

Crustal structure of the Reykjanes Ridge near 62°N, on the basis of seismic refraction and gravity data

Wolfgang R. Jacoby^{a,*}, Wilfred Weigel^b, Tanya Fedorova^a

^a Institut f. Geowissenschaften, Johannes Gutenberg-Universität Mainz, D-55099 Mainz, Germany

^b Berliner Str. 71, D-21244 Buchholz, Germany

Received 9 July 2005; received in revised form 23 July 2006; accepted 5 September 2006

Abstract

Explosion deep seismic sounding data sections of high quality had been obtained with RV Meteor in the Reykjanes Iceland Seismic Project (RRISP77 [Angenheister, G., Gebrande, H., Miller, H., Goldflam, P., Weigel, W., Jacoby, W.R., Pálmason, G., Björnsson, S., Einarsson, P., Pavlenkova, N.I., Zverev, S., Litvinenko, I.V., Loncarecic, B., Solomon, S., 1980. Reykjanes Ridge Iceland Seismic Experiment (RRISP 77). *J. Geophys.* 47, 228–238]) which close an information gap near 62°N. Preliminary results were presented by Weigel [Weigel, W., 1980. Aufbau des Reykjanes Rückens nach refraktionsseismischen Messungen. In: Weigel, W. (Ed.), *Reykjanes Rücken, Island, Norwegischer Kontinentalrand. Abschlusskolloquium, Hamburg zur Meteor-Expedition*, vol. 45. DFG, Bonn, pp. 53–61], and here we report on the data and results of interpretation. Clear refracted phases to 90 km distance permit crustal and uppermost mantle structure to be modelled by ray tracing. The apparent P-wave velocities are around 4.5, 6–6.5, 7–7.6 and 8.2–8.7 km/s, but no wide-angle reflections have been clearly seen. Accompanying sparker reflection data reveal thin sediment ponds in the axial zone and up to 400 m thick sediments at 10 Ma crustal age. Ray tracing reveals the following model below the sediments: (1) a distinct, 1–2 km thick upper crust (layer 2A) with V_p increasing with age (to 10 Ma) from <3.4 to 4.9 km/s and with a vertical gradient of 0.1–0.2 km/s/km, (2) a lower crust or layer 3 beginning at depths of 2 (axis) to 4 km (10 Ma age) below sea level with 6.1–6.8 km/s and similar vertical gradients as above, (3) the lower crust bottoms at 5.2–9.5 km depth below sea level (0–10 Ma) with a marked discontinuity, underneath which (4) V_p rises from about 7.5–7.8 km/s (0–10 Ma) with a positive vertical gradient of, again, 0.1–0.2 km/s/km such that 8 km/s would be reached at 12 km and deeper near the axis. Our preferred interpretation is that the mantle begins at the distinct discontinuity (“Moho”), but a deeper “Moho” of V_p ≈ 8 km/s cannot be excluded. From Iceland southward to 60°N several experiments show a decrease of crustal thickness from 14 to 8 km. Velocity trends with age across the ridge reflect cooling and filling of cracks, and thickness trends probably suggest volcanic productivity variations as previously suggested.

Gravity inversion concentrates on a profile across the ridge with the above seismic a priori information; with 0.2–0.5 km depth uncertainty it leads to a good fit (±2.5 mGal where seismic data exist). Best fitting densities are (in kg/m³) for sediments, 2180; upper crust, 2450–2570; lower crust, 2850–2940; mantle lithosphere, 3215–3240 with a deficit for an asthenospheric wedge of no more than –100 kg/m³. The morphological ridges and troughs superimposed on the SE ridge flank are partly correlated, partly anti-correlated with the Bouguer anomaly and suggest that variable crustal density variations accompany the morphology variations. © 2006 Elsevier Ltd. All rights reserved.

Keywords: Reykjanes Ridge; North Atlantic; Iceland; Crust–mantle; Seismic refraction; Gravity inversion

* Corresponding author. Tel.: +49 6131 39 23170.

E-mail address: jacoby@uni-mainz.de (W.R. Jacoby).

1. Introduction

Reykjanes Ridge SW of Iceland is unique, a slow ridge without a rift valley. This probably results from hot asthenospheric SW-ward flow from the Iceland plume (Vogt, 1974). Studying the crust–mantle structural consequences is a challenge and was the motivation for marine geophysical experiments with RV Meteor. In 1977 the large-scale land–sea seismic experiment RRISP77 (Angenheister et al., 1979, 1980; Weigel, 1980) included three seismic lines on Reykjanes Ridge at about 62°N which so far are not yet published in any detail. They do, however, deserve publication in view of their good quality and because they complement other crustal studies to the SW and NE. Additional motivation is the comparison to Iceland itself in one special issue concentrating on Iceland and its immediate surroundings. Hence, the plume-influenced northern part of the Reykjanes Ridge is the main subject, and studies extending further south beyond the plume influence (transition to more normal Mid-Atlantic Ridge at about 56°45'N; Appelgate and Shor, 1994) are considered for comparison.

The seismic data and their interpretation are backed up and substantiated by a gravity inversion. It has the potential of constraining density and temperature and thus highlighting dynamic processes of mantle plume flow and crustal formation, as shown, e.g. by Peirce et al. (2005) for a profile extending along the Reykjanes Ridge axis from ~57°N to ~62°N near the present study area. Though principally ambiguous, the combination with seismic a priori information in gravity inversion sheds light on the interpretation of the seismic refraction and wide-angle reflection observations.

The existing seismic lines between about 59°N and 63°N are shown in Fig. 1a. Profiles 3, 4 and 5 were observed with ~1.5 km shot spacing, which for technical reasons is less dense than in more recent airgun work, but the records show excellent first arrivals (see below; Fig. 2), and models can be constructed from these data with confidence. Profile lengths are 130 km (profile 3), 150 km (profile 4) and 90 km (profile 5). Two of the profiles (4 and 5) intersect the morphological ridge axis at right angles and one (3) is nearly parallel (~6°) to it, <30 km to the SE from the axis. The morphological axis is about 30° oblique to the spreading normal, evident in fissure eruptions and narrow volcanic ridges (Jacoby, 1980; Appelgate and Shor, 1994; Searle et al., 1998), the seismic profiles are oblique to them. The sub-parallel profile 3 follows essentially the edge or slope of the central horst structure of Reykjanes Ridge at about 2.5 Ma seafloor age. The horst structure narrows SW-ward and was interpreted by Vogt (1974) to be part of a V-shaped ridge associated with a plume pulse and the initiation of the SE volcanic zone in Iceland.

Profiles 3–5 were part of the large-scale RRISP77 experiment that concentrated on the 640 km long profile 1 (see below), parallel to profile 3, extending with big, 1–4 t, shots from about 60°40'N, 26°W along the SE flank of Reykjanes Ridge and across Iceland to the NE coast till 65°45'N and 14°50'W. The present results should be seen in the context of the whole experiment.

Published data along the Reykjanes Ridge between 60°N and Iceland include several deep seismic sounding experiments which image crust and uppermost mantle structure (Bunch, 1980; Goldflam et al., 1980; Ritzert and Jacoby, 1985; Smallwood and White, 1998; Weir et al., 2001). At about 60°N three ridge-parallel, 100 km long profiles were acquired by Bunch (1980) called here B1, B2 and B3, at 0, 3 and 9 Ma crustal age, respectively (Fig. 1). Between 61°20' and 62°N Smallwood and White (1998) acquired three ridge-normal and two ridge-parallel profiles, <100 km

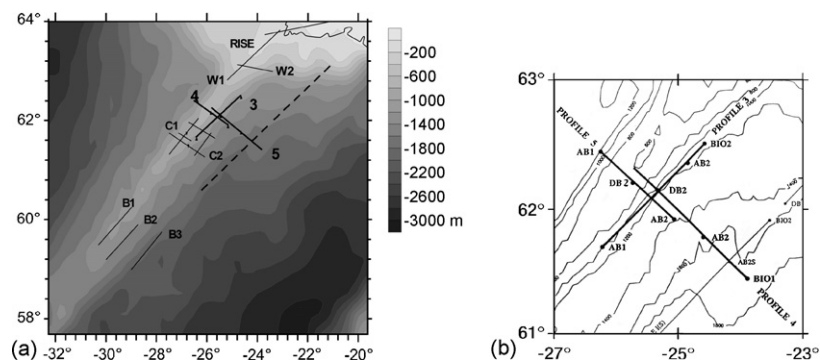


Fig. 1. Location map of seismic lines discussed in this paper. (a) Overview of the various experiments in the study region, bathymetric contours annotated for reference. (b) Details of RRISP profiles plotted on bathymetry contours annotated in m; thin line, profile 1.

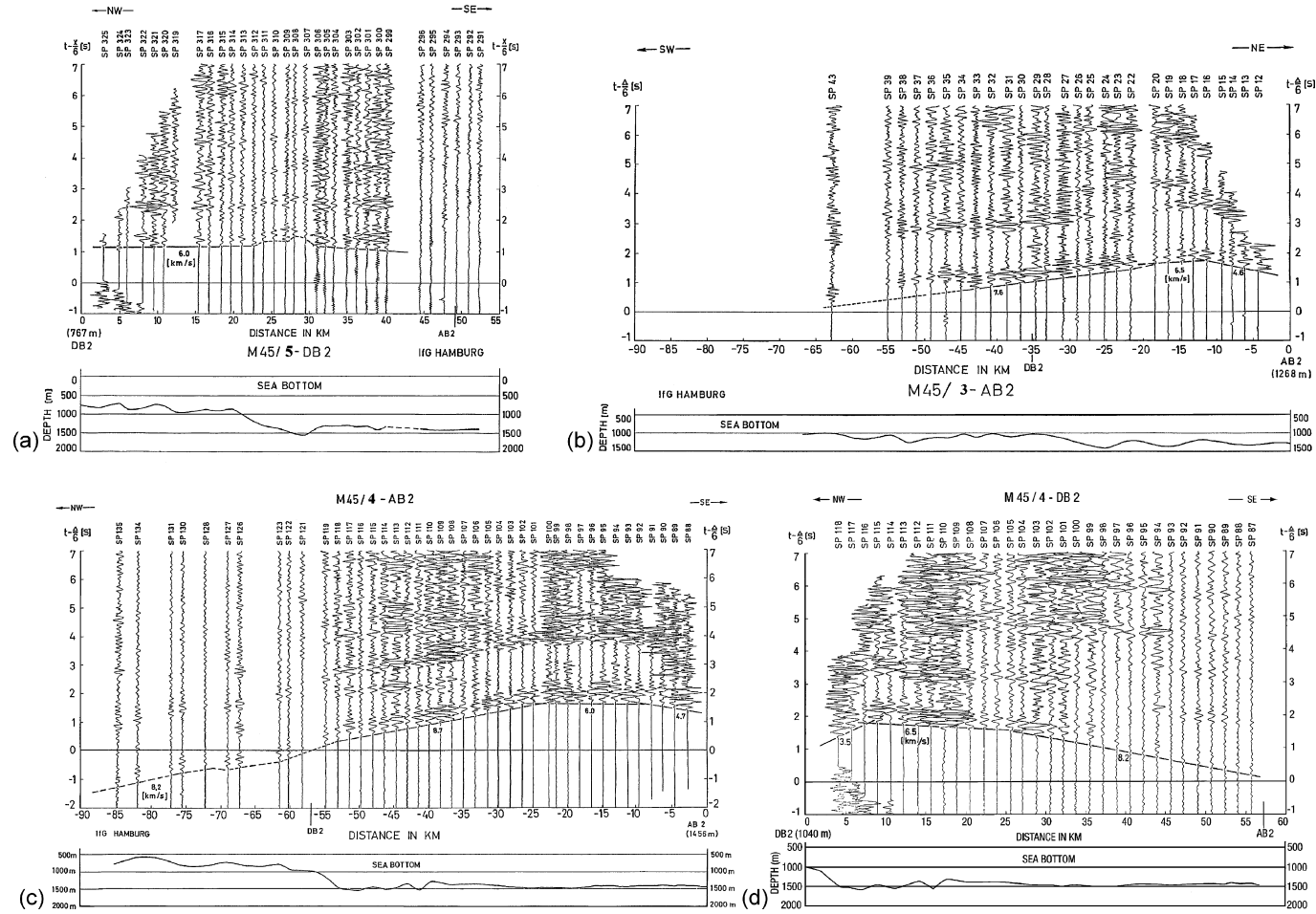


Fig. 2. RRISP record sections by codes “profile-instrument” with recording direction indicated; profile 3 is parallel to the ridge axis, profiles 4 and 5 are across the ridge: (a) 5-DB2, (b) 3-AB2, (c) 4-AB2 and (d) 4-DB2; for positions of the stations see Fig. 1. Plotted at bottom is bathymetry; the sections (a) 5-DB2 and (d) 4-DB2 begin on the axial horst or its edge, while section (c) 4-AB2 ends on the horst.

long, called here C1 and C2 at 0 and 5 Ma crustal age, respectively. To the NE from the present study area Weir et al. (2001) acquired an axial profile W1 between ~ 63 and $\sim 64^\circ\text{N}$, continuing at an angle on the Reykjanes Peninsula, and an 80 km long, approximately EW cross-profile W2 near 63°N extending to ~ 6 Ma crustal age. RRISP77 profile 1 (Goldflam et al., 1980; Ritzert and Jacoby, 1985) follows the eastern ridge flank at a small angle along a V-shaped ridge, about 10–12 Ma old from $61^\circ 20'$ to the Iceland shelf at 63°N .

Bunch (1980) observed a refractor at >7 km depth below sea level (under 1 km deep water) at the axis, deepening to about 11.5 km below sea level for 9 Ma old seafloor (2 km water depth). The Moho is represented by small velocity steps or steep gradient zones of 0.66 s^{-1} ($=\text{km/s/km}$). The Pn velocity increases from 7.1 to 8.2 km/s. Wide-angle reflections are not evident. At 61°N along C1 and C2 at 0 and 5 Ma age (Smallwood and White, 1998), wide-angle reflections are interpreted to represent the Moho with a velocity contrast of 0.5–0.7 km/s at about 10 and >8 km depth below seafloor, respectively (11 and 10 km below sea level); thus their crustal thickness decreases with age in contrast to that of Bunch (1980). The crust consists essentially of two gradient zones from 3 to 6 km/s and increasing to 7–7.2 km/s. The transition occurs at 3.5 ± 0.5 km depth below seafloor (see Fig. 5). Weir et al. (2001) present similar axial models for their RISE profile from $62^\circ 50'$ northward into Reykjanes Peninsula at 64°N , however with an overall crustal thicknesses increasing from 10 (13 km at axis) to 21 km into Iceland; here a reflector is seen with $V_p \geq 7.5$ km/s at 9–11 km depth at the tip of Reykjanes Peninsula. Across the ridge near 63°N , crustal thickness appears to be greatest near the axis (>13 km, see below; Fig. 5) thinning to 10 km at 5 Ma crustal age. Weir et al. (2001) and Smallwood and White (1998) explain the crustal thickness variations by temporal plume temperature variation. In this context, the morphology and gravity anomalies of the ridge must also be taken into account (see below).

At the ~ 10 Ma old SE flank, along RRISP77 profile 1 (Ritzert and Jacoby, 1985) water depth increases away from Iceland from 1 to 2 km. Covered by a few hundred metres of sediments, the upper crustal layer thins from >3 to 2 km thickness and has a vertical P-wave velocity gradient of $>0.2\text{ s}^{-1}$ (4.3 to ~ 5 km/s). The lower crust is ~ 4 km thick, its bottom rises SE-ward from 11 to 10 km depth; V_p rises with depth from 6.3 ± 0.1 to ~ 7.3 km/s with indications for a low-velocity layer at the bottom. The travel times constrain the Moho depths to about ± 0.5 km in a trade-off with the crustal velocities. At least such an uncertainty exists in all results discussed here and must be kept in mind when comparing them. The uppermost mantle velocity is almost 8 km/s, nearly constant with depth until another transition occurs at about 20 km depth to 8.5 km/s sub-parallel to the ridge axis. Very high P-wave velocities are supported by recordings at Icelandic land stations of RRISP shots along Reykjanes Ridge (Einarsson, 1979) showing an average velocity of 8.7 ± 0.3 km/s, which cannot be explained as an upslope apparent velocity. A high velocity parallel to the ridge axis is somewhat surprising, but without more data, it is only a speculation that velocity anisotropy is the reason with the high velocity parallel to the ridge axis. It might be generated by horizontal axial outflow from the plume below 20 km depth; if such a flow field dominates the vertical upflow related to plate divergence, it might explain a preferred olivine crystal orientation opposite to the usual situation (Ritzert and Jacoby, 1985).

Further south, the plume influence tapers out, the central horst is replaced by a median valley south of $\sim 57^\circ\text{N}$. Multidisciplinary geophysical studies (wide-angle seismic reflection–refraction, electromagnetic sounding and magnetotellurics, gravity, magnetics) between 57° and $>58^\circ\text{N}$ (Sinha et al., 1998; Peirce et al., 2005) show that under an axial water depth of 1.5–2 km, the upper crust is ~ 2.5 km thick, and the lower crust (6.5–7 km/s) is ~ 5 km thick or its base at ~ 9 km below sea level (bsl). Subcrustal P-wave velocities are calculated to be 7.9–8.2 km/s, rising to 8.4 km/s at 50 km depth. In addition, combination of seismic velocities and electrical resistivities suggested a recent intrusive event showing the magmatic activity at the ridge to be probably episodic (as it is in Iceland; see e.g. Björnsson, 1985).

Comparison with the studies further north (above) reveals a decrease of the “reflection Moho” depth from ~ 21 km bsl near the tip of Reykjanes Peninsula to ~ 9 km at $\sim 58^\circ\text{N}$. The two studies that did not unambiguously observe the wide-angle reflection, termed PmP (Bunch, 1980 and the present work, see below), obtained “refraction Moho” depths of >7 and 6–7 km, respectively. A weak trend of increasing V_p values may be recognized from ~ 7.8 km/s near Iceland to >8 km/s near $\sim 57^\circ\text{N}$. The gravity interpretation of Peirce et al. (2005) suggested a continuous southward thinning of the crust of approximately 10–7 km from the region of the present study to $\sim 57^\circ\text{N}$ and an accompanying decrease of the uppermost mantle density of 3230–3000 kg/m^3 . Gravity is thus important for the question of what crust and Moho mean in the vicinity of the Iceland plume.

After a description of our data collection, record sections are shown and an outline of the interpretation methods is given. The resulting models are then presented and discussed. The interpretation includes a gravity inversion after an explanation of the method. A geodynamic discussion closes the paper.

2. The seismic experiment

2.1. Data acquisition and treatment

Along the profiles mostly small shots (25–100 kg dynamite) were fired every ~ 1.5 km and recorded by three to four receivers (thus receivers and sources were reversed from the usual procedure with few shots and many receivers along profiles). Ocean bottom hydrophones (OBH: AB1, AB2 and DB2) and two ocean bottom seismometers (OBS: BIO1 and BIO2) were deployed; the latter did not supply many usable data from the small shots for the present profiles (in contrast to profile 1; see above). AB and DB stand for analogue and digital buoys, respectively; the numbers identify the instruments, not the positions; the three instruments were moved during the experiment and a unique identification of the records is by profile and instrument number, e.g. 5-DB2. Fig. 1b shows the instrument positions during the experiment. Mounted on frames with buoyant chambers and removable anchors, the OBHs were deployed, connected via cable to buoys from which the signals were radioed directly to the ship and recorded on magnetic tape and paper. The analogue records on magnetic tape were digitized onboard or back home. The systems had a nearly flat response from ~ 10 to 25 Hz with the 3 dB point near 7 Hz, and the digitized signals had a dynamic range of 52 dB and a sampling rate of 100 Hz; amplifier adjustment was remotely controlled.

The shots were fired from the ship. The dominant signal excitation was about 5 Hz of the bubble pulse; this is at the lower end of the hydrophone frequency range. Combining the shot and receiver frequency bands demonstrates that at the low end, say at 3 Hz (important in some wide-angle reflections), the system response is reduced; on the other hand, 5 Hz are recorded well, as seen in the record sections (Fig. 2, below). Charges were dropped into the water and, for safety reasons, a minimum distance had to be reached by the ship before they were ignited electrically by a shooting machine using a clock signal with 10^{-9} s error. The charges sink at a rate of about 0.7 m/s and, at shooting time, reach an optimized depth if possible (25 kg, 40–50 m; 50 kg, 70 m; 100 kg, 80 m), but a compromise had to be maintained with ship safety. A hydrophone was pulled near the ship; the electrical trigger impulse and the hydrophone signal were recorded for timing, shot distance from the ship and shot depth control (from the time difference of the direct wave and the surface reflection).

The ship moved with a constant speed (usually between 6.5 and 7 kn corresponding to 12–13 km/h). The accuracy of shot coordinates relative to the ship was estimated to be a few meters, and depth was continuously measured by an echo sounder. The onboard integrated navigation system INDAS IV, based on LORAN C (Chain SL7W-SL7X), automatically updated by repeated satellite fixes (before the GPS era). Buoy (Table 1) and shot positions were taken at instrument deployment at the time of water contact, but lateral drift during sinking may introduce some errors, especially in the case of deep water deployments. The travel times of the direct water waves measured on the seismic records permitted the source–receiver range to be checked; the agreement turned out to be within ± 200 m.

Reflection seismic acquisition was carried out in parallel using a single-channel hydrophone streamer spread of 50 m in length. The signals were generated by a 4 kJ sparker source. From the sections, the sediment thickness was calculated with an assumed P-wave velocity of 1.7 km/s. The results are included in the refraction modelling.

The onboard Askania sea gravimeters (instruments Gss3 Nos. 1 and 55, mounted on a gyro-stabilized platform close to the ship's centre of gravity) was operated when circumstances (constant speed and course, no maneuvers)

Table 1
A priori velocities V_p , densities ρ and density contrasts $\Delta\rho$ and a posteriori densities

Layer	A priori			A posteriori	
	V_p [km/s]	ρ_i [kg/m ³]	$\Delta\rho_{i,i-1}$	$\rho_i - \delta\rho$ (in steps)	$\rho_i - \delta\rho$ (in steps) [kg/m ³]
Sed	1.7	2200	$-400 \pm 50/2600$	2200	2200
UC	3.4–4.9	2600	$+400 \pm 50/\text{sed}$	2490–2600	2480–2605
LC	6.2–7.2	3000	$+400 \pm 50/\text{UC}$	2930–3000	2940–3000
UM	7.5–7.8	3250	$+250 \pm 50/\text{LC}$	3190–3250	3220–3280
A		3250	$(-30, -70, -100) \pm 50/\text{L}$	3150–3220	3170–3240

Sed, sediments; UC, upper crust; LC, lower crust; UM, upper mantle; A, asthenosphere; L, lithosphere. In $\Delta\rho_{i,i-1}$ i and $i-1$ stand for the symbols in column 1 (density contrasts between layers); $\rho_i - \delta\rho$ (in steps) describes the range of densities within layers UC, LC, UM and A.

allowed. The data were processed and archived at the Hamburg institute as well as distributed to marine data centres; they contributed to the database used in this paper (see below: bathymetric and gravity data).

2.2. Record sections

OBHs generally (if not at a profile end) received shots from both sides, as the ship passed over the station letting off shots at intervals. Record sections for each OBH were plotted with travel times reduced at $v_{\text{red}} = 6$ km/s. Four of the sections are shown in Fig. 2a–d, identified as explained above: 5-DB2 (a), 3-AB2 (b), 4-AB2 (c) and 4-DB2 (d), each with the bathymetry plotted underneath. The direct water wave has been cut off. Sections that are shorter and had more data gaps are not shown. The dominant frequency content of the traces usually peaks at 6–8 Hz with individual non-systematic scatter. There may be a tendency of a very slight frequency decrease with distance. The amplitudes are plotted normalized with shot size and $1/r^2$ (r = distance) and band-pass filtered at 2–20 Hz. Mainly the relative amplitudes of different phases along the records can be distinguished. Most traces show little noise and clear onsets for each shot. No principal difference can be seen between digital and analogue recording.

Apparent velocities were determined from the first arrivals by adopting a 1D slope-intercept method; they range between 4.6 and 8.7 km/s; the latter high value occurs along a cross-ridge-uphill section, actually in the axial region. The correlations chosen are supported by several bottom multiple phases. The properties of the seismic “discontinuities” are not very obvious from the record sections.

The individual records start with stronger and weaker wave groups varying with shot distance at intervals which differ from section to section; no general and simple amplitude-distance correlation is evident, nor can any amplitude relationships be recognized with location relative to the ridge axis. Energy transmission across the ridge axis seems normal with no significant absorption, as ray tracing (Fig. 3) shows; however, only shallow rays (<10 km depth) have been observed in the axial region. The sections shown do not traverse the axial region; 5-DB2 begins near the axis, however on the same side as the records to the SE; 4-AB2 ends in the axial zone and shows there no distinct attenuation. Clear first arrivals are visible generally to 80 km distance, and mostly a stronger wave group follows the arrival within about 0.5 s.

There is little evidence for wide-angle PmP reflections as seen by other studies (Smallwood and White, 1998; Weir et al., 2001). The question arises whether they were overlooked. The reduced sensitivity of our OBHs at the low frequency end (see above) is not favorable for recognizing the phase. Comparison suggests moreover, that most of our records of Fig. 2, especially along profile 5 may be too short to clearly recognize the phase; weak indications can be seen as a wave group following the first arrivals which we interpret as Pn, at distances beyond 40 km in the sections 3-AB2, 4-AB2 and perhaps 4-DB2, but its delay after Pn, about 0.5 s, should decrease with distance, if it is the same PmP as that identified by Weir et al. (2001); this may possibly be the case in section 3-AB2. In the present records, it was not considered a significant or convincing separate phase, but its hypothetical existence is taken into account in gravity modeling (see below).

2.3. Interpretation—methods, evaluation and modeling

The interpretation proceeded in three steps. (1) Horizontal constant velocity layers were calculated for each travel time diagram (mostly one to both sides of an OBH) on the basis of the apparent velocities and the corresponding intercept times; the sediment thickness was taken from the reflection sections with an assumed $V_p = 1.7$ km/s. (2) Construction of 2D velocity sections was carried out by connecting and interpolating the nearest 1D velocity–depth models. (3) Inversion of the combined set of source–receiver travel time sets on the basis of the initial section of step (2).

In step (1), beside the visual straight-line correlation, the onsets were picked with an error estimated to 10–30 ms. The best-fitting segment apparent velocities and intercept times were calculated by least-squares regression where the formal standard errors came out to about ± 20 ms.

In step (2) the individual or “local” 1D $V_p(z)$ models were interpolated linearly, and their connection resulted in inclined boundaries and velocity gradients between the layer boundaries or discontinuities. The velocity gradients, both vertical and horizontal, were moderate (except in the thin “boundary layers”, see below). The depths estimated for the sedimentary cover from the reflection profiles were included in the models as “fixed” a priori information.

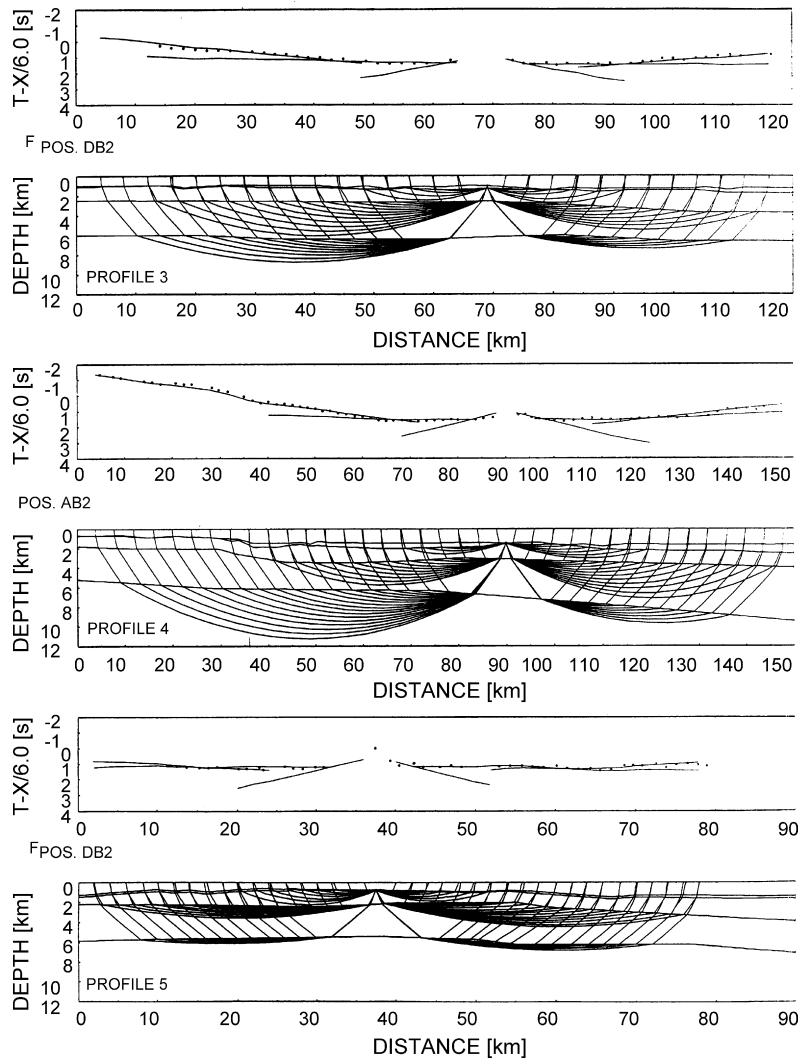


Fig. 3. Examples of ray tracing along the three profiles for the shot arrays recorded by ocean bottom receivers as identified by profile and receiver numbers. Note that $t - t/6$ is plotted downward, as usual in reflection sections.

In step (3) the method of Zelt and Smith (1992) was used for a non-linear iterative inversion for optimized 2D velocity sections. The layers are defined by their upper and lower boundaries of variable depth discretized at ~ 0.2 km intervals and by internal velocity gradients described by 2D polynomials (in x and z); the boundaries themselves are parameterized as thin (0.1 km thick) gradient layers connecting the bottom velocities of the upper layer to the top velocities of the lower layer. For the non-linear inversion of all observed travel times within their error bounds, the initial assumptions were the parameters defining the layers and the velocities, given the freedom of being adjusted and optimized. Travel times were then calculated by applying ray tracing according to Cerveny et al. (1984) to the initial models and compared with the observations. Fig. 3 shows three examples of ray tracing through the final models (3, DB2; 4, AB2; 5, DB2) and the travel times (reduced at 6 km/s, plotted downward as in reflection sections). Compare to record sections 4-AB2 and 5-DB2 in Fig. 2.

As indicated by Figs. 2 and 3, refracted phases to 85 km distance permit crustal and uppermost mantle structure to be satisfactorily modelled. With the inversion routine, velocity models were optimized within assumed limits. The models do, of course, reflect the assumptions including discontinuities delimiting the layers which themselves are permitted to have a smooth 2D velocity variation, i.e. space-variable gradients. Incidentally, in these optimized models observed

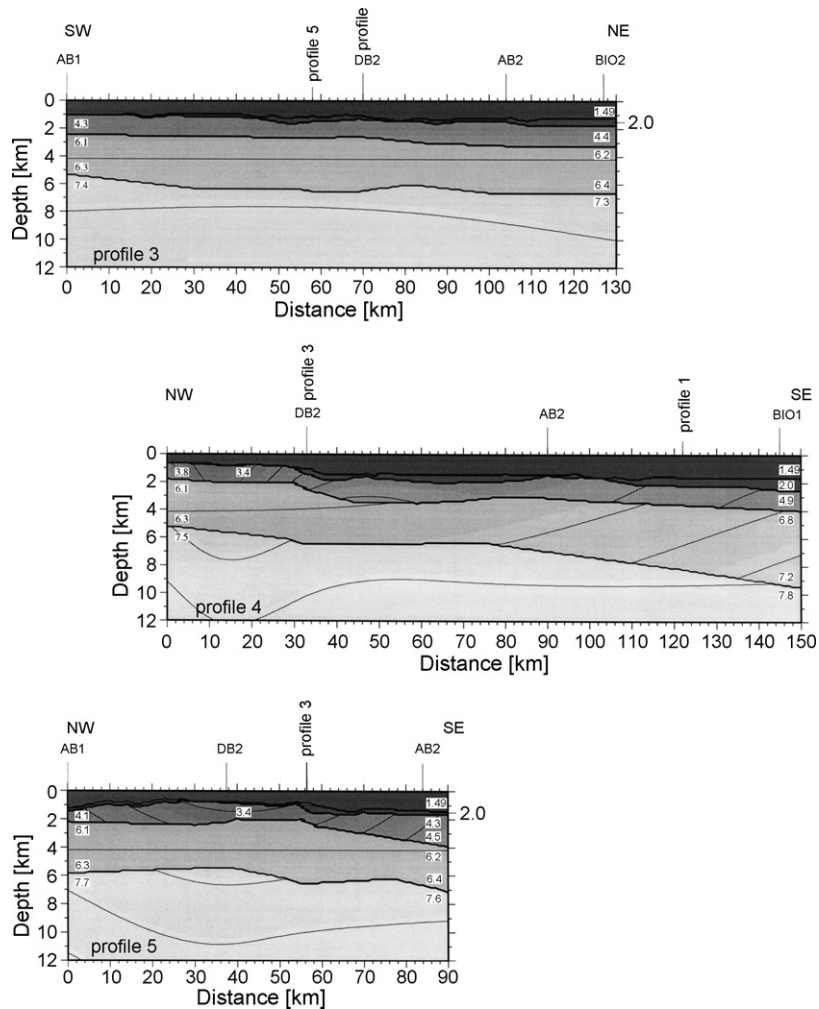


Fig. 4. Velocity sections calculated for profiles 3, 4 and 5 contoured and shaded in grey (smallest velocities are darkest); mutual intersections with the other profiles are indicated. The lowest boundary shown is interpreted preferentially as the Moho. Note that the sections are extrapolated at depth beyond the seismic ray coverage to the sides.

rays penetrate a maximum depth of approximately 12 km (Fig. 3), which is the limit of information obtained; the ends of the profiles are also not covered by seismic rays and at depth the models are extrapolated some 10 km.

2.4. Results of seismic travel time inversion: the sections

The results are presented in Fig. 4 for the three profiles 3, 4 and 5. First, profiles 4 and 5 across the axis (middle and bottom in the figure) are described. Profile 4 begins close to the axis, while profile 5 intersects it at about $x = 30$ km. The sparker reflection data indicate thin sediment ponds in the axial zone, reaching 750 m thickness at 75 km distance from the axis and slightly decreasing again further away to 400 m thickness at 10 Ma crustal age (Goldflam et al., 1980); the sediments are traditionally called layer 1. The layer below is called the upper crust (or layer 2A), and the next is called lower crust (or layer 3). The following “layer” appears bottomless and is called mantle. All sections are characterized by layers of moderate vertical velocity gradients separated by distinct velocity contrasts.

- (1) The distinct upper crust is 1–2 km thick and has a P-wave velocity increasing at the layer top from <3.4 to 4.9 km/s with crustal age (0–10 Ma); these velocities correspond approximately to those generally found for layer 2A (Grevemeyer and Weigel, 1996) or are slightly higher; the vertical gradient of V_p is about 0.2 s^{-1} (km/s/km)

- indicating diagenesis as well as crack and vesicle filling by minerals (Flóvenz, 1980). Around $x = 10$ km on profile 4 and around $x = 30$ – 45 km on profile 5, a pronounced velocity minimum is indicated by the 3.4 km/s contour; the minima in both profiles are offset to the SE from the morphological ridge axis by about 10 km.
- (2) The lower crust begins at depths of 2–4 km and bottoms at 5.2–9.5 km depth below sea level (0–10 Ma) and is thus ~ 3 km thick near the axis and fairly variable and increasing to ~ 5 km at 10 Ma crustal age at the SE flank; the P-wave velocity increases across the upper/lower crust boundary by nearly 2 km/s, modeled as a very thin high-gradient layer, which would be indistinguishable from a transition across < 1 km (typical wavelengths < 1 km); at the top V_p ranges from 6.1 to 7 km/s; the vertical gradients are similar to those in the upper crust; the layer is thought to consist of solid basalt with increasing dyke frequency fostering rapid alteration and solidification. No axial velocity minimum is recognized in this layer.
 - (3) The transition from the lower crust to the deepest layer of the models is again a marked discontinuity of > 1 km/s. At the top, V_p is 7.5–7.9 km/s (0–10 Ma); it has a positive gradient of $< 0.1 \text{ s}^{-1}$, and the velocity contours are depressed near the ridge axis describing a 30–50 km wide low velocity region; again, as in the upper crust, the minimum is shifted 10 km SE from the morphological axis; in the case of profile 4, there are no seismic rays defining the NW flank of the minimum, such that the velocity contours are here less reliable. Near the minimum $V_p = 7.8$ km/s is reached at 11–12 km depth, and extrapolation to 8 km/s would give 14–15 km depth, if that has any significance. Farther from the axis 8 km/s may be encountered at shallower 11–12 km depth.

The two cross-profiles 4 and 5 are quite similar, both in velocity and geometry. In the upper crustal layer the velocities are strongly reduced near the axis, while the effect in the lower crust is weaker and broader and not seen in profile 5. The axial velocity depression below the Moho is about 0.2 km/s. A distinct 1 km downward step of the volcanic basement away from the axis occurs in the distance range 16–25 km off-the axis, corresponding to 1.5–2.5 Ma crustal age. Below the step the sediments have a thickness maximum. The upper crust is generally 1–1.5, locally to 2 km thick, with a minimum (< 1 km) at the step, while the lower crust has a gentle thickness maximum, with the Moho at 6–6.5 km depth bsl. Beyond about 70 km (~ 7 Ma) from the axis the lower crust gently thickens, but here the data are less certain.

Profile 3 (Fig. 4, top) was shot along the ridge flank at about 2.5 Ma crustal age sub-parallel to the axis. Sediment thickness, on the basis of the sparker reflection data, increases towards NE from near 0 to ~ 500 m; this partly reflects the fact that the profile follows the horst flank and actually descends northwestward down the slope by some 200 m (Fig. 1b); partly it is related to the approach towards Iceland. Crustal thickness shows less variation than the profiles across, but some irregularities are visible, the most obvious being the slight deepening and thickening of the layers toward Iceland. The upper crust is fairly constant, the lower crust thickens by about 0.5 km. The Moho depth increases in a somewhat irregular fashion from 5.2 to 6.7 km, and in the middle section a slightly greater Moho depth correlates with enhanced upper mantle P-wave velocities. The velocity gradients are less obvious than in the cross-profiles, and the contours are subhorizontal (in contrast to profiles 4 and 5).

At the two intersections of the three profiles (marked in Fig. 4) velocity–depth functions have been constructed, and the layer boundaries agree in depth very well, with a scatter of < 100 m. The upper crust P-wave velocity is slightly smaller across the ridge than along (~ 0.5 km/s), which may be real, related to the volcanic fissures. The velocity discrepancy is very small in the lower crust. In the uppermost mantle both cross-profiles show a P-wave velocity about 0.1 km/s higher than along which may or may not be real. Quantification of the errors of the a posteriori models is not easy as they involve discontinuities and continuous velocity variations, but for the estimation of the error bars in the gravity inversion (see below) the depth and velocity errors are important. Although the modeling does not permit the errors to be determined quantitatively and definitively, the above discussion suggests 0.2 km for the upper/lower crust boundary and 0.3–0.5 km for the Moho, and for velocity generally better than 0.3 km/s, where systematic differences between across and along velocities are neglected.

In Fig. 5 the velocity–depth functions near the ridge axis are plotted for profiles 3 and 4 together with those from Smallwood and White (1998) and Weir et al. (2001). Obviously there is no gross difference in the velocities at any depth down to 12 km; the curves “meander” within a “band” of 0.8 ± 0.1 km/s width. Different are the magnitudes and depths of the velocity discontinuities and the velocity–depth gradients. In the depth range from 1 to 4 km (from sea level) the present interpretation has a significant discontinuity at about 2.5 km instead of a continuous function in the other studies (gradient 0.5 – 1.0 s^{-1}) and a > 1 km/s discontinuity at about 6.5 km depth instead of a weaker gradient (0.1 – 0.2 s^{-1}) down to a 0.5 km/s discontinuity at depths between 9 and 14 km. Below the 6.5 km discontinuity, the present models have velocity gradients $< 0.1 \text{ s}^{-1}$. Waves traveling down to 6.5 km depth and up again, have very similar travel times in

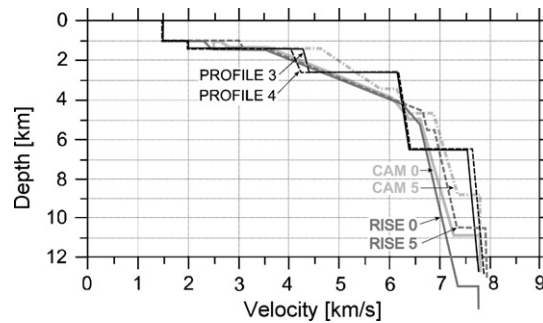


Fig. 5. Velocity–depth functions of three experiments, all referenced to $z = 1$ km at the sea floor which is approximately the axial value; the off-axis water depths are between 1.2 and 2 km. The present result (profile 3 along, profile 4 across the ridge) are compared with results CAM0 and CAM5 (Smallwood and White, 1998) as well as with RISE0 and RISE5 (Weir et al., 2001) where 0 and 5 refer to crustal age in Ma.

both model types, while waves penetrating into depths to about 10 km have distinctly shorter travel times amounting to about 0.5 s across a 50 km distance, indicating real, data-based model differences. In the more continuous models it is essentially only a change in velocity gradient characterizing the crustal layers, and it is not intuitively obvious to which extent the magnitudes of the gradients and the depths and step widths of the discontinuities are interrelated. Undoubtedly, however, the different a priori assumptions affect the inversion results.

The reversed P-wave velocity of 7.5 km/s seems high for lower crust and low for uppermost mantle although it may be “normal” for the axial mantle. The discrepancy between the interpretations would more or less disappear if the respective axial mantle velocity gradient zone of the present models would be interpreted as lowermost crust. The other studies in question (except Bunch, 1980) interpret wide-angle reflections to indicate a deeper ~ 7.8 km/s Moho; the brightness of those reflections is not strong, but is enhanced by bandpass filtering. No clear PmP is seen in the present data, perhaps because of the reduced sensitivity at low frequencies, but doubtful indications for it exist in section 3-AB2.

The present velocity sections are characterized by first-order discontinuities with intervening layers of relatively weak gradients. The gradients are not generally vertical. This is very distinct in the upper crust at 25–40 km distance where V_p increases laterally from 3.4 to 4.3 km/s with a further slow increase to 5 km/s at 100 km. In the lower crust, $V_p \approx 6.2 \pm 0.1$ km/s; it shows little change out to 80 km and then a slight increase to 7.2 km/s at 150 km distance. The lowermost layer has a velocity at its axial top of 7.4 km/s, slowly increasing laterally to 7.8 km/s at 130 km. The axial velocity low of about 0.2 km/s appears as a “syncline” of the contours with a “halfwidth” of 20–30 km. The “Moho” increases in depth from axial 6–7 to <10 km at 150 km distance. The velocity trends with age reflect cooling and filling of cracks, while thickening trends suggest crust–mantle differentiation and volcanic productivity variations.

The different interpretations of what is crust and what mantle and what is the Moho partly rest on the modeled velocity gradients and the corresponding discontinuities and contrasts between the layers, which may somewhat depend on the methods applied. Travel time interpretation, as presented here, is not sensitive to stronger or weaker gradients, and the moderate gradients found by travel time inversion are probably affected by the initial assumptions of weak gradients. Bunch (1980), Smallwood and White (1998), Weir et al. (2001) and also Ritzert and Jacoby (1985) and Peirce et al. (2005) emphasized velocity gradients and calculated synthetic seismograms by which also the amplitudes are taken into account. The results favour stronger gradients in the upper part of the crust such that a discrete boundary between the upper and lower crust layers becomes insignificant. The lower part of the crust is characterized by weaker gradients, but about three times stronger than in our models (see Fig. 5). The cumulative travel times through the crust to some 6 km depth is about the same in both model types; hence the most significant difference is the existence or non-existence of a significant refractor at 6–7 km depth below sea level. The present travel times hardly permit a different interpretation, though the discontinuity or sharp transition zone may be extended over a wider depth interval of, say 1, perhaps 2 km, such that its bottom may reach 8 km, even 9 km, but the definition of the Moho would refer to the mean transition depth between 7 and 8 km. The wide-angle reflections, elsewhere called PmP, do not originate at this “refraction Moho” but from deeper down; in the present data set it is not convincingly seen. Similar reflections observed in Iceland (already by Gebrande et al., 1980) are suggested by Björnsson et al. (2005) to originate from melt lenses in, or at the bottom of, a transition zone.

For comparison, the study by Flóvenz (1980) of crustal phases in Iceland shows that the observed amplitudes strongly favour velocity gradients and do not require a discontinuity to exist between upper and lower crust. If compared to a model that fits the travel times with constant-velocity layers with a step of V_p from ~ 4.5 to ~ 6.3 km/s at < 3 km depth, the stronger upper-crust gradient extends to > 4 km depth; underneath the gradient is small (< 0.1 s $^{-1}$). If applied to the present data, the upper part of the crust of a strong velocity gradient would appear thicker with its bottom extending to, perhaps, 3–4 km depth, instead of 2–3 km. This would not change the model depth of the “Moho refractor” by more than 1 km (since the cumulative travel times in the crust hardly change). The consequences for the gravity modeling (below) are only minor.

A precise numerical estimate of the model uncertainties is not possible because they include uncertainties of the model type, which is up to choice; this includes the programs used. The above discussion clearly shows this. From the statistical scatter and the seismic ray tracing results (Figs. 3 and 4) it is estimated that depths of discontinuities and velocity contours are resolved to better than 0.5 km (i.e. $< 10\%$) and at any well resolved point, the velocity uncertainty is about 0.2–0.3 km/s ($5 \pm 1\%$). The corresponding errors of density are estimated below. These are important input quantities for the gravity inversion to be carried out.

3. Gravity modeling

As pointed out above, morphology and gravity are important for understanding especially the variations of crustal thickness. Reykjanes Ridge is not quite symmetric; the inversion concentrates on a profile across the ridge, especially on the SE ridge flank; the axis-parallel profile 3 shows less variation and is too short for telling much about the variation of the plume influence. On a general scale, gravity reflects density and thickness variations of the crust and upper mantle including cooling and thickening lithosphere in the wavelength range of > 100 km where morphology and Bouguer anomaly are negatively correlated. On shorter scales between 10 and 100 km, the situation may be more complicated with incomplete local isostasy, combined crustal density and thickness variations and 3D effects, e.g. of topography. If topography is compensated by thickness variation alone, a negative correlation with the Bouguer anomaly results; if topography is uncompensated the correlation is positive. The discussion is deferred until after gravity inversion on the basis of the seismic models.

3.1. Bathymetry and gravity data

Topography and gravity data are from the Eysteinsson and Gunnarsson (1995) digital compilation of topography, gravity and magnetics comprising land, ship-borne and satellite-derived gravity (converted to free-air gravity anomalies, short: FA), elevation from Iceland, underway ship soundings and the ETOPO5 database of the world relief. Topography and gravity profiles were computed by data reduction; profile values at 5 km intervals were calculated from the digital files by several averaging methods, because these values are more representative for a strip rather than single observations. Gravity is in the IGSN71 reference system, Bouguer reduction had been made for density 2600 kg/m 3 . The errors are a combination of the individual observational and reduction errors and the average scatter. This type of scatter may locally reach 100 m for bathymetry and 30 mGal for the Bouguer anomaly but is mostly much smaller. Generally the errors relevant for the interpretation are estimated to be about 10 m for topography/bathymetry and 0.5–3 mGal for the Bouguer anomalies.

The whole region, including Reykjanes Ridge (Vogt et al., 1990), is characterized by shoaling of the ocean floor toward Iceland. The ocean basins, surrounding the Icelandic Plateau, are ≤ 2400 m deep, i.e. 1000–1200 m shallower than normal ocean of that age. The crustal section under consideration is that defined by the seismic profiles 4 and 5, complemented along the ridge by profile 3. Moderately positive FA values generally characterize the spreading ridges. The Bouguer anomaly (BA) differs from the FA by the Bouguer reduction by which the density contrast of the seafloor is removed and emphasizes density anomalies of crust and mantle; the BA has generally an inverse relationship with topography/bathymetry, demonstrating mass compensation. The spreading Reykjanes Ridge has a relative crestal minimum embedded in Bouguer anomalies of up to +200 mGal over the deeper basins. The anomaly is not monotonous but has a step-wise structure of ups and downs which is enhanced by subtracting a smooth component of the BA from the actual data profile as shown in Fig. 6. The BA along profile 4/5 varies between 100 mGal over the axis to 160 mGal over the flanks at the distance of about 200 km. The strong BA minimum above the crestal zone suggests a strong density deficit. The main source for this large-scale feature is the thermal anomaly in the mantle (asthenosphere), its boundary

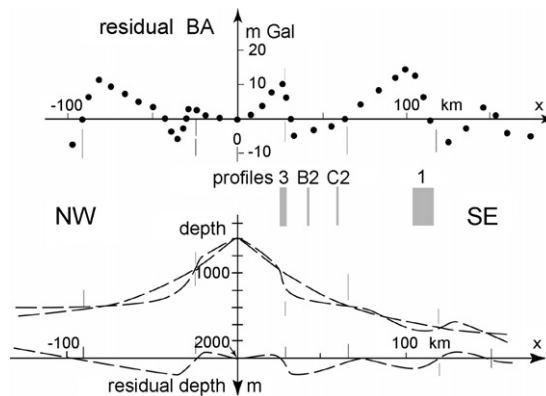


Fig. 6. Residual bathymetry and Bouguer anomaly in a strip covering the transverse profile 4/5.

versus the lithosphere was estimated on the basis of age and the model of thickening lithosphere, $h_{\text{lith}} \approx 7.5t^{1/2}$. The crustal density deficiency in the axial region, determined by the seismic model, also contributes to the BA low. The bathymetry is treated the same way as the BA, by subtraction of the smooth “cooling ridge topography” from the observed profile.

There is no uniform relationship between the gravity and bathymetry variation along the RRISP profiles except for the longest wavelengths. This is evident in the residual ridges and troughs (wavelengths 50–100 km) visible in Fig. 6; the edges of the central horst, at ~ 2.5 Ma crustal age, are accompanied by a “dipolar” gravity anomaly, positive above the high side and negative on the low side suggesting a largely positive correlation. Profile 3 follows the high side of the edge, while profiles B2 and C2 of Bunch (1980) and Smallwood and White (1998) follow the low side at about 4 Ma age; this difference may partly explain the different crustal thicknesses found. The relationship between topography and gravity partly shows a negative correlation, e.g. in the region of RRISP profile 1, near 12 ± 2 Ma crustal age; a gravity low follows a weakly developed morphological ridge; gravity interpretation must take this into account.

3.2. Bayesian inversion with a priori information

The relationship between the seismic models, morphology and gravity is studied for a better understanding of structure and evolution of the ridge. Bayesian gravity inversion is applied with the program package INVERT (Smilde, 1998). In contrast to ordinary inversion starting with some arbitrary initial model, Bayesian inversion starts with an a priori model that is assumed to have some probability of being correct which is described by the parameter error bars. The approach permits to explore the model space. The assumed model parameters are fitted simultaneously with the gravity data, each within their “error bars”. The model parameters are coordinates and densities with errors < 0.5 km and $10\text{--}20 \text{ kg/m}^3$, respectively, as converted from velocity uncertainties of $0.1\text{--}0.3 \text{ km/s}$ (see below for the assumed density–velocity relationships which have, however, not been applied strictly).

The targets or the unknowns in the inversion are the densities and the geometry, represented by 2D polygonal cross-sections at right angles to the morphological ridge axis. The so-called Talwani et al. (1959) method is used for computing gravity effects at observation points. The gravity observations are at sea level. As the gravity effects are non-linear functions of the geometrical parameters, the inversion scheme is non-linear and iterative with stepwise linearization which is performed numerically with finite difference coefficients at all observation points with respect to all model parameters. The normal equations are based on the linearized “observation equations” and are solved for the parameter adjustment which is repeated to some pre-set limit. The condition is usually the least-squares norm (L_2), but other norms can be chosen.

Judging the results requires experience, because formal standard errors are misleading as they describe only the consistency between observations and calculated model effects, no matter how close the model is to, or how far from, the geological reality. This is due to the principal ambiguity of potential field inversion. Whether a result is accepted or not is ultimately a subjective decision, especially, if the uncertainties of the initial assumptions are themselves uncertain. Mathematical inversion as such does not guarantee objectively reliable results.

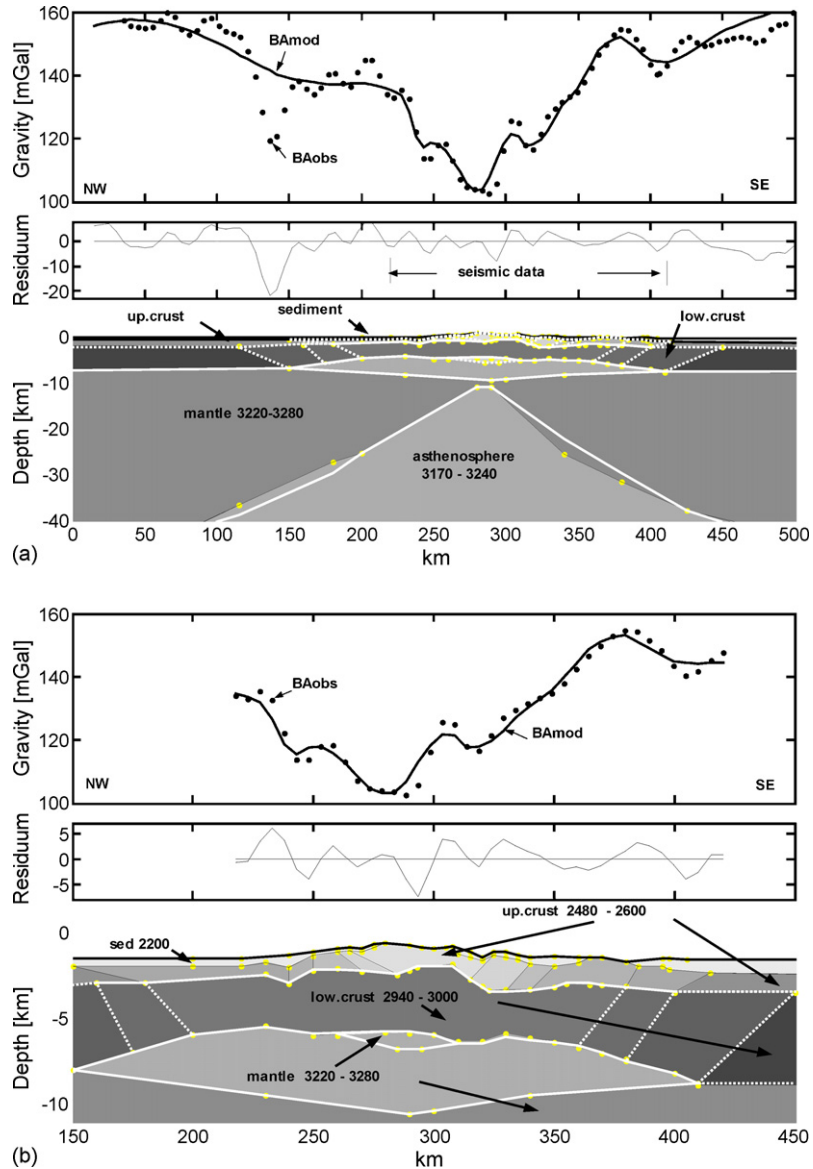


Fig. 7. Gravity profile across Reykjanes Ridge, along profile 4/5 combined from seismic profiles 4 and 5 and inverted for densities. (a) Five hundred kilometers profile extrapolated from the 200 km long seismic profile 4/5; (b) enlargement of a 300 km central section emphasizing the crustal structure. The lines indicate body boundaries and velocity contours (realized in the models as density contrast boundaries).

3.2.1. The *a priori* models

As a *a priori* input, the seismic models for the closely neighbouring parallel profiles 4 and 5, <10 km apart (Figs. 1b and 4), are combined to one, called profile 4/5. The seismic models of Fig. 4 span a distance of about 200 km while the gravity data are extended further; the model profile is thus extrapolated to nearly 500 km length (Fig. 7a) where the layered structures have been calculated within wide *a priori* limits to fit gravity. The ends were extended even beyond the 500 km limits, in order to avoid edge effects. Fig. 7b shows an enlargement of a 300 km segment that contains the 200 km where seismic data exist. Profile 3 was also modelled and extended beyond its ends for the same reasons as above. The BA and the seismic model show only small variations along the profile, and the inversion which is not shown here, demonstrates a good fit to the data; the density values are very close to those obtained for profile 4/5. For additional information on the ridge-parallel variations compare with the gravity model of Peirce et al. (2005), which, of course, suffers from the same ambiguity problem.

The density parametrisation of the models is briefly explained. Instead of taking the absolute density values (as may be done), the lateral density contrasts are taken, which means that a uniform horizontal layer (or layers) is subtracted from the model with lateral edges at large arbitrary distance. It is, however, not necessary to define the model bodies relative to a common reference density; a more economic way is to model each seismic layer boundary geometrically by a polygonal line between the endpoints on each side, which are taken at the same depth such that the line connecting them is a horizontal side of the polygon which produces no laterally varying gravity effect. At each boundary the density contrast $\Delta\rho_{i+1,i}$ follows from the contrast of V_p from layer to layer, because it is this contrast combined with the boundary geometry which produces the gravity effect of interest. The absolute density of any one layer is the sum of the layer contrasts from the surface down to its top.

An important aspect of the models is that the velocity variations within each layer are taken into account by corresponding density variations $\Delta\rho$ described by the velocity contours (Fig. 4); $\Delta\rho$ is superimposed on the constant background value ρ_i for each layer i , such that at any point the local density value is the sum $\rho_i + \Delta\rho$. The constant ρ_i is assumed to be the off-axis value (at the edges) and $\Delta\rho$ is the density decrease towards the axis; $\Delta\rho$ is defined stepwise as depicted by the contours which are parameterized by polygons across which the density is incremented by a fixed value (between the adjacent contours, density is assumed uniform). The parametrisation is realized most economically by circumscribing the structure along the (polygonal) contours in spiral fashion such that a “2D volume” or area of, say, 1 density increment is surrounded once, of two increments twice, etc., as many times as needed.

The initial density values ρ_i and the increments $\Delta\rho$ are estimated from the seismic velocities on the basis of the relationships of Birch (1960, 1961) or Carlson and Herrick (1990) and Zelt (1992), as applied also by Darbyshire et al. (2000); $\Delta V_p = 0.2\text{--}0.3\text{ km/s}$ relates to $\Delta\rho$ of about $20\text{--}30\text{ kg/m}^3$. The relationships refer to different velocity ranges and appear appropriate; small differences are immaterial since the a priori assumptions are given the freedom to be adjusted in the gravity inversion. When modeling the ridge and adjusting the densities by fitting the Bouguer anomalies, it must be considered that the Bouguer reduction had been applied to the data with a Bouguer density ρ_B of 2600 kg/m^3 ; this figure is relevant only to the “filling” up of the ocean water to rock density. For the sediments with an assumed a priori total density ρ_1 of 2200 kg/m^3 , this means that $\Delta\rho_1 = 2200\text{--}2600 = -400 \pm 20\text{ kg/m}^3$. The upper crust has $V_p \approx <3.4\text{ km/s}$ (near the axis) to 4.9 km/s (at the profile ends), and density varies accordingly from 2490 to 2600 kg/m^3 ; hence the density contrast versus sediments is (initially) $+400 \pm 50\text{ kg/m}^3$. The lower crust with velocities increasing from 6.2 km/s near the axis to 7.2 km/s at the flanks suggests density to vary from about 2930 to about 3000 kg/m^3 ; this is a contrast of $+400 \pm 50\text{ kg/m}^3$ versus the upper crust. The initial uppermost mantle densities are estimated as $3190\text{--}3250\text{ kg/m}^3$ from the velocities of about $7.5\text{--}7.8\text{ km/s}$, suggesting a further density increment across the Moho of $250 \pm 50\text{ kg/m}^3$. For the asthenosphere (or thickening lithosphere) about which the seismic models of Fig. 4 give no information but which is assumed to exist and is needed to fit gravity, the initial density assumption is $3150\text{--}3220\text{ kg/m}^3$ in two consecutive steps of -30 and -70 kg/m^3 (± 50) versus mantle lithosphere.

The density uncertainties or errors are estimated from the velocity errors with the same ρ – V_p relationships. An error of 0.3 km/s , corresponds to $\sim 50\text{ kg/m}^3$. Depth uncertainty for the upper layers is about $0.1\text{--}0.2\text{ km}$ and for the Moho $0.3\text{--}0.5\text{ km}$. The error of the BA is assumed $\pm 5\text{ mGal}$ above the seismic model and $9\text{--}10\text{ mGal}$ for the ends of the profile; gravity cannot be fitted within $0.5\text{--}3\text{ mGal}$ of the basic data with the chosen model parametrization; this must be taken into account. The parameters are “adjusted” by the inversion to fit gravity and the initial assumptions within the error limits. Inversion calculations were carried out with various input assumptions, e.g. with and without changing the geometry of the seismic model.

As velocity–density relationships are not unique and vary with rock types or chemical composition and physical state, and the “real” (non-observational) scatter is considerable, an inversion for densities with good seismic constraints gives more representative results than average ρ – V_p relationships give. Therefore the a posteriori densities are of prime interest and the a priori densities and a priori velocities are also shown in Table 1.

3.2.2. Results of gravity inversion

The model shown was obtained after eight iterations. Where constrained by the initial seismic model, it fits the observed BA with a mean error of about 2 mGal and the residuals range within about $\pm 5\text{ mGal}$, but beyond the seismically constrained geometry, i.e. for $x < 200\text{ km}$ and $x > 400\text{ km}$, the errors or residuals are much larger, up to 20 mGal . The extrapolation is necessary for the calculation of the gravity effects without distortion by edge effects. Extrapolation may also be aided by gravity modeling on the basis of the observed anomalies. However, this has not been considered an aim of the present study, especially not on the NW flank.

The coordinates as given by the a priori seismic model have been modified by the gravity inversion only slightly, usually within the error bounds. The fit is not very sensitive to the assumed geometrical error bounds. The gravity residuals are not quite random having a “wavy” character of ~ 25 km wavelength. This is largely a consequence of the model parametrization with a coarser spatial resolution than the data intervals permit, but it is considered sufficient. Outside the seismic section (Fig. 4) the larger residuals have longer “wavelengths” (< 50 km), again due to the coarse parametrization of the extrapolated structures, largely in horizontal layers.

Interpolation in regions of seismic data gaps between the seismic models of Smallwood and White (1998) and Weir et al. (2001), e.g. along the “longitudinal” profile 3 (not shown), is more reliable than extrapolation. The considerable differences, especially in Moho depth, will be considered in the discussion section.

The improvements to the initial densities by the inversion were small, of the order of $10\text{--}20\text{ kg/m}^3$, thus the velocity-based predictions were quite close to the final densities. The trend line has roughly a slope of 0.16 kg/m^3 per km/s which may be acceptable for largely crystalline rocks. The mantle rocks seem to be systematically denser than predicted by the trend. It appears likely that partial melting plays a role which reduces density less than it reduces the velocities. Moreover, a purely vertical density increase $\rho(z)$ which may be superimposed on the models does not generate gravity variations and thus escapes the gravity inversion.

As mentioned, criteria for “good” gravity models are difficult to define in the light of their principal ambiguity. A literally taken a priori model is easily disproved, but adjustments may make it acceptable. This pertains especially to velocity-deduced densities. A soft criterion for modeling success may be the ease of fitting all relevant data on gravity and the seismic (or other) a priori models without too much coercion. The present results satisfy such a criterion and are hence relatively trustworthy.

4. Discussion of Reykjanes Ridge structure and evolution: similarities and differences

The most striking differences between the present results and the ones to the NE and SW, best seen in Fig. 5, are (1) the reduced crustal thickness, < 6 km at the axis versus 10 km (profile C1 at about 61°N) to the SW and 13 km to the NE (profile RISE at axis near 63°N) and (2) the apparent crustal thickness increase with age in contrast to an apparent thickness decrease. Profile 4/5 shows an increase from < 6 to ~ 8 km at about 15 Ma age, while profile C1 indicates a decrease from 10 to ~ 8 km at 5 Ma, and the corresponding RISE values nearer to Iceland are 13 and 10 km. The greater axial crustal thickness and the lateral decrease with age is explained by a temporal increase in plume flow and/or temperature and melting since about 2–3 Ma which coincides with the beginning of rifting in the Eastern Volcanic Zone (EVZ) of south Iceland. This is supported by the axial horst structure instead of a rift and the clearly developed volcanic ridges (Jacoby, 1980; Appelgate and Shor, 1994). While it is not implausible, our results indicate the crust to thicken as it ages. Or is it possibly a problem of interpretation in terms of crust–mantle transition?

Several possibilities exist:

- (1) In a technical sense, systematic differences in the results may result from the differences in the nature of the experiments (shot density, frequency content) and interpretation methods between the different studies. The present data display refracted waves (used for the modelling), but no unambiguous wide-angle reflections, which, on the other hand, are seen by Smallwood and White (1998) and Weir et al. (2001) in their sections, especially when filtered by a 2–8 Hz bandpass, and interpreted as PmP. Consequently, the a priori assumptions in the inversions differ and affect the results. In the present study, layers with moderate velocity gradients and significant first-order discontinuities between them have been assumed a priori, versus layers with stronger gradients and very small velocity discontinuities. These differences cannot explain the Moho depth differences in the models, but it is not clear to what extent the initial assumptions affect them.
- (2) A geographical aspect is that profile 3 lies on the edge of the central horst (at about 2.5 Ma crustal age) while profiles B2, C2 and partly the RISE data (Bunch, 1980; Smallwood and White, 1998; Weir et al., 2001) follow the foot of the slope at about 4 Ma crustal age. Crustal thickness varies with age as suggested by morphology and gravity (Figs. 1, 6 and 7); the relations were discussed in connection with Fig. 6. This might partly explain the difference between apparent crustal thickening and the apparent thinning, but it cannot explain the different absolute crustal thicknesses found or the thinning toward the axis in profile 4/5. On the > 100 km scale cooling leads to a negative correlation of morphology and Bouguer anomaly, on shorter scales, incomplete local isostasy and crustal density and thickness variations may lead to positive or negative correlation, depending on the situation

that predominates. To the SW, near 58°N, the Bouguer anomaly–morphology relationship is different, where a BA maximum is located above the crest. The interpretation of the FA by Peirce et al. (2005) indicates a continuous crustal thickening toward Iceland up to the present study area, on the a priori assumption of the “reflection Moho”; under this assumption it does not contradict the present models.

- (3) The discrepancy of crustal thickness at the axis is subject to the interpretation of the seismic data. While Figs. 4 and 5 indicate a distinct velocity discontinuity interpreted to be the Moho, the axial low-velocity lens underneath suggests that P-wave velocities close to 8 km/s are reached at about 12 km depth. It cannot be ruled out that the lens may be lower crust, in which case the discrepancies between the various studies would disappear. It is a question of definition of crust and mantle, but this is ambiguous in the present range of the values of the geophysical parameters seismic velocity and density; gravity inversion with a slightly lower, more “crust-like” a priori density of the lens leads to equally acceptable results as those of Fig. 7; this is being investigated further. Nevertheless, the clear refractor seen in the present models is considered the more likely “refraction Moho”. The little indication in the present data (Fig. 2) of wide-angle reflections, is considered less convincing for the existence of a deeper “reflection Moho”.
- (4) The fundamental definition of crust and mantle lies in the petrology and composition of the material between the “refraction Moho” and the “reflection Moho”. Is it the basaltic product of mantle melting or the ultramafic residuum? Do the geophysical characteristics require it to be predominantly the former or predominantly the latter? Or might it be a transition layer between the two and transient in the sense of evolving during a few million years from the axial transition into a Moho that separates the ultramafic mantle from a thickened mafic crust? And could not in such a transition zone melt lenses in its deeper part naturally occur and give rise to frequent wide-angle reflections? Such an interpretation for the equivalent layer under central Iceland has been favoured by Björnsson et al. (2005), Kaban et al. (2002) and Fedorova et al. (2005).

It is interesting to compare the present results to recent ones from Kolbeinsey Ridge (KR) (Kodaira et al., 1997). The results are very similar to the present ones from Reykjanes Ridge (RR) which may be related to the fact that both studies were carried out at about the same distance from the supposed plume centre. The significant axial velocity reduction, especially in the upper crust, found at KR is quite similar to that at RR (Fig. 4). The same can be said about the seismic velocities generally in the upper and middle or lower crustal layers. It is noteworthy that the similarity is not due to applying the same interpretation methods, which was not the case. The similarity of RR and KR structures is remarkable in view of the following dissimilarities. The main axial plume flow seems to be channelled along the RR while possibly blocked from KR by the Tjörnes Fracture Zone (TFZ). The KR is characterized by a central rift north of about 67°50'N with water depths >1000 m, in contrast to RR that is <1000 m deep for ~400 km SW of Iceland and has no rift for >800 km up to ~57°N.

The different interpretations of the data from RR have consequences for the geodynamic models. The discrepancies in crust–mantle interpretation strongly resemble similar discrepancies in the interpretation of seismic models in Iceland (e.g. Gebrande et al., 1980; Darbyshire et al., 2000; Fedorova et al., 2005). Clearly, the situation there is anomalous, and the terms “crust” and “Moho” should be used with caution and not literally equated with the continental models. It seems likely that the situation at the axial Reykjanes Ridge, at least for a few hundred kilometres from the plume centre is similar to that in Iceland.

5. Summary and conclusions

Explosion deep seismic sounding data of high quality from ~62°N as part of the Reykjanes Iceland Seismic Project (RRISP77) are presented. Clear refracted phases are modelled and optimized by crust and upper mantle structure of layers with moderate velocity gradients. They are taken as a priori information for Bayesian inversion of gravity data. The models differ from the published seismic models to the NE and SW by a lower axial crustal thickness and an apparent thickness increase with crustal age instead of a decrease. The discrepancy is a matter of interpretation of the seismic data; the axial low-velocity lens, where an 8 km/s P-wave velocity is reached near 12 km depth, has been interpreted here as an anomalous part of the upper mantle, but may alternatively be taken to belong to the lower crust. In that case the discrepancies between the various studies would disappear. The gravity inversion only slightly modifies the seismic model by fine-tuning both in structure and properties. Thus, the seismic and gravity data are in excellent agreement.

Preference is given to the interpretation of the axial lense-shaped body of rather low P-wave velocities (7.4–7.7 km/s) to be part of the mantle rather than lowermost crust for the following reasons. (1) In the present data there is minimal evidence for wide-angle reflections which are regarded by others as PmP, i.e. coming from a deeper Moho. (2) On the other hand, the refracted phase that we interpret as Pn is very distinct on all RRISP77 profiles from 0 to 10 Ma crustal age; the corresponding refractor is, as generally in refraction seismology, taken to define the seismic Moho or crust–mantle boundary. Along the ridge, irrespective of this discrepancy, the different experiments show a decrease of crustal thickness from about 14 to 8 km from Iceland southward to 58°N.

Finally, our preference must not be taken for a claim of absolute knowledge of the truth. As in the thin-crust versus thick-crust controversy in Iceland, reconciliation may lie in kind of a compromise on the basis of knowledge about the composition and nature of the material between the respective refractor and reflector. Such a knowledge can be gained only indirectly, e.g. by combining as many methods of investigation as possible. The layer in question may be best envisioned (Björnsson et al., 2005; Kaban et al., 2002; Fedorova et al., 2005) as a transition layer of melt accumulation within a largely solid matrix, transient in the sense that it is differentiating and evolving in several million years partly into the lowest crust and partly into the uppermost mantle as found under the older parts of Iceland (Darbyshire et al., 2000) or along RRISP profile 1 along the ~10 Ma old SE flank of Reykjanes Ridge (Ritzert and Jacoby, 1985). Such a process would naturally explain the discrepancies of interpretation and especially the thickening of the crust with age. Certainly the terms “crust” and “Moho” cannot be equated with the corresponding continental features, neither in Iceland, nor at the axial Reykjanes Ridge for several hundred kilometres from the plume centre.

The authors of the studies SW and NE from ours explain variation of crustal thickness increasing towards the axis by a temporal increase in plume temperature and/or outflow since about 2–3 Ma which coincides with the beginning of rifting in the Eastern Volcanic Zone of south Iceland. The plume influence is undoubtedly evident in the axial horst structure instead of a rift and the clearly developed volcanic ridges (Jacoby, 1980; Appelgate and Shor, 1994; Peirce et al., 2005). While a “recent” increase in plume productivity seems to be real, the depth variation can be equally well explained by a transient evolving layer. A currently high plume activity would, indeed, enhance such an effect. A combination of both aspects is thus possible. Velocity trends with crustal age across the ridge in all layers very likely reflect lithosphere cooling, filling of cracks and, at greater depth, crystallisation and differentiation, while some of the crustal thickness variations support the suggestion of changes in the volcanic productivity.

Acknowledgements

Captain and crew of RV Meteor during cruise 45 did their best to make the experiment a success, and so did all students, technicians and scientists onboard. We thank especially Rolf Herber for excellent coordination of the work at sea and handling of the explosives onboard. We thank for the help we received in preparing this contribution. Bosco Loncarevic, Bedford Institute of Oceanography, Dartmouth, Nova Scotia, Canada, participated in the cruise, provided ocean bottom seismometers (OBS) and helped analyzing the data. Sean Solomon, Massachusetts Institute of Technology, Cambridge, MA, USA, likewise took part and provided an OBS. Mohamed Rahal picked the seismogram arrival times and carried out the seismic inversions for crustal structure. Elke Hillermann contributed a lot by her efficient data management and computations. Petra Koppenhöfer excellently prepared some of the figures. Two anonymous reviewers made very useful suggestions about the seismic interpretation, especially of wide-angle reflections, on relevant literature and the work of other groups. They helped us clarify the presentation and improve the English.

References

- Angenheister, G., Gebrande, H., Miller, H., Weigel, W., Goldflam, P., Jacoby, W., Pálmason, G.G., Björnsson, S., Einarsson, P., Zverev, S., Loncarevic, B., Solomon, S., 1979. First results from the Reykjanes Ridge Iceland Seismic Project 1977. *Nature* 279, 56–60.
- Angenheister, G., Gebrande, H., Miller, H., Goldflam, P., Weigel, W., Jacoby, W.R., Pálmason, G., Björnsson, S., Einarsson, P., Pavlenkova, N.I., Zverev, S., Litvinenko, I.V., Loncarevic, B., Solomon, S., 1980. Reykjanes Ridge Iceland Seismic Experiment (RRISP 77). *J. Geophys.* 47, 228–238.
- Appelgate, B., Shor, A., 1994. The northern Mid-Atlantic and Reykjanes Ridges: spreading center morphology between 55°50'N and 63°00'N. *J. Geophys. Res.* 99, 17935–17956.
- Birch, F., 1960. The velocity of compressional waves in rocks to 10 kbar, 1. *J. Geophys. Res.* 65, 1083–1102.
- Birch, F., 1961. The velocity of compressional waves in rocks to 10 kbar, 2. *J. Geophys. Res.* 66, 2199–2224.
- Björnsson, A., 1985. Dynamics of crustal rifting in NE Iceland. *J. Geophys. Res.* 90, 10151–10161.

- Björnsson, A., Eysteinnsson, H., Beblo, M., 2005. Crustal formation and magma genesis beneath Iceland: magnetotelluric constraints. *Geol. Soc. Am. Spec. Pap.* 338, 665–686.
- Bunch, A.W.H., 1980. Crustal development of the Reykjanes Ridge from seismic refraction. *J. Geophys.* 47, 261–264.
- Carlson, R.L., Herrick, C.N., 1990. Densities and porosities in the oceanic crust and their variations with depth and age. *J. Geophys. Res.* 95, 9153–9170.
- Cerveny, V., Klimes, L., Psencik, I., 1984. Paraxial ray approximation in the computation of seismic wavefields in inhomogeneous media. *Geophys. J. R. Astr. Soc.* 79, 89–104.
- Darbyshire, F.A., White, R.S., Priestley, K.F., 2000. Structure of the crust and uppermost mantle of Iceland from a combined seismic and gravity study. *Earth Planet. Sci. Lett.* 181, 409–428.
- Einarsson, P., 1979. Travel times recorded at Icelandic seismograph stations during the Reykjanes Ridge Iceland Seismic Project (RRISP). *Publ. Sci. Inst. Univ. Iceland*, RH-79-10.
- Eysteinnsson, H., Gunnarsson, K., 1995. Maps of gravity, bathymetry and magnetics for Iceland and surroundings. *Orkustofnun*, OS-95055/JND-07.
- Fedorova, T., Jacoby, W.R., Wallner, H., 2005. Crust–mantle transition and Moho model for Iceland and surroundings from seismic, topography, and gravity data. *Tectonophysics* 396, 119–140.
- Flóvenz, Ó.G., 1980. Seismic structure of the Icelandic crust above layer three and the relation between body wave velocity and the alteration of the basaltic crust. *J. Geophys.* 47, 211–220.
- Gebrande, H., Miller, H., Einarsson, P., 1980. Seismic structure along RRISP—profile I. *J. Geophys.* 47, 239–249.
- Goldflam, P., Weigel, W., Loncarevic, B.D., 1980. Seismic structure along RRISP—profile I on the southeast flank of the Reykjanes Ridge. *J. Geophys.* 47, 250–260.
- Grevenmeyer, I., Weigel, W., 1996. Seismic velocities of the uppermost igneous crust versus age. *Geophys. J. Int.* 124, 631–635.
- Jacoby, W.R., 1980. Morphology of the Reykjanes Ridge crest near 62°N. *J. Geophys.* 47, 81–85.
- Kaban, M., Flovenz, O., Pálmason, G., 2002. Nature of the crust–mantle transition zone and the thermal state of the upper mantle under Iceland from gravity modelling. *Geophys. J. Int.* 149, 281–299.
- Kodaira, S., Mjelde, R., Gunnarsson, K., Shiobara, H., Shimamura, H., 1997. Crustal structure of the Kolbeinsey Ridge, North Atlantic, obtained by use of ocean bottom seismographs. *J. Geophys. Res.* 102, 3131–3151.
- Pearce, C., Gardiner, A., Sinha, M., 2005. Temporal and spatial cyclicity of accretion at slow spreading ridges—evidence from the Reykjanes Ridge. *Geophys. J. Int.* 163, 56–71. doi:10.1111/j.1365-246X.2005.02738.x.
- Ritzert, M., Jacoby, W.R., 1985. On the lithospheric seismic structure of Reykjanes Ridge at 62.5°N. *J. Geophys. Res.* 90, 10117–10128.
- Searle, R.C., Keeton, J.A., Lee, S.M., Owens, R.B., Mecklenburgh, R., Parsons, B., White, R.S., 1998. The Reykjanes Ridge: structure and tectonics of a hot-spot influenced, slow spreading ridge, from multibeam bathymetry, gravity and magnetic investigations. *Earth Planet. Sci. Lett.* 160, 463–478.
- Sinha, M.C., Constable, S.C., Pearce, C., White, A., Heinson, G., MacGregor, L.M., Navin, D.A., 1998. Magmatic processes at slow spreading ridges: implication of the RAMESSES experiment at 57°45′ on the Mid-Atlantic Ridge. *Geophys. J. Int.* 135, 7381–7745.
- Smallwood, J.R., White, R.S., 1998. Crustal accretion at the Reykjanes Ridge 61°–62°N. *J. Geophys. Res.* 103, 5185–5201.
- Smilde, P.L., 1998. Verwendung und Bewertung von a priori Information bei potentiell singulären Inversionsproblemen am Beispiel der gravimetrischen Bestimmung von Dichteverteilungen. *Deutsche Geodät. Komm.* C 490 (München).
- Talwani, M., Worzel, J.L., Landisman, M., 1959. Rapid gravity computations for two-dimensional bodies with application to the Mendocino submarine fracture zone. *J. Geophys. Res.* 64, 49–59.
- Vogt, P.R., 1974. The Iceland phenomenon: imprints of a hot spot on the ocean crust, and implications for flow below the plates. In: Kristjánsson, L. (Ed.), *Geodynamics of Iceland and the North Atlantic Area*, pp. 105–126.
- Vogt, P.R., Nishimura, C., Jarvis, J., 1990. Prospecting for active volcanism. A promising role for long-range aircraft along the mid-oceanic ridge. *EOS* 71, 773–774.
- Weigel, W., 1980. Aufbau des Reykjanes Rückens nach refraktionsseismischen Messungen. In: Weigel, W. (Ed.), *Reykjanes Rücken, Island, Norwegischer Kontinentalrand. Abschlusskolloquium, Hamburg zur Meteor-Expedition*, vol. 45. DFG, Bonn, pp. 53–61.
- Weir, N.R.W., White, R.S., Brandsdóttir, B., Einarsson, P., Shimamura, H., Shiobara, H., 2001. Crustal structure of the northern Reykjanes Ridge and Reykjanes Peninsula, southwest Iceland. *J. Geophys. Res.* 106, 6347–6368.
- Zelt, C.A., Smith, R.B., 1992. Seismic traveltime inversion for 2D crustal velocity structure. *Geophys. J. Int.* 108, 16–34.
- Zelt, C.A., 1992. Documentation of “tramp” and related programs. Available at <http://zephyr.rice.edu/departments/faculty/zelt/rayinv.html>.