Crustal structure beneath the Faroe Islands from teleseismic receiver functions

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SUMMARY
We use teleseismic receiver function analysis to constrain the crustal structure beneath the Faroe Islands on the northwest European volcanic continental margin. A cluster of 45 broad-band seismometers on the Glyvursnes peninsula, Streymoy Island recorded 10 teleseismic events ranging in magnitude from 6.1 to −8.0 during July–December 2003. Receiver functions show a clear $P_s$ peak for the first time in this region. The depth of the converting boundary is estimated as 29–32 km using forward and inverse modelling and thickness versus $V_P/V_S$ ratio stacking techniques. Modelling experiments suggest that this estimate may represent the conversion from a region of high-velocity lower crust rather than the crust–mantle transition at the Moho. The best-fit modelling results were achieved with a gradational high-velocity region at least 6 km thick in the lower crust. This is interpreted as due to the emplacement of sills into pre-existing continental crust rather than the simple underplating of a block of high velocity igneous material at its base.

Key words: Continental margin; Crustal structure.

1 INTRODUCTION
The Faroe Islands are a presumed fragment of continental crust on the northwest European continental margin, but there is currently no measurement of the crustal thickness or the lower crustal velocity directly beneath them. Our first objective was to use receiver functions from a cluster of 45 broad-band seismometers to constrain the crustal thickness beneath the Faroe Islands. The islands are covered by flood basalts formed at the time of the Tertiary continental break-up at ca. 55 Ma, which created the North Atlantic (White & McKenzie 1989). The basalt flows reach at least 7 km thick (Waagstein 1988; White et al. 2003; Passey & Bell 2007). No outcrops of underlying crust pre-dating the basalts have been found on the islands, however, plate reconstructions (e.g. Nunns 1983; Knott et al. 1993) suggest that the region from East Greenland to the Shetlands was continuous prior to breakup. Basalt sequences can be correlated from east Greenland to the Faroe Islands (Larsen et al. 1999). Outcrops of Lewisian basement in East Greenland and the Shetlands suggest that the same basement is likely to be found beneath the Faroes if continental lithosphere lies beneath the basalt. A second objective of our study was to use receiver functions to investigate whether there is continental crust beneath the Faroe Islands.

Lewisian basement has a typical $P$-wave velocity of 6.0–6.5 km s$^{-1}$ (Hall & Simmons 1979). When there are large volumes of extruded basalts, it is likely that there is at least as much magma intruded into the lower crust as extruded (Cox 1980; White et al. 2008). The intruded, fractionated magma has a much higher seismic velocity than the country rock, and therefore creates a high-velocity lower crust (HVLC) with velocities typically in excess of 7.0 km s$^{-1}$. Our third objective was to seek evidence for an HVLC indicative of lower crustal intrusion.

Bott et al. (1974) provided an early estimate of the crustal thickness beneath the Faroe Islands of 27–38 km. Later authors (Richardson et al. 1998, 1999; Smallwood et al. 1999), using offshore shooting into seismometers on land in the Faroes on the FLARE, FAST and FIRE profiles (locations marked in Fig. 1), suggest that the crustal thickness is likely to be towards the upper end of this range. Staples et al. (1997) estimated the crustal thickness beneath the Faroe Islands as 35–40 km, of which they interpreted 20–30 km as original continental crust and the remainder as intruded or extruded igneous rock. Slightly thinner, but not dissimilar crustal thicknesses of 27–32 km are reported from the nearby submarine areas of Hatton Bank (White et al. 1987; Fowler et al. 1989), Rockall Bank (Bunch 1979) and the Fugloy Ridge (White et al. 2008). As the Faroe Islands are subaerial, we would expect the crustal thickness beneath them to be larger than on the adjacent, submarine portions of the continental margin if they are in isotatic equilibrium.

Regions of HVLC are ubiquitous on the North Atlantic continental margins (e.g. Fowler et al. 1989; Morgan et al. 1989; White...
The HVLC has often been referred to as ‘underplated’ igneous rock, though as White et al. (2008) discuss, it is more likely to be a region of heavily intruded continental crust. Comparisons of receiver function estimates of crustal structure around the British Isles with wide-angle controlled source seismic constraints have been published by Tomlinson et al. (2006) and Shaw Champion et al. (2006). They report similar results from the two methods when the crustal velocity is known a priori, but also show that it can be difficult to resolve the presence of an HVLC unless it is more than 4 km thick. For at least some of the locations, Tomlinson et al. (2006) suggest that the converting horizon from which the thickness is calculated in the receiver function analysis represents the top of the HVLC rather than the Moho. The receiver function thickness is therefore smaller than the true crustal thickness. Later in this paper we discuss the effect of an HVLC on the receiver functions from velocity profiles appropriate for the Faroe Islands. Shaw Champion (2005) attempted to use data from a single seismometer on the Faroe Islands to calculate receiver functions, but the data were noisy and they did not find a consistent converted arrival from the base of the crust. In our study we have used a tight cluster of 45 broad-band seismometers on Streymoy, Faroe Islands (Fig. 1), which after stacking produces receiver functions with good signal-to-noise ratios and a clear converted arrival from the lower crust.

2 RECEIVER FUNCTIONS

Receiver functions represent the seismic response beneath a seismometer to an incoming teleseismic P-wave. P-to-S conversions are generated at any interface with a significant velocity contrast (Fig. 2). The incident P waves have a steep angle of incidence (in this case 16–29° from the vertical for the direct P-wave and 8–16° from the vertical for their corresponding Ps conversions), hence the particle motion is polarized so that P waves dominate the vertical component and S waves are preferentially recorded on the radial
component. The source time function of the event is removed from the recorded teleseismic waveform by deconvolution, as described by Ammon (1991), in an attempt to isolate the conversions due to local structure from changes to the seismic waveform caused by structure near the source and during propagation through the Earth’s mantle between the source and the receiver.

We calculate receiver functions in the frequency domain using codes by Ammon (1991), which are modified from Langston (1979). A water-level method developed by Clayton & Wiggins (1976) is used to fill spectral troughs and thereby produce stable receiver functions. A Gaussian low-pass filter was used during deconvolution to remove high-frequency noise. The filter is given by

$$G(\omega) = e^{-\omega^2/4a^2},$$

where $a$ is a width factor of the filter. After preliminary trials with different values, $a$ was set to 2.0. This produces a 90 per cent cut-off at 1 Hz on the high frequency side, to match the energy of the teleseismic arrivals, which is predominantly below this frequency. The wavelength at 1 Hz is 6–7 km, so we do not expect to be able to resolve lower crustal layers thinner than about one half-wavelength, or 3–4 km in vertical thickness.

We assume that the velocity structure is 1-D (i.e. there are no significant dips or lateral variability in the crustal structure over the region extending approximately 15 km from the seismometer cluster that is sampled by the different primary and multiple phases used in the analysis and shown in Fig. 2.) In the absence of noise or lateral variability there should be no energy on the tangential component of the receiver function. As we show later, the tangential receiver functions are much smaller than the radial receiver functions, giving us confidence in this assumption.

### 3 FORWARD MODELS OF RECEIVER FUNCTIONS

The depth to a converting boundary inferred from a receiver function depends on the $P$-wave velocity of the crust above that boundary and on the $V_P/V_S$ ratio of the crust. The forward models of receiver functions shown in Fig. 3 illustrate the changes in the traveltimes and amplitudes of the main phases derived from a two-layer model of the crust and mantle with changes in the $V_P/V_S$ ratio of the crust. Multiple peaks, such as the $PpPs$ peak, are delayed differently for these two cases, and it is therefore possible to discriminate between them using the multiple phases. Receiver functions forward modelled from these velocity models were generated using a Gaussian filter with width $a = 2$ to match the inversions made subsequently on the observed seismic data.
3.1 Resolution of high-velocity lower crust (HVLC)

In order to test which features of the HVLC can be resolved using the receiver function method, we generated three types of simple crustal velocity profiles. They represent a simplified form of the actual crustal velocity profile under the Faroe Islands, which has basalts outcropping at the surface and no sedimentary overburden. The type A crustal model contains a discrete layer of HVLC; type B, a gradational region of HVLC; and type C a high-velocity layer with a sharp top and a gradational base. These models (Fig. 4) were then used to generate synthetic radial receiver functions. For each case the thickness $\Delta z$ of the HVLC layer was varied from 0 to -9 km in 3-km steps. All results were compared with the receiver function generated by a control case, a simple two-layer model of the crust and mantle ($\Delta z = 0$) with the boundary at 25 km below the surface (which is also the top of the HVLC in all models).

![Figure 4](https://academic.oup.com/gji/article-abstract/177/1/115/728403)

**Figure 4.** The model types A, B and C represent three possible crustal velocity models generated to investigate the resolution capabilities of the receiver function method. Receiver functions forward modelled from these velocity models were generated using a Gaussian filter with width $a = 2$ to match the inversions made subsequently on the observed seismic data. The ability to distinguish between a layer and a sharp discontinuity ($\Delta z = 0$) is tested by varying $\Delta z$ (0, 3, 6 and 9 km steps are depicted), and between a discrete layer and gradational region by comparing type A with type B and C receiver functions. The dominant phase conversion horizon in all these models is the top of the HVLC rather than the Moho, so depth estimates from receiver function inversions are likely to constrain the depth to the top of the HVLC rather than the Moho depth unless the HVLC is sufficiently thick to be recognized as a separate layer. Values of $\Delta z$ in figure are in kilometres.
The discrete HVLC (type A) is representative of a layer often depicted on crustal profiles, commonly termed ‘underplate’. The gradational HVLC (type B) may be formed if the lower crust is intruded by sills with the density and/or thickness of sills decreasing upwards away from the Moho. If only the lowermost crust is heavily intruded by high-velocity igneous sills, one might expect a greater contrast between the upper crust and the top of this intruded HVLC than between the HVLC and the underlying mantle below (as represented by type C crust).

Experiments using type A crustal models demonstrate that a high-velocity lower crustal layer less than 6 km thick would be indistinguishable from the control two-layer case on the basis of the $P_s$ peak alone, particularly in the presence of noise. Once the thickness $\Delta z$ of the HVLC is increased above 6 km, two different $P_s$ peaks from the bottom and top of a high-velocity layer can be resolved (Fig. 4a). These peaks have a delay time separation that is proportional to the thickness of the layer and amplitudes that depend on the magnitude of the velocity contrast across each boundary. The HVLC layer can be distinguished from the control case at a much smaller $\Delta z$ using the $P_{pP}$ peak. This has already split into two peaks on the multiple $P_{pP}$ arrivals by $\Delta z = 3$ km.

Receiver functions resulting from a gradational region at the base of the crust are also hard to distinguish from the control two-layer case or from a discrete high-velocity layer using the $P_s$ peak alone where $\Delta z \leq 6$ km (Fig. 4b). The $P_s$ peak does not split into two for greater $\Delta z$ thicknesses, but in this case continues to broaden and to reduce in amplitude. The $P_s$ peak does exhibit a slight increase in delay time as $\Delta z$ increases (by $\Delta z = 9$ km the delay time has increased by ca. 0.6 s). Changing the boundaries in the model from sharp to gradational results in a more obvious broadening of the $P_{pP}$ multiple peak and a decrease in its amplitude. This effect increases with increasing $\Delta z$. As the region thickens the $P_{pP}$ peak is therefore more easily hidden by noise.

The receiver functions generated from the type C profiles show a combination of effects (Fig. 4c). The $P_s$ peak remains similar to the control case in delay time (i.e. this peak is due to conversion from the top of the layer), but it decreases in amplitude and broadens slightly as the thickness $\Delta z$ of the gradational HVLC increases. The $P_{pP}$ multiple develops a distinct shape, a peak (resulting from a reflection and then conversion from the top of the layer), with a shoulder (as a result of the gradational region). The shoulder is again of lower amplitude and would be easily obscured by noise in real data.

This analysis shows the importance of the $P_{pP}$ multiple in constraining the crustal structure, but also the difficulty of discriminating between different models of the HVLC for thicknesses less than ca. 6 km. The receiver functions are governed by the depth of the main layer at which phase conversion occurs, which in the models we have used is dominantly at the top of the HVLC rather than at its base. The inferred crustal thickness from the receiver function method in the presence of an HVLC is likely therefore to be the top of the HVLC rather than the Moho, unless the HVLC is thicker than 3–6 km and can be recognized as a separate layer at the base of the crust.

4 ANALYSIS

4.1 Sources of data

Forty-five Guralp 6TD seismometers (bandwidth 0.03–50 Hz) were deployed during July–December 2003 in a tight cluster mainly within a 400 m by 400 m array on the Glyvursnes peninsula, Streymoy, Faroe Islands (see Fig. 1). Ten teleseismic events selected for receiver function analysis all had magnitudes $\geq$6.1, with angular distances between 30$^\circ$–90$^\circ$ (see Fig. 5 for locations and Table 1 for details). They are clustered in back azimuth with four events near Japan, two in Siberia, three near the Aleutians and one lone event on the Carlsberg Ridge. The small spatial extent of the seismometer cluster meant that it could not be used for array processing, but the seismic data were stacked to improve the signal-to-noise ratio.

4.2 Receiver functions

Receiver functions were generated for each event, with all stations in the cluster stacked. The resultant radial receiver functions show a consistent $P_s$ peak, which is $3.5 \pm 0.1$ s after the direct $P$ arrival (Fig. 6). In an ideal case, with flat isotropic horizontal layers, no energy from conversions should be recorded on the tangential component. Dipping layers, anisotropy and scattered energy can, however, all introduce energy. The tangential receiver functions from individual events show a significant amount of energy, but the uneven distribution of events in back azimuth means that we cannot constrain whether this is due to anisotropy or dipping layers, or is simply noise from local heterogeneity.

Receiver functions for individual events were stacked in clusters of similar back azimuth, resulting in an improvement in the signal-to-noise ratio of the radial receiver functions (Fig. 6b). A clear $P_s$ peak occurs at $3.5 \pm 0.1$ s, with a broader peak at $10.5 \pm 0.4$ s interpreted as the $P_{pP}$ multiple, which becomes more apparent after stacking. It is not possible to identify peaks later than this due to the background noise. The amplitudes of the corresponding tangential receiver functions are greatly reduced by stacking compared to those for individual events, suggesting that the majority of the energy in the latter is a result of scattering or background noise rather than 3-D structure. For this reason a 1-D approximation is henceforth assumed. The stack of receiver functions from all events achieved the best signal-to-noise ratio on the radial component with least energy on the tangential component (Fig. 6c).
5 CRUSTAL STRUCTURE FROM OBSERVED RECEIVER FUNCTIONS

We used three different methods to constrain the crustal structure from the stacked receiver functions. First, we made an inversion using Ammon’s (1991) method. Second, we made a grid search using a $h - V_p/V_s$ stacking method developed by Zhu & Kanamori (2000). Third, we forward modelled the main phases by hand to fit the amplitudes and relative arrival times of the main phases. The difference between these methods is primarily in the way that noise affects the results. The inversion weights all the parts of the waveform equally irrespective of their source, whereas forward modelling by hand enables us to concentrate explicitly just on those phases that we can identify, while ignoring other parts of the waveform that may be noise. In the following section we discuss the overall constraints on crustal structure that can be inferred from the results of applying these three different methods to the same data.

5.1 Inverse modelling

The programs used for inversion modelling are described by Ammon (1991). Ammon’s inversions use a Poisson’s ratio of 0.25 throughout. Modified versions of the programs were therefore created to allow the $V_p/V_s$ ratio to be input into the velocity model layer-by-layer and hence to allow variation with depth for greater geological realism. Controlled source wide-angle seismic profiles were used to generate an approximate starting velocity model, then this was perturbed randomly multiple times and the inversion was completed for each perturbed input model. In Fig. 7 we show the mean velocity-depth profile from these multiple inversions for each of the regionalized input receiver functions (coloured lines in Fig. 7), as well as for the stack of all events (thicker black line in Fig. 7).

$P$-wave velocities of ca. 7.6 km s$^{-1}$, which are typical of mantle rocks are reached at a depth of about 32 km beneath the surface. This also corresponds to a change to a smaller velocity gradient at the same depth (broken horizontal black line in Fig. 7), which therefore makes it a good candidate for the Moho. The precise details of the variations in velocity structure below 32 km are beyond the resolution of the inversion, and the smoothing necessary for the inversion forces boundaries to be gradational in the modelling, so we conclude that the best estimate for the crustal thickness from this method is about 32 km.

The overall structure of the 32-km thick crust is marked by a high-velocity gradient in the upper 10 km, beneath which there is either a decrease in velocity gradient (as shown by inversion of arrivals from the Carlsberg Ridge), or an actual decrease in velocity causing a low-velocity zone (remaining inversions). The velocity gradient increases again in the lowermost quarter of the crust beneath 25 km depth (all inversions show this).

The velocities of 4.5–6.5 km s$^{-1}$ and the high-velocity gradient in the upper 10 km of the crust are consistent with the thick-layered basalt sequence seen everywhere on the Faroe Islands, which is known from outcrop and drilling to be at least 7 km thick. The low-velocity zone beneath this could be real, and would then represent continental crust. It may also be an artifact of the limited bandwidth of the seismometers that may cause artificial negative peaks in the receiver functions which then translate into low-velocity zones in the inversion. Even if the actual velocities do not decrease, but rather exhibit a greatly reduced velocity gradient in the mid-crust, we would still interpret this as due to the presence of continental crust beneath the Faroes, overlain by extrusive basalts. The structure of the Fugloy Ridge derived from wide-angle controlled source seismic profiles (red dotted line in Fig. 7), which lies along strike from the Faroes shows a similar velocity profile that is interpreted by White et al. (2008) as representing continental crust overlain by basalts. Note that the Fugloy Ridge is submarine, and has lost some of the near-surface basaltic section by erosion, so it is not surprising that it now has a thinner overall crustal thickness of 27 km than the subaerial Faroe Islands.

The increase in velocity gradient in the lower 10 km of the crust is exactly what we would expect to be produced by igneous intrusions in the lower crust.

5.2 $h - V_p/V_s$ stacking

This method, after Zhu & Kanamori (2000), stacks receiver functions in the crustal thickness $h - V_p/V_s$ ratio domain. Arrival times of the $Ps$, $PpPs$ and $PpSs + PsPs$ phases are calculated for each receiver function, using the appropriate ray parameter for that event, and across a range of values of $h$ and $V_p/V_s$ ratios. The stack is produced by summing the amplitudes of each receiver function at each of these traveltimes. The advantage of this method is that we only stack signals at the time when we expect arrivals. However, the disadvantage is that in the presence of noise, the signal at the expected traveltimes of the $PpPs$ and $PpSs + PsPs$ phases may be dominated by noise because the arrivals from the teleseismic event are only small in any case; we would then be stacking noise, not signal. The variation of ray parameter from the different events is 0.045–0.069 (Table 1). We experimented with a realistic range of different values of the average crustal $P$-wave velocity and found that the stacks are rather insensitive to the precise value adopted, since they depend mainly on the traveltime difference between the

<table>
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<tr>
<th>Event</th>
<th>Date</th>
<th>Origin time</th>
<th>Location</th>
<th>Depth (km)</th>
<th>Ray Parameter</th>
<th>Backazimuth (^)</th>
</tr>
</thead>
<tbody>
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<td>16/06/2003</td>
<td>22:08:02</td>
<td>Kamchatka, Aleutians</td>
<td>6.9</td>
<td>0.0601</td>
<td>8</td>
</tr>
<tr>
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<td>02:36:53</td>
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<td>6.5</td>
<td>0.0534</td>
<td>22</td>
</tr>
<tr>
<td>3</td>
<td>25/09/2003</td>
<td>19:50:08</td>
<td>Hokkaido, Japan</td>
<td>8.0</td>
<td>0.0532</td>
<td>22</td>
</tr>
<tr>
<td>4</td>
<td>27/07/2003</td>
<td>06:25:32</td>
<td>near SE coast of Russia</td>
<td>6.8</td>
<td>0.0554</td>
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</tr>
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<td>0.0512</td>
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<td>0.0690</td>
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<td>0.0574</td>
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<td>Rat Islands, Aleutians</td>
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<td>0.0576</td>
<td>358</td>
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Figure 6. (a) Radial (left) and tangential (right) receiver functions for each event (see Table 1 for details). (b) Receiver functions, grouped and stacked by back azimuth as annotated. The $P_s$ peak stands out more clearly, noise is reduced and the amplitudes of the tangential receiver functions are reduced significantly by the stack. (c) Stack of all events, which markedly reduces energy on the tangential receiver function. Positive amplitudes are shaded. See Fig. 2 for definitions of $P$, $P_s$ and $PpP_s$ phases.

The maximum value of the stack using our data is for a crustal thickness of 26 km, with a $V_P/V_S$ ratio of 1.88 (Fig. 8), but there is a trade-off in the highest values of $h - V_P/V_S$ shown by the diagonal band across Fig. 8. An average crustal $V_P/V_S$ ratio of 1.88 is unrealistic for the rock types likely to be present beneath the Faroe Islands. Basalt has the highest $V_P/V_S$ ratio of the lithologies present, measured as typically 1.84 from boreholes in the Faroes and from wide-angle seismic measurements (Christie et al. 2006; Eccles et al. 2007, 2008; Bais et al. 2008), while the other extreme is the continental basement rock of Lewisian Gneiss, which may have a $V_P/V_S$ ratio as low as 1.73 (Christensen 1996). A likely overall crustal average beneath the Faroes is 1.78 (Christensen 1996; Eccles et al. 2008). The red line in Fig. 8 shows the range of inferred crustal thickness of 27–31 km using the two extremes $V_P/V_S$ ratios of basalt and Lewisian Gneiss, with the thickness from the most likely value of $V_P/V_S$ being about 29 km (red dot in Fig. 8). As Tomlinson et al. (2006) have shown from synthetic studies of $h - V_P/V_S$ stacking, any departure from the assumed 1-D crustal velocity structure (for example, a dipping Moho), would affect the delay times and amplitudes of the peaks and lead to an underestimate...
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5.3 Forward modelling

Forward modelling using the reflectivity method (Kennett 1983), was carried out by hand. We aimed to fit the amplitudes and relative timing of the three main peaks produced by the direct $P$, $Ps$ and $PpPs$ phases and where possible to also fit a negative peak that may represent the $PpSs+PsPs$ conversion. The noise level of the signal prevented identifying and fitting any further multiples. The quality of fit of the synthetic seismograms was judged by its agreement with the relative timing, amplitude and width of each observed peak. The latter was particularly useful for multiple phases, for example the width of $PpPs$ arrivals are diagnostic of gradational velocities in the lower crust (Fig. 4). $V_p/V_S$ ratios appropriate for the crust under the Faroe Islands were assumed as described in the previous section.

A preliminary study was carried out using a two-layer model of the crust and mantle. For subsequent studies the minimum number of layers was used that created an adequate fit to the different peaks. Adding a low velocity top layer to the model generated the trough, which commonly follows the direct $P$ peak. This layer represents the basalt flows that outcrop on the island. The $P$-wave velocity of this layer was chosen on the basis of borehole data (Petersen et al. 2006; Bais et al. 2008) and is therefore well constrained. However, the negative lobes around the main arrivals may also be artifacts of the receiver function generation caused by limited bandwidth of the seismometers (0.03–50 Hz). The width of the peaks was poorly modelled by a sharp crust–mantle discontinuity. A gradational velocity increase at the same depth resulted in a definite broadening of the multiple peak, improving the match to the data (as expected from Fig. 4).

The best-fit velocity model produced by forward modelling the stack of all receiver functions is shown in Fig. 9. Similar velocity models were produced for individual receiver functions and suggest that the base of the crust is marked by a gradational region of the order of 4–6 km thick with the Moho at a depth of about 30 km.

6 DISCUSSION

We consistently find evidence from the three methods described above for a crustal thickness of 29–32 km beneath the Faroe Islands. The mid-crustal velocities are consistent with the presence of continental crust similar to that inferred beneath the along-strike continental fragments of Fugloy Ridge (White et al. 2008), Hatton Bank (Fowler et al. 1989; Morgan et al. 1989) and Edoras Bank (Barton & White 1997). This continental crust lies beneath an...
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Figure 9. The best forward-model fit to the receiver function generated from a stack of all events. The shallowest layer in the velocity model represents a basalt layer and has considerable impact on the first few seconds of the receiver function. There is a good match to the delay time, the width and the amplitude of the $P_S$ peak. The broad $PpP_s$ multiple is matched by the choice of a gradational HVLC region at the base of the crust.

upper layer about 10 km thick with velocities of 4.5–6.5 km s$^{-1}$ and a high-velocity gradient which is interpreted as extrusive basalts.

There is evidence for an HVLC at the base of the crust. Furthermore, it appears to be gradational over a thickness of about 6–10 km. If a discrete layer of high-velocity lower crust with a strong impedance contrast (an ‘underplated’ region) were present, it would have to be less than 6 km thick to fit the observed $P_S$ peak in the receiver functions. The $PpP_s$ peak observed is poorly modelled by such a layer, and we conclude from our modelling that a gradational HVLC is more likely. We interpret this as caused by the intrusion of igneous sills into the lower crust, with the density of sills increasing downwards towards the Moho.

The crustal thickness we infer of 29–32 km is within the uncertainty estimates of thicknesses from wide-angle Moho reflections to the east and west of the Faroe Islands, although it is at the lower end of those estimates. Our Faroe Island estimates are also similar to crustal thicknesses of 27 km and 35 km calculated from receiver functions from the conjugate east Greenland coastal area at Ittoqqortoormiit in Scoresbysund and Sodalen, respectively (Dahl-Jensen et al. 2003; Kumar et al. 2007). The receiver function method relies on mode conversion to define boundaries, and therefore it is possible that the crustal thickness we estimate is actually the depth to a region within the HVLC where there are significant velocity contrasts rather than to the top of the mantle. This would be consistent with the apparent thinnest crust derived from receiver functions on the conjugate east Greenland coastal region being for stations lying directly above the track of the Iceland plume, where the thickest HVLC and concomitantly thicker crust overall would be expected. In the Ethiopian Rift, which is an analogous region of early continental breakup above a mantle plume, Stuart et al. (2006) report that receiver function results underestimate the crustal thickness due to phase conversions from a layer of HVLC. Our inversions shown in Fig. 7 suggest that there could be several kilometres of highly intruded rock beneath 32 km depth with velocities in excess of 7.6 km s$^{-1}$, although when there is such a large amount of igneous intrusion it becomes a moot point whether to interpret this as heavily intruded continental crust or as upper mantle. In any case, we expect there to be at least as much intruded igneous rock as extruded basalts on this volcanic continental margin, consistent with our results.

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