Crustal accretion at the Reykjanes Ridge, 61°–62°N

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Abstract. We report results of a seismic, gravity, and magnetic survey of the Reykjanes Ridge spreading center at 61°–62°N, about 600 km from the center of the Iceland mantle plume. Anomalously shallow water on the ridge crest enabled us to record seismic refractions on a 2.4 km hydrophone streamer. The velocity within layer 2A is $2.4 \pm 0.3$ km s$^{-1}$, and its mean thickness is $400 \pm 100$ m. The velocity at the base of layer 2A is $3.3 \pm 0.3$ km s$^{-1}$ on the ridge axis, increasing with crustal age to $4.0$ km s$^{-1}$ at 1.5 Ma and $4.5$ km s$^{-1}$ at 5 Ma. Assuming that seismic layer 2A on the ridge axis is also the extrusive layer, i.e., the magnetic source layer, we have successfully modeled the variations in amplitude of the magnetic field. The best magnetic model includes enhanced magnetization within layer 2A at the sites of recent volcanic activity as independently recognized in side-scan sonar data. We also present a full crustal seismic model, based on wide-angle seismic recordings on digital ocean bottom hydrophones and disposable sonobuoys. The seismic model is complemented by gravity modeling, which further suggests that the ridge crest is in isostatic equilibrium. The zero age crust is 10.0 km thick, while crust of age 5 Ma is 7.8 km thick. These crustal thicknesses are greater than those of normal oceanic crust, which we attribute to the presence of anomalously hot asthenospheric mantle beneath the spreading center. We suggest that the variation in thickness between 0 Ma and 5 Ma crust is caused by temporal variation in the plume-fed asthenospheric temperature beneath the Reykjanes Ridge.

1. Oceanic Crust Near the Iceland Plume

The anomalously hot mantle of the Iceland mantle plume profoundly affects the formation of oceanic crust at the Mid-Atlantic Ridge, causing the spreading center to be subaerial on Iceland and anomalously shallow on the adjacent submarine Reykjanes and Kolbeinsey Ridges. We report results from a survey of the Reykjanes Ridge collected aboard RRS Charles Darwin cruise CD70 at 61°–62°N, some 600 km from the center of the mantle plume.

Away from fracture zones and mantle plumes, the average thickness of oceanic crust is $7.1 \pm 0.8$ km [White et al., 1992]. Elevated asthenospheric temperatures beneath the spreading centers around Iceland, however, cause decompression melting to start deeper and thus to generate thicker crust than at spreading centers underlain by normal temperature mantle [White et al., 1995]. Along-axis, the crustal thickness of the Reykjanes Ridge decreases and its depth increases away from the plume center, with the axial high near Iceland giving way to a median valley south of 59°N [Laughton et al., 1979].

Spreading is at a full rate of 20.2 mm yr$^{-1}$ at 61.5°N [DeMets et al., 1990], in a direction 25°–30° oblique to the ridge axis normal. Along the ridge, a long-wavelength bathymetric swell and associated gravity anomaly south of Iceland, and the geochemical signature of dredged rock samples [Taylor et al., 1995] are indicative of a pulse of anomalously hot asthenospheric mantle presently traveling away from Iceland [White et al., 1995; White, 1997]. In this study we demonstrate that anomalously thick crust has been generated by this pulse of hot mantle and suggest that similar variations in crustal thickness underlie the V-shaped gravity anomaly and topographic lineations that flank the Reykjanes Ridge, cutting across the magnetic stripe isochrons. These V-shaped ridges were recognized by Vogt [1971] and are now understood to record the passage of anomalously hot asthenospheric mantle away from the plume center [White et al., 1995; White, 1997].

As a slow spreading mid-ocean ridge, the Reykjanes Ridge is unusual in the absence of transform faults, with just one small offset occurring near 58°N [Searle et al., 1994]. Local segmentation is marked instead by axial volcanic ridges (AVRs), which are typically 10–40 km long and over 100 m high [Searle and Laughton, 1981; Keeton et al., 1997]. The AVRs generally overlap each other by up to 50% of their length and exhibit a dominant strike of N014°E, intermediate between the
spreading orthogonal direction and the ridge azimuth of N036øE [Searle and Laughton, 1981; Parson et al., 1993]. During the AVR building phase, magmatism accounts for a large part of the plate separation. Subsequently, they may become disaggregated by faulting as magmatism decreases, and they are moved away from the ridge axis [Murton and Parson, 1993; Parson et al., 1993].

2. CD70 Reykjanes Ridge Survey

Multichannel near-normal incidence and wide-angle seismic, bathymetric, magnetic, and gravity data were collected on CD70. A multibeam bathymetric survey over the area was completed subsequently on cruise CD87 [Keeton et al., 1997]. The CD70 survey (Figure 1) consisted of five intersecting seismic lines, CAM 71–CAM 75. CAM 71 lies along the ridge axis, running obliquely over several axial volcanic ridges, and CAM 73 lies parallel to it along magnetic chron 3n (~5 Ma) in the free-air gravity low between highs defining the first and second V-shaped ridges [Vogt, 1971; White et al., 1995]. CAM 72, CAM 74, and CAM 75 run orthogonally to CAM 71, across the ridge axis and out to crust of age ~7 Ma.

The seismic source was a 12-gun, 4666 inch³ (76.5 L) airgun array, with a ~100 m shot spacing. The shallow water (~800 m) of the spreading axis allowed us to record seismic refractions from the uppermost crust on the 2.4 km hydrophone streamer and hence to determine the shallow velocity structure of seismic layer 2A.

3. Upper Crustal Structure

3.1. Seismic Structure of Layer 2A

Although originally defined as the shallow, highly magnetized layer of the oceanic crust [Talwani et al., 1971], “layer 2A” is now generally used as a seismological term [Houtz and Ewing, 1976]. Purdy [1987] used a fixed ocean bottom hydrophone and charges detonated near the seafloor to determine seafloor velocity of 2.1 km s⁻¹ and a velocity gradient of 4 s⁻¹ within the median valley of the Mid-Atlantic Ridge (MAR) at 23°N. Subsequently, there has been little published work to date specifically on layer 2A in the Atlantic [see Grevemeyer and Weigel, 1996], although there have been several studies of the East Pacific Rise (EPR) and Juan de Fuca ridge systems [e.g., Kappus et al., 1995]. The total thickness of layer 2A is 100–300 m on the EPR axis, increasing to double this within 0.5–3 km off-axis. Hooft et al. [1996] have developed a stochastic model of lava eruptions and dike emplacement which recreates the observed thickening of layer 2A. The uppermost igneous velocity is typically ~2.4 km s⁻¹ with a vertical velocity gradient of ~1 s⁻¹ at the top, increasing with depth.

At spreading rates lower than those of the EPR, there is more variability in the thickness of layer 2A. The best constrained layer 2A velocity structure on the intermediate spreading rate Juan de Fuca Ridge is for the area around 45°N [McDonald et al., 1994]. Their preferred average model is again a layer 250 m thick with a low vertical velocity gradient from ~2.7 km s⁻¹ at the seabed, underlain by a layer with a higher velocity gradient taking the velocity to 4.8 km s⁻¹ at ~350 m below the seafloor. McDonald et al. [1994] report that layer 2A shows significant seismic anisotropy (higher velocities parallel to the ridge) and that the thickness varies between 200 and 550 m nonsystematically with respect to the ridge axis. White and Clowes [1990] and Cudrak and Clowes [1993] present models from an ocean bottom seismometer refraction survey at 48°N on the Juan de Fuca Ridge showing thickness variation of layer 2A from 250 to 650 m using a spatially constant velocity model: 2.5 km s⁻¹ at the seabed, overlying an unconstrained vertical velocity gradient of ~1 s⁻¹.

We identified two distinct refracted phases from the uppermost igneous crust on the ridge axis from shot gather records on the 2.4 km, 48-channel hydrophone streamer. Neither of these two first-arrival phases displayed the concave curvature that would be expected from a structure with a seismic velocity which increases significantly with depth (Figure 2), although a curving phase corresponding to the continuation of the second
refraction was identified in some of the digital ocean bottom hydrophone (DOBH) sections (section 4). We have constructed a simple two-layer model of the upper crust using straight-line least squares fits to the travel times of the two refracted phases. We corrected the apparent phase velocity for the dip of the seabed between the ray entry and exit points on the seafloor for each shot.

A weak first-arrival phase asymptotic to the seabed reflection, interpreted as the seafloor refraction (Figure 2a), was observed sparsely through the data set. We attribute the rarity of this phase to the highly fractured, vesicular and unconsolidated nature of the young volcanic rocks on the ridge axis, which causes the refraction to be highly attenuated.

Where the water depth was less than ~1200 m, we frequently observed a strong refracted phase as a first arrival on up to 20 of the farthest offset traces (Figure 2). This phase was not tangential to the seabed reflection phase, so was not a refraction which had traveled immediately below the seabed. A continuous velocity gradient from the seafloor downward, as assumed by Ewing and Purdy [1982] for the uppermost oceanic crust, would not produce the two distinct refracted phases that we observe; there must be a discontinuity in the velocity gradient or a first-order velocity step to produce two phases with distinct apparent velocities.

There is no continuous reflector in the stacked multichannel seismic (MCS) data at the depth that has been interpreted as the base of layer 2A from the refraction data, although there are hints of events coinciding with the expected horizon [Smallwood, 1997]. The absence of a base 2A near-normal incidence reflection is not unusual: the same combination of a wide-angle layer 2A event without significant energy at near-normal incidence has been noted in data sets from the MAR and the EPR [Christeson et al., 1996]. The absence of a near-normal incidence reflection suggests that the base of layer 2A is usually gradational or is laterally discontinuous on the scale of the seismic wavelengths (~50-200 m).
3.2. Seafloor Refraction Velocity

The seafloor velocity on CAM 71 ("zero age" crust) is $2.4 \pm 0.3 \text{ km s}^{-1}$ (number of samples, $n=16$). On the across-axis lines CAM 72 and CAM 74, seafloor velocities are $2.7 \pm 0.1 \text{ km s}^{-1}$ ($n=9$) and $2.5 \pm 0.1 \text{ km s}^{-1}$ ($n=21$), respectively. In determining these velocities, a few solutions yielding velocities $> 3.1 \text{ km s}^{-1}$ (close to the mean of the base 2A refraction velocity) were discarded, as these were interpreted as coming from below the base of layer 2A rather than representative of the velocity above the refractor.

We attribute these low igneous velocities of $2.4-2.7 \text{ km s}^{-1}$ as due to the presence of fractures and a high degree of porosity within layer 2A. Using Carlson and Herrick's [1990] velocity-porosity relationship suggests that such velocities indicate basalt porosities of about 60%, although extrapolation outside the data constraining their relationship is required to produce this estimate.

It might be expected that the seafloor refraction velocity would increase as the crust ages [Grevemeyer and Weigel, 1996]. However, we find all the seafloor velocities to be the same within their error bounds. One possible explanation is that although newly erupted volcanics may be at their maximum vesicular porosity, subsequent tectonic disruption of the volcanic section may increase the bulk porosity or introduce other types of fractures, decreasing the seismic velocity. Another possible explanation for the absence of a significant increase in layer 2A velocity with age is that the rise of the seafloor velocity off-axis may be initiated by the sedimentary sealing of the crustal hydrothermal circulation system [Rohr, 1994]. In our area, significant sediment accumulation is present only on crust older than 1.2 Ma, while our streamer refraction data were only obtained from crust younger than 1.2 Ma.

3.3. A Refraction Model of Layer 2A

For modeling purposes we assumed that the base of layer 2A was locally parallel to the seafloor and assigned the mean determined velocity of $2.45 \text{ km s}^{-1}$ to layer 2A. From each gather of picks we determined a velocity for the refraction from directly beneath the base of 2A (hereafter called the base 2A velocity) and layer 2A thickness.

A series of consistent and smoothly varying determinations of layer 2A thickness vindicated the assumed geometry, and Figure 3 shows a final model based on a smoothed thickness variation. If there were lateral variations in the seismic velocity within layer 2A, then our thickness determinations would be correspondingly affected, with the error on the velocity used ($\pm 0.3 \text{ km s}^{-1}$) giving an approximate error of $\pm 50 \text{ m}$ on the individual layer 2A thickness determinations.

There is no correlation between the thickness of layer 2A and the water depth (correlation coefficient $r = 0.02$), but a weak correlation between the base 2A depth and the base 2A refraction velocity ($r = 0.47$) on CAM 71 suggests that below layer 2A the velocity increases with depth. Corrected for seafloor dip, the mean base 2A refractions velocity on CAM 71 was $3.29 \pm 0.46 \text{ km s}^{-1}$ ($n=269, \sigma$).

With our data it is not possible to rule out a vertical velocity gradient within layer 2A, although a discontinuity in the velocity profile must be maintained at its base. Given the paucity of refracted arrivals observed from within layer 2A itself, a velocity gradient could not be well constrained, but ray trace modeling of individual shot gathers suggests that a gradient of up to $\sim 0.5 \text{ s}^{-1}$ would still allow the travel time data to be satisfied. In this case the $2.45 \text{ km s}^{-1}$ velocity determined from straight-line fits to the seafloor refraction phase would likely be an average value from within layer 2A and the thickness estimates of layer 2A presented here would not change significantly.

3.4. Layer 2A Seismic Results

The mean layer 2A thickness along the ridge axis (CAM 71) is $400 \pm 130 \text{ m}$ ($n=269, \sigma$). The mean layer 2A thickness is similar for the across-axis lines (Table 1),

![Figure 3. CAM 71 layer 2A model, along the ridge axis. Layer 2A is shown in grey, constructed assuming a constant velocity of 2.45 km s$^{-1}$. Individual thickness determinations shown as white dots, with layer 2A base drawn as a smoothed line through these points. Vertical exaggeration, 1:22. Arrows show intersection positions of current volcanic systems with CAM 71. Letters indicate areas discussed in text.](image-url)
although in both cases the spatial mean may be slightly larger than our observations since refractions from beneath a thicker layer 2A are less likely to be observed on the 2.4 km streamer. These thicknesses were calculated assuming a constant velocity within layer 2A over the whole area. If the velocity within layer 2A increases with crustal age (which our data have not resolved), then the layer 2A thickness must decrease correspondingly with crustal age.

The base 2A refraction velocity is higher in the older crust than it is along axial line CAM 71, with a trend of increasing velocity away from the axis (Figure 4). While the base 2A refraction velocity for zero age crust is 3.3 ± 0.3 km s⁻¹, that for 1.5 Ma crust is 4.0 ± 0.4 km s⁻¹. The rate of increase of velocity with crustal age is greater than that reported by Greverneyer and Weigel [1996], who suggest on the basis of a data compilation that the evolution of upper crustal velocities with age is complete in ~10 Ma rather than the 40 Ma suggested by Houtz and Ewing [1976]. The rate of increase in base 2A velocity with crustal age found here is, however, similar to the rate of increase in layer 2A velocity on the Juan de Fuca Ridge which, from stacking velocities, was found to rise from 3.0–3.5 km s⁻¹ on the axis to 5.0 km s⁻¹ at 1.2 Ma [Rohr, 1994].

Streamer refractions were not received where the water depth was greater than 1200 m, away from the ridge axis, and therefore direct velocity measurements for layer 2A could not be made in these areas. However, the CD70 DOBH data from the smallest offsets (section 4) yielded a base 2A velocity together with an upper-crustal travel time delay. The time delay attributable to layer 2A decreases by a factor of 4 between zero age and 5 Ma crust. With the present data we are unable to resolve the relative contributions of a layer 2A velocity increase or a layer 2A thickness decrease to this delay time change. A similar decrease in upper crustal delay time with age has been noted elsewhere on the Reykjanes Ridge [Lilwall et al., 1980; Mochizuki, 1995; Sinha et al., 1997], although the geometry of these surveys was such that it was not possible to isolate the contribution of layer 2A specifically.

As noted above, the streamer refraction data give the base 2A refraction velocity for zero age crust as ~3.3 km s⁻¹ and that for 1.5 Ma crust as ~4.0 km s⁻¹. The DOBH data constrain the base 2A velocity at 5 Ma to be ~4.5 km s⁻¹. Such an increase in upper crustal velocities away from the ridge axis has been widely reported [e.g., Greverneyer and Weigel, 1996] and is attributed to mineral growth and deposition within the pore space [e.g., Wilkens et al., 1991].

### 3.5. Magnetization of the Uppermost Crust

The extrusive basalts of young oceanic crust typically exhibit high magnetic intensity [de Boer, 1975; Tivey, 1996], although this may be rapidly reduced as titanomagnetite, the primary phase bearing the remanent magnetization, undergoes low-temperature oxidation [Klitgord, 1976]. The central anomaly magnetic high (CAMH), a short-wavelength anomaly peak found over the spreading axis of many mid-ocean ridges, is commonly used to infer the area of most recent volcanism on spreading center axes. The CAMH is not observed universally; an axial magnetization low within the CAMH on the Endeavour Ridge axis of the northern Juan de Fuca Ridge is thought to be caused by thinning of the magnetic source layer (i.e., the extrusive section) [Tivey and Johnson, 1993]. Positive correlations between the magnetic source layer and the thickness of seismic layer 2A have also been reported from the Cleft and Vance segments of the southern Juan de Fuca Ridge [Tivey and Johnson, 1993; Tivey, 1994], and from the EPR.

Gee and Kent [1994] suggest a resolution of the paradox between the off-axis thickening of seismic layer 2A on the EPR and the presence of a CAMH: the alteration process causing magnetization to decay occurs over as

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**Table 1. Layer 2A Seismic Observations**

<table>
<thead>
<tr>
<th>Line</th>
<th>Layer 2A Thickness, m</th>
<th>Base 2A Refraction Velocity, km s⁻¹</th>
<th>n</th>
</tr>
</thead>
<tbody>
<tr>
<td>CAM 71</td>
<td>398 ± 132</td>
<td>3.29 ± 0.46</td>
<td>269</td>
</tr>
<tr>
<td>CAM 72</td>
<td>378 ± 91</td>
<td>4.03 ± 0.74</td>
<td>178</td>
</tr>
<tr>
<td>CAM 74</td>
<td>387 ± 113</td>
<td>4.13 ± 0.85</td>
<td>133</td>
</tr>
<tr>
<td>CAM 75</td>
<td>376 ± 113</td>
<td>3.87 ± 0.77</td>
<td>237</td>
</tr>
</tbody>
</table>

The errors given are 1σ. The mean layer 2A thicknesses are identical within uncertainty, whereas the base 2A refraction velocity is higher off the ridge axis.

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**Figure 4.** Base 2A refractor velocity, v, across the ridge axis for (a) CAM 72; (b) CAM 75; and (c) CAM 74. Points show the velocities corrected for seafloor dip. Large dot in center of each plot is the mean velocity for CAM 71, with a 2σ bar. Data are scattered but show an increase in refractor velocity with mean crustal age, particularly CAM 72 NW, CAM 75 SE, and CAM 74 NW.
Figure 5. CAM 71 magnetic models. Top section of each box shows observed (thick) and modeled (thin) magnetic anomaly. Bottom section shows layer 2A model. (a) Constant thickness source layer, magnetization 25 A m\(^{-1}\); (b) using seismically determined layer 2A as magnetic source layer, constant magnetization, 25 A m\(^{-1}\); amplitude variations are fairly well matched except at 60–65 km; (c) same thickness model but variable magnetization (shown below). Arrows show intersection positions of current volcanic systems with CAM 71. These match the areas of increased magnetization. Dashed lines show the intersection of CAM 71 with (left to right) CAM 72, CAM 75 and CAM 74.

3.6. Magnetization Along CAM 71

The base of layer 2A is commonly interpreted as a porosity boundary [Whitmarsh, 1987; Harding et al., 1989; Rohr, 1994]. Furthermore, in young (<1 Ma) oceanic crust on fast spreading ridges, the wide-angle seismic reflections from the steep velocity gradient at the base of layer 2A are widely interpreted as originating from the extrusive/sheeted dike contact. We therefore use our seismic layer 2A model along the ridge axis to examine the effect of a variable thickness magnetic source layer, assuming that seismic layer 2A is equivalent to the extrusive section.

We here model just the zero age profile, along the spreading axis (CAM 71), thus avoiding complications...
due to the magnetic anomaly caused by magnetic reversals, by age-related changes in remanent and induced magnetization, by variations in spreading history, and by the detailed geometry of crustal accretion. The CAMH in the CD70 area shows large-amplitude variations along strike, which are likely to be at least partly due to variations in the source layer thickness.

If the magnetic layer is assumed to have a constant thickness of 400 m and a constant magnetization, the predicted magnetic anomaly does not match the observed anomaly (Figure 3a). Therefore there must be a variation in layer thickness and/or magnetization to match the variations in the data. When we use seismic layer 2A along CAM 71 to define the thickness variations in the magnetic source layer but maintain the same constant magnetization of 25 A m\(^{-1}\), the model produces a magnetic anomaly variation which reproduces the observed variations of amplitude and has an RMS misfit of 530 nT (Figure 5b).

In the preliminary models, we chose a constant magnetization of 25 A m\(^{-1}\) based on natural remanent magnetization (NRM) measurements from Reykjanes Ridge samples of 10–60 A m\(^{-1}\) from 61° to 62°N (de Boer, 1975), and a mean of 18.8 ± 12.5 A m\(^{-1}\) from slightly farther north (Sempéré et al., 1990). However, weathering may reduce the magnetization (Gee and Kent, 1994), while increased fractionation at crustal levels may increase the magnetization (Sempéré, 1991). We therefore show a third model allowing variable magnetization within layer 2A which improves the fit further (Figure 5c). The magnetization within layer 2A was estimated by first running a constant layer thickness inversion for magnetization (Parker and Heustis, 1974) and then scaling this value by the thickness of the model layer 2A along the line. This forward model magnetization distribution gives a magnetic anomaly that matches all the main features of the data and has a much improved RMS misfit of 250 nT (Figure 5c).

Tests with constant layer-thickness models suggested that the three dimensionality of the seafloor topography could affect the magnetization solution by up to ~10%. Given these possible effects of three-dimensionality in the structure, and the uncertainties in the layer 2A thickness, further refinement of the magnetisation distribution model would have rather limited value. In this model we did not include any annihilator (a magnetisation distribution that produces no external field), as for this along-axis line the technique of adding annihilator to balance magnetization amplitudes at known reversal boundaries could not be applied. Instead, the magnetisation solution was offset by a DC level to minimize the misfit in the field. The solution we present in Figure 5c is not the only magnetisation distribution that could fit the data, as magnetic inversion is inherently nonunique, although it perhaps represents the most plausible situation.

The areas of enhanced magnetization correspond to the points at which the current volcanic systems identified from backscatter data in the neovolcanic axis (Figure 1) intersect CAM 71 (arrows, Figure 5c). It is likely that the areas of particularly high magnetization are the areas in which volcanic activity has been most recent. Our successful modeling of the observed magnetic field using the seismically measured layer 2A as a magnetic source layer and the correlation between areas of enhanced magnetization with regions of recent eruptive activity support our assumption of the equivalence between seismic layer 2A and the extrusive layer. This correlation is widely accepted for the young crust on the EPR, but this is the first meaningful report of such a correlation on the MAR.

Having established by the magnetic modeling that seismic layer 2A is likely to be coincident with the volcanics along the ridge axis, we now return to discuss the observed thickness variations of layer 2A along the ridge axis. It exhibits a fairly constant thickness along the northeastern part of CAM 71 (Figure 3). At 60 km the base of layer 2A fingers up, perhaps indicating a bulk of intrusive rocks associated with dikes feeding the eruptive fissure that is shown by backscatter data (Applegate and Shor, 1994) to run along the axis of the overlying AVR (area A, Figures 1 and 3). Layer 2A thickness increases to its maximum constrained thickness of ~600 m at a distance of 45–55 km, flanking the largest AVR in the survey region (area B, Figures 1 and 3). Between 20–40 km, layer 2A has below average thickness. The thin region is associated with a narrow AVR that barely intersects CAM 71 (area C, Figures 1 and 3), having a poorly defined recently active volcanic system with little recent extrusive activity. Centered on 20 km distance is a graben, where an AVR has recently been rifted into two halves (area D, Figures 1 and 3), the active volcanic vent being still active or reinvigorated within the graben. The layer 2A thickness determinations here are consistent with an extrusive graben fill being dropped down by normal faults on each side.

4. Crustal Wide-Angle Seismic Data

We modeled the crustal structure along the four seismic profiles using data recorded on five DOBHs and five disposable sonobuoys. Crustal models were constrained by travel time forward and inverse modeling and synthetic seismogram forward modeling and were confirmed by gravity modeling. The disposable sonobuoys recorded upper crustal phases to offsets up to 20 km. We used Bruguer and Minshull's (1997) method to cope with sonobuoy drift in the modeling.

Four of the DOBHs each recorded two orthogonal seismic lines and were located near the intersection of these lines. The fifth instrument, DOBH 11, recorded only a single line. There were therefore nine seismic sections (Figures 6–9).

4.1. DOBH Data and Travel Time Picks

The data recordings from the five DOBHs are discussed in sections 4.1.1–4.1.4 according to seismic line, CAM 71–CAM 74 (Figure 1). Sample synthetic seismogram sections are plotted adjacent to their corresponding data sections (Figures 6–9). Several of the sections exhibit strong receiver seafloor multiples of the ground waves.
Figure 6. CAM 71 crustal modeling. Seismic data are scaled linearly with range and are zero phase filtered at 5–13 Hz. (a) DOBH 13 data with model travel times; (b) DOBH 13 synthetic seismograms; (c) DOBH 14 data with model travel times; (d) CAM 71 ray tracing through crustal velocity model; (e) CAM 71 crustal velocity model, selected velocity contours annotated in km s\(^{-1}\).

Uncertainties on the pick times of the crustal turning phases are estimated as 40 ms at small offsets, increasing to 75 ms at 30 km offset, while the \(P_nP\) picks were allocated an uncertainty of 100 ms. CAM 72, CAM 73 and CAM 74 yielded picks to offsets sufficient to check satisfaction of reciprocity.

4.1.1. CAM 71 (Figure 6). The shortest offset turning rays on DOBH 13 can be correlated with the base 2A refraction observed on the streamer: they curve with offset out to about 10 km, indicating a velocity gradient in the upper crust. Arrivals with an apparent velocity of \(\sim 7\) km s\(^{-1}\) persist out to an offset
Figure 7. CAM 72 crustal modeling. Seismic data are scaled linearly with range and are zero phase filtered at 5–13 Hz. Position of ridge axis is indicated. (a) DOBH 14 data with model travel times; (b) DOBH 15 data with model travel times; the travel time curve paralleling the first arrivals at offsets below 20 km is for an S wave converted at the midcrustal interface, with \( v_p/v_s = 1.81 \); (c) DOBH 15 synthetic seismograms; (d) CAM 72 ray tracing through crustal velocity model; (e) CAM 72 crustal velocity model, selected velocity contours annotated in km s\(^{-1}\).

of -55 km on DOBH 14. The furthest offset arrivals on both DOBHs are interpreted as the Moho reflection, \( P_mP \), arriving almost concurrently with, but just behind, the crustal diving phase \( P_g \). At smaller offsets, \( P_mP \) is present \( \sim 0.5 \) s behind the primary at \( \sim 25 \) km offset and becomes indistinguishable from \( P_g \) at \( \sim 45 \) km offset. Converted shear phases are not clear on either of these ridge axis sections, in contrast to the off-axis profiles.

4.1.2. CAM 72 (Figure 7). Both DOBHs recorded arrivals almost from end to end of the 76 km line. When the source was over sedimented areas, converted
shear waves were recorded on DOBH 15, which was deployed on sediment, in contrast to DOBH 14, which sat directly on igneous acoustic basement. We interpret an arrival paralleling the first arrival at offsets below 20 km on DOBH 15 to be a shear wave converted at the mid-crustal interface. On both DOBHs, the $P_mP$ phase is recorded behind the crustal diving phase, with an amplitude comparable to the first arrival, from offsets of ~22 km, until the phases merge beyond 40 km. The $P_mP$ phase is particularly clear in the multiple.
4.1.3. CAM 73 (Figure 8). The three record sections parallel to the ridge axis are of similar character, each showing crustal phases to offsets of ~40 km and a less distinct \( P_mP \) reflection than the other lines. A reflected phase cutting across crustal turning phases on DOBH 15 (~25 km) and DOBH 12 (~30 km) is tentatively interpreted as \( P_mP \), although this was not picked for travel time modeling. We attribute the indistinct character of the \( P_mP \) reflection to the presence of a smaller impedance contrast across the Moho than on the other lines.

4.1.4. CAM 74 (Figure 9). As on CAM 72, the
topography of the ridge affects the apparent velocities observed on CAM 74. DOBH 12 recorded $P_mP$ at -30 to -35 km. DOBH 13 recorded crustal diving rays $P_d$ out to the end of the line beyond 60 km range. $P_mP$ merges with $P_d$ at offsets around 40 km.

4.2. Travel Time Modeling

For each line, a velocity model was set up using water depths from the echo sounder measurements and sediment thickness from the stacked MCS sections. The upper crustal structure deduced from the streamer data was included where available. With two DOBHs on each line which recorded arrivals to significant offsets, supplemented by upper crustal arrivals recorded by the sonobuoys, reversed ray coverage was achieved for at least the portion of each line between the DOBH positions (Figure 1).

The wide-angle seismic travel time picks from the sonobuoys and DOBHs were modeled using Zelt and Smith's [1992] algorithm applied to a two-dimensional model of the crustal $P$ wave velocity structure. We built the model from the top downward, using the minimum structure that would satisfy the data within the estimated travel time uncertainties.

The first-arrival travel time data were initially broadly satisfied by a model typical of oceanic crust, if slightly thicker: an upper crustal layer with a high vertical velocity gradient ($\sim$0.5–1.0 s$^{-1}$) and a lower crustal layer with a lower velocity gradient ($\sim$0.1 s$^{-1}$). Increasing 2-D structure was introduced as forward and inverse travel time modeling was continued simultaneously for velocity and interface depth until the travel time data were satisfied; that is, the $\chi^2$ measure and the RMS misfit were reduced as much as possible within the model parameterization (Table 2). The modeling was carried out as four separate lines rather than simultaneously on all four intersecting lines for three main reasons. First, any velocity anisotropy between the ridge-parallel and ridge-perpendicular lines would go undetected and would reduce the validity of the models. Second, since the lines intersected at the same position as the DOBHs, the resolution on velocity nodes beneath each DOBH is rather lower than that for nodes between the DOBHs at midcrustal and lower crustal depths. Third, a narrow velocity anomaly at the ridge axis, for instance, would pull the velocities on the crosslines away from a better broad model.

4.3. Synthetic Seismograms

We calculated synthetic seismograms using an asymptotic ray theory code TRAMP [Zelt and Ellis, 1988]. Inclusion of an additional, midcrustal layer in the model allowed a smoother decrease in amplitude with offset (Figure 6b) and allowed fitting of second arrival crustal phases (Figure 9b).

The $P_mP$ reflected phase is present in the majority of the data sections. Although this phase is coherent, its amplitude is mostly below the threshold required for picking. However, an excellent agreement is apparent when calculated model travel times are overlaid on data from which picks were not made. The validity of the Moho position is thus confirmed. The sub-Moho velocity is not constrained by the travel time modeling since a mantle refracted or turning phase was not observed unambiguously in any of the data, perhaps due to an extremely low velocity gradient in the mantle or a high level of attenuation in the mantle. Synthetic seismograms were therefore computed to estimate the mantle velocity. A velocity of 7.8±0.2 km s$^{-1}$ (i.e., an increase in velocity of 0.5 km s$^{-1}$ across the Moho) provided synthetic seismograms that gave, qualitatively, a good match for the extent and amplitude of $P_mP$.

4.4. Uncertainty of Model Parameters

The uncertainty of the parameters was estimated by perturbing the velocities and depths in the model until the fit to the data was statistically degraded [Zelt and Smith, 1992]. For CAM 71, for example, this method suggests that the uncertainty on the Moho depth is ±0.25 km based on the degradation of the fit to the $P_mP$ phase only. Lower crustal velocities above 7 km s$^{-1}$ are not sampled directly by turning phases on CAM 71 (Figure 6d), although such velocities are sampled on other lines. The preferred CAM 71 model is based on extrapolation of the velocity gradient to the $P_mP$ reflector. If, alternatively, the velocity gradient decreases, then the depth to the reflector shallows by up to 0.1 km (for 7 km s$^{-1}$ continued to the Moho). A similar uncertainty should perhaps be included for the possibility of an increase in velocity gradient, leaving the final estimated uncertainty as ±0.5 km.

In conclusion, in the regions of the model that are constrained by ray coverage (see Figures 6–9), estimated errors on velocity are ±0.3 km s$^{-1}$ in the upper crust, ±0.2 km s$^{-1}$ for the midcrust, from which the bulk of picked arrivals turn, and ±0.3 km s$^{-1}$ for the lower crust. Estimated errors on the depth of the midcrustal interfaces are ±0.4 km, and estimated errors on the Moho depth are ±0.5 km.

5. Gravity Modeling

The mean velocities for each layer of the crust in our seismic models were converted to densities using Carlson and Herrick's [1990] relationships, except for layer 2A, where the velocities fall well beneath the region.
Table 3. Parameters of the Mantle Thermal Model

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spreading rate, mm yr⁻¹</td>
<td>10.0</td>
</tr>
<tr>
<td>Thermal diffusivity κ, m² s⁻¹</td>
<td>10⁻⁶</td>
</tr>
<tr>
<td>Mantle reference density (STP), Mg m⁻³</td>
<td>3.30</td>
</tr>
<tr>
<td>Thermal expansion coefficient α, °C⁻¹</td>
<td>3 x 10⁻⁵</td>
</tr>
<tr>
<td>Potential temperature of asthenosphere, °C</td>
<td>1350</td>
</tr>
<tr>
<td>Temperature at base of lithosphere, °C</td>
<td>1100</td>
</tr>
</tbody>
</table>

constrained by Carlson and Herrick [1990]. A density of 2.4 Mg m⁻³ was assigned to layer 2A, using results from the EPR [Stevenson and Hildebrand, 1994].

A good fit between observed and modeled gravity assuming a two-dimensional structure along each of the profiles is obtained without any modification to the crustal structure derived from seismics (Figure 1). The short-wavelength misfits of a few milliGals may be ascribed to out-of-plane seafloor and layer 2A topography or to local density variations, particularly in the upper crust. There is sufficient scatter in the velocity-density relationship to allow these small variations to be fitted perfectly within the density uncertainties if so desired.

At spreading centers it is necessary to remove the effect of the age-dependent thermal (and hence density) structure of the lithosphere. We use a half-space lithosphere cooling model to calculate the gravitational contribution from first-order thermally caused density variations and add this to the gravity signal from the seismically determined crustal model. We assume that the top 2 km of the crust are effectively quenched at the ridge axis by hydrothermal circulation and therefore do not give a time-varying gravity signal. Parameters used in the half-space cooling model are shown in Table 3. The asthenospheric temperature of 1350°C was chosen to produce 9.5 km of igneous crust on decompression melting [from Bown and White, 1994]. In the immediate vicinity of the ridge crest the half-space cooling model isotherms were adjusted by imposing a conductive temperature gradient through the lower part of the crust (2-11 km depth) in accordance with the observation of a brittle crust to 7-8 km depth at the axis [Mochizuki, 1995]. The shape of the gravity contribution resulting from the lithosphere cooling is insensitive to reasonable changes in the mantle thermal parameters: a 10% change in any of these parameters only changes the gravity curve by a few milliGals over 60 km.

Along and parallel to the ridge axis the gravity field shows only minor variations within the survey area (Figure 1 inset). Although the crustal thickness is only constrained seismically in the center of the seismic lines, the fit to the gravity data supports the assumption of a constant thickness crust parallel to the ridge axis. The residual mantle bouguer gravity anomaly along the Reykjanes Ridge does not show the "bull's eye" pattern reported from some other slow spreading ridges but has a smooth character. An explanation for the absence of along-axis crustal thickness variations is that the anomalously large thickness of oceanic crust causes the lower section to lie in the ductile regime; thus any crustal thickness variations that would otherwise be frozen in at the spreading center are removed by lower crustal flow [Bell and Buck, 1992].

The isostatic balance (Figures 10b and 10d) shows that the topography of the elevated ridge axis, with its thickened crust, can be supported by the mass deficit in the mantle suggested by the simple lithospheric thickening model. The residual pressure at a depth below the base of the lithosphere is smaller than 5 MPa. The model presented here does not require that the axial high is supported by the strength of the lithosphere, although a small flexural topographic and gravity "moat" edging the axial high is expressed in Figure 10 as a small negative excursion below the isostatic equilibrium line.

![Figure 10. Gravity models and isostatic balance. Mean upper and lower crustal densities converted from best fitting seismic models using Carlson and Herrick's [1990] velocity-density relations (see text). (a) CAM 72 observed and modeled free-air gravity, with and without mantle thermal contribution; (b) isostatic pressure variation, arbitrary zero level; (c) CAM 74 observed and modeled free-air gravity, with and without mantle thermal contribution; (d) isostatic pressure variation, arbitrary zero level.](image-url)
In conclusion, the gravity modeling provides support for the across-axis crustal thickness variation postulated from the seismic modeling, does not require crustal thickness variations parallel to the ridge axis and suggests that the axial high is supported isostatically.

6. Discussion of Crustal Models

The velocity structures of 0 and 5 Ma crust are distinctly different, as shown by average velocity-depth profiles from CAM 71 and CAM 73 (Figure 11). A major difference is that the seismic velocity of 5 Ma crust is higher at all depths through the model than the velocity at the ridge axis. The steep (~1 s⁻¹) vertical velocity gradient of the upper crust is characteristic of oceanic crust and can be understood in terms of increasing closure of cracks as lithostatic pressure increases with depth. In the upper crust the increase in velocity with age is probably caused by the infilling of pores and cracks by precipitation of hydrothermal minerals. However, the agreement between laboratory and seismic measurements of gabbro velocities [Carlson and Miller, 1997], and the low vertical velocity gradient in the lower crust, suggests that porosity is low in the lower crust. Therefore porosity changes are unlikely to cause the velocity change that we observe. A more likely cause is that higher temperatures at the ridge axis reduce seismic velocities. In the absence of a magma chamber, the thinner lithosphere at the ridge axis will have a higher conductive temperature gradient than the older off-axis lithosphere. Estimating the lithospheric thickness increase to be ~25 km over 5 Myr and extrapolating a conductive gradient downward from the base of the seismogenic layer to the base of the crust at the ridge axis suggests a difference in temperature of more than 500°C between the base of the crust at 0 and 5 Ma. Velocity-temperature relations from Murase and McBirney [1973] and Khitarov et al. [1983] predict that this temperature difference might cause a difference in P wave velocity of ~0.5 km s⁻¹ at the base of the crust, decreasing upward. Our observations are of the right order of magnitude (Figure 11), falling within the uncertainties of the laboratory results. We therefore infer that a temperature effect is the likely explanation for the velocity differences observed in the lower part of the velocity models. A similar increase in lower crustal velocity with age is reported by Bunch and Kennett [1980].

Another difference is that the crustal thickness decreases from 10.0 km on the ridge axis to 7.8 km on 5 Ma crust. The change in crustal thickness of 2.2 km has an uncertainty of about ±1 km. Assuming that the igneous crust is generated by decompression melting of passively upwelling mantle beneath the spreading axis, such a change in crustal thickness would be produced by a change in mantle temperature of about 35°C [Bown and White, 1994]. There is evidence that a pulse of hot asthenospheric mantle has migrated southward through the CD70 area from bathymetry [Murton and Parson, 1993; Keeton et al., 1997], from seamount distribution [Magde and Smith, 1995], from the petrology of dredged rocks [Taylor et al., 1995], and from rare earth element inversions [White et al., 1995]. Our seismic results support the thesis that the crustal thickness change is due to a change in mantle temperature.

Profile CAM 73, over 5 Ma crust, lies in the gravity and bathymetry low between the present axis and the first V-shaped ridge. We conclude that it is likely that thickened crust is present beneath all the V-shaped ridges and that these record variations of around ±30°C in the temperature of the mantle fed southward beneath the Reykjanes Ridge from the Iceland mantle plume [White et al., 1995; White, 1997].
The 10 km thick igneous crust currently being formed on the ridge axis falls within a trend of decreasing thickness southward along the Reykjanes Ridge away from the Iceland plume [Smallwood et al., 1995]. If the melt is generated by passive mantle decompression, the mantle potential temperature in the CD70 survey area is 1360°C [Bown and White, 1994]: this too falls in a pattern of abnormally hot mantle introduced by the flow field associated with the Iceland mantle plume [White et al., 1995; White, 1997].

We do not find any evidence for a crustal magma chamber in our survey area, in common with the great majority of other seismic surveys over slow spreading ridge axes. The only well-documented slow spreading ridge magma chamber to date is from farther south on the Reykjanes Ridge at 57°43'N [Sinha et al., 1997].

Thermal estimates suggest that a magma body such as that discovered at 57°43'N, while containing enough magma to form 20,000 years worth of crust, will freeze in about 1000 years. If it is valid to extend this estimate spatially, then it is reasonable to expect to find an extant magma chamber under only 1 in 20 AVRs. The absence of a magma chamber in the CD70 area is thus not surprising.

7. Summary and Conclusions

Thickness variations of seismic layer 2A have been determined using streamer refraction travel time modeling where the water depth is less than ~1200 m. The mean velocity and thickness of layer 2A are 2.4 ± 0.3 km s⁻¹ and 400 ± 100 m, respectively. The seismic velocity at the base of layer 2A averages 3.3 ± 0.3 km s⁻¹ on the ridge axis, increasing with age to ~4.0 km s⁻¹ at 1.5 Ma and to ~4.5 km s⁻¹ at 5 Ma. An off-axis decrease in delay time through the upper crust is observed on the DOBHs, suggesting either that the velocity within layer 2A roughly doubles in 5 Myr or, more likely, that there is an increase in velocity and a decrease in layer thickness.

Along the spreading axis, the signature of a magnetic layer corresponding to the seismically measured layer 2A provides a good match to magnetic anomaly data, particularly when magnetization within the layer is enhanced at the sites of the most recent volcanism. We therefore suggest that seismically measured layer 2A is coincident with the extrusive layer on the youngest oceanic crust.

The crustal velocity structure at 61°–62°N on the Reykjanes Ridge is generally typical of oceanic crust, although the crust is anomalously thick, which we attribute to the presence of anomalously hot asthenospheric mantle beneath the spreading center. Upper crustal velocities increase with crustal age, which we attribute to precipitation of hydrothermal minerals in the pore-space. Lower velocities on the ridge axis than in 5 Ma crust, particularly in the lower crust, are attributed to higher axial crustal temperatures: there is no evidence for a crustal melt body in the CD70 area.

The crust is 10.0 km thick on the ridge axis and thins to 7.8 km at an age of 5 Ma. We attribute the change in crustal thickness to a temporal change in asthenospheric temperature beneath the spreading center: passive upwelling of mantle with a potential temperature of 1360°C produces 10.0 km of crust by decompression melting, while a mantle potential temperature of 1325°C will yield 7.8 km. We postulate that somewhat thickened crust is present beneath the gravity and bathymetric highs of the V-shaped ridges that flank the Reykjanes Ridge axis.

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