S3. Selection of modelled 1D geotherms 0-14.5 Myr after the onset of extension representing the Tertiary and Pliocene-Pleistocene areas of Iceland. The panel at left shows geotherms 0.1 Myr after emplacement of each 1100-m thick layer of mafic melt. The panel at right shows the geotherms 0.5 Myr after emplacement. Curve parameter is the total time in Myr. Curve '0' represents the initial steady-state continental geotherm. The dashed line indicates Moho.

Modelling of seismic velocities and densities. The seismic velocities of solid rocks depend on mineral composition, texture, pressure, temperature, porosity and the properties of the pore fluid. In calculating seismic velocities for different geotherms (pressure-temperature conditions), we use the databases of Christensen and Mooney²⁶ and Hacker and Abers²⁷. Christensen and Mooney²⁶ report laboratory measurements of seismic velocities and densities for 30 different rock types at 5-50 km depths at 5 km depth intervals for low, average and high geothermal gradients. Hacker and Abers²⁷ provided an Excel macro for calculating seismic velocities for given modal compositions, pressures and temperatures. Their database was applied for calculating the velocities and densities of upper mantle lherzolite and harzburgite, while all other rock types

were modeled using the Christensen and Mooney ²⁶ database. Values corresponding to our geotherms at 5 km depth intervals were interpolated. We constructed geologically feasible rock type suites and calculated the average seismic velocities and densities of the mixtures. By varying the volume proportions of the different rock types, we fit our model with experimental V_P and density vs. depth profiles. Averages were calculated as arithmetic means weighted by the vol% of rock types. Porosity was assumed to be zero.

We compared the modeled V_P with the generic V_P profile of Iceland (Fig. 10 in ref 4). Velocity of the sub-Moho mantle (7.85 km/s) was taken from receiver function results ⁴³. Densities corresponding to the V_P profile were calculated using density-velocity relationships compiled for crustal rocks ¹⁹. The density of the sub-Moho mantle (3240 kg/m³) was adopted from gravity modelling results of the ICEMELT transect ¹⁹.

Figures

Figure 1

Map of the Iceland region showing bathymetric contours and tectonic and geological features. Thick black lines: axes of Reykjanes and Kolbeinsey ridges (RR, KR), grey: spreading segments of the main neovolcanic zones, white: glaciers. Black dots: locations of rocks dated at 16 ± 0.3 Myr and 12.92 ± 0.14 Myr in NW and E Iceland 11-13. Colour (bar at top right) shows crustal thickness. Red contours on land: measured surface conductive heat flow in mW/m² 14, 15. Blue dashed lines: limit of Miocene and Lower Pliocene rock outcropping at the surface (i.e. rocks older than 3.3 Myr).

Figure 2

Model geotherm (thick pink line) applied in lithological modelling of Icelandic crust, assuming extension of continental crust by $\beta = 1.4$ and emplacement of mafic dykes in the upper crust. Geotherm represents temperatures at a dyke emplaced 400 kyr BP in a 2D model. Mantle adiabat ($T_P = 1353^\circ\text{C}$) and selection of solidi curves are for dry peridotite 23 (PER Dry) and wet peridotite 24 (PER $<0.4\%$ H₂O). Intermediate values for 0.05 and 0.1% H₂O were interpolated. MS, BT, HB, indicate the dehydration melting temperatures of muscovite, biotite, hornblende. WET or-ab-qz is solidus in the wet orthoclase-albite-quartz system 25.

Figure 3

a) Thin lines: P-wave velocities vs. depth calculated for various rock types using the geotherm from Fig. 2. Blue dashed line: generic P-wave velocity vs. depth profile 4, b) P-wave velocities modelled using continental rock type mixtures (red line and squares). Boxes give modelled compositions at 5 km depth intervals. Subscripts indicate vol% proportions of each rock type. BGN, biotite (tonalite) gneiss; AMP, amphibolite; MGR, mafic granulite; GGR, mafic garnet granulite; BAS, basalt; DIA, diabase; GAB, gabbro-norite-troctolite; ECL, mafic eclogite; HBL, hornblendite; PYX, pyroxenite; HRZ, harzburgite; LHZ, lherzolite.

Figure 4

Crustal densities estimated from generic P-wave velocity vs. depth profile for Icelandic crust 4, 19 (blue dashed line) and vertical variation of density of the lithological model in Fig. 3b (green line).