P-wave velocity distribution in basalt flows of the Enni Formation in the Faroe Islands from refraction seismic analysis

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Received October 2011, revision accepted January 2012

ABSTRACT

The main objective of this work is to establish the applicability of shallow surface-seismic traveltime tomography in basalt-covered areas. A densely sampled ~1300-m long surface seismic profile, acquired as part of the SeiFaBa project in 2003 (Japsen et al. 2006) at Glyvursnes in the Faroe Islands, served as the basis to evaluate the performance of the tomographic method in basalt-covered areas. The profile is centred at a ~700-m deep well. \( V_p \) and \( V_s \) and density logs, a zero-offset VSP, downhole-geophone recordings and geological mapping in the area provided good means of control.

The inversion was performed with facilities of the Wide Angle Reflection/Refraction Profiling program package (Ditmar et al. 1999). We tested many inversion sequences while varying the inversion parameters. Modelled traveltimes were verified by full-waveform modelling.

Typically an inversion sequence consists in several iterations that proceed until a satisfactory solution is reached. However, in the present case with high velocity contrasts in the subsurface we obtained the best result with two iterations: first obtaining a smooth starting model with small traveltime residuals by inverting with a high smoothing constraint and then inverting with the lowest possible smoothing constraint to allow the inversion to have the full benefit of the traveltime residuals.

The tomogram gives usable velocity information for the near-surface geology in the area but fails to reproduce the expected velocity distribution of the layered basalt flows. Based on the analysis of the tomogram and geological mapping in the area, a model was defined that correctly models first arrivals from both surface seismic data and downhole-geophone data.

Keywords: Full waveform, Ray tracing, Shallow seismic, Tomography, Velocity analysis.

INTRODUCTION

Since the start of the hydrocarbon exploration era in the Faroese area the sub-basalt imaging problem has been recognized as one of the major risk factors. The problem in obtaining sub-basalt information from reflection seismic data has usually been attributed to the quite different physical properties of basalt versus those of the overlying and underlying sediments, which also have large local variations. These large differences, especially in acoustic impedances, often cause poor transmission of energy, scattering, strong multiple reflections, multiple mode conversion and low-pass filtering of the energy that propagates through a stack of basalt flows (e.g. White et al. 2003; Ziolkowski et al. 2003).

More recent work has shown that although the seismic signal deteriorates as described above when propagating through...
P-wave velocity distribution from refraction seismic analysis

Figure 1 a) Location of the Faroe Islands relative to the extent of flood basalts. The international border of the Faroes territory is shown as a blue line (modified from Sørensen 2003). b) Geological map of the Faroe Islands showing the locations of deep boreholes (Vestmanna-1, Glyvursnes-1 and Lopra-1/1A) and the distribution of the three Palaeogene basalt formations (modified from Rasmussen and Noe-Nygaard 1970). c) Geological section through the Faroe Islands with the three wells (modified from Waagstein 1988). The location of the profile is shown on b).

A stack of basalt flows, there is still sufficient primary energy from below the basalts for sub-basalt imaging. It has been suggested that the problem reduces to focusing of the primary energy (e.g., Gallagher and Dromgoole 2008). Whatever processing sequence is used, the focusing of primary energy depends critically on the quality of the velocity models used in the processing.

The Faroe Islands Basalt Group, part of the North Atlantic Igneous Province, attains a thickness of more than 6.6 km. Previous studies have given us the means for a good understanding of the emplacement of the Faroe Islands Basalt Group and of its seismic properties. Knowledge of these properties can ultimately provide a basis for improved velocity models in the Faroese area (Rasmussen and Noe-Nygaard 1970; Hald and Waagstein 1984; Nielsen, Stefánsson and Tulinius 1984; Waagstein 1988; Waagstein and Andersen 2003; Boldreel 2006; Chalmers and Waagstein 2006; Passey and Bell 2007; Passey 2007; Passey and Jolley 2009).

The stratigraphic sequence of the Faroe Islands Basalt Group (Fig. 1) consists mainly in the following 3 formations with different characteristics: the ~3-km thick Beinisvøørð Formation (also referred to as the Lower Basalt Formation in the literature) is characterized by thick-sheet lobes of tabular-classic facies architecture, with thicknesses up to about 60 m but typically 5–25 m and with thin, mostly 0.5–2 m thick, volcaniclastic beds. The ~1.4 km thick Malinstindur
Table 1  Overview of velocity and attenuation information from three wells in the Faroes (Superscripts refer to: 1: Waagstein and Andersen 2003, 2: Christie et al. 2006, 3: Shaw 2006, 4: Shaw et al. 2008).

<table>
<thead>
<tr>
<th>Well name</th>
<th>Lopra-1/1A</th>
<th>Vestmanna-1</th>
<th>Glyvursnes-1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Section of stratigraphic sequence covered by well</td>
<td>Beinisvørð Fm below 900 m and uppermost 1000 m of Lopra Fm</td>
<td>Variation mainly between 5 and 6 km/s whilst the average velocity from VSP is 5.29 km/s.(^2)</td>
<td>Uppermost 350 m of Malinstindur Fm and lowermost 350 m of Enni Fm</td>
</tr>
<tr>
<td>(V_P) (velocity)</td>
<td>High-frequency variations mainly between about 4 and 6 km/s relating to flows and beds whilst the average velocity from VSP is 5.25 km/s.(^2)</td>
<td>(V_P) from log (V_S) ratio (1.84^2)</td>
<td>(V_P) from VSP (1.81^2)</td>
</tr>
<tr>
<td>(V_P)/(V_S) ratio from log</td>
<td>(1.84^2)</td>
<td>(1.8^1)</td>
<td>(1.8^1)</td>
</tr>
<tr>
<td>(V_P)/(V_S) ratio from VSP</td>
<td>(1.81^2)</td>
<td>(1.9^4)</td>
<td>(2.0^9)</td>
</tr>
<tr>
<td>Quality factor, (Q)</td>
<td>35 on average(^2)</td>
<td>25 on average(^4)</td>
<td>25 on average, 15 for the Enni Fm and 30 for the Malinstindur Fm(^4)</td>
</tr>
</tbody>
</table>

Formation (also referred to as the Middle Basalt Formation in the literature) consists mainly of thin flow lobes, about 1–2 m thick, forming lava flows of compound-braided facies up to about 20 m thick. Volcaniclastic beds are very thin or absent in the Malinstindur Formation. The Enni Formation, \(~0.9\) km thick, also referred to as the Upper Basalt Formation, consists in a mixture of simple and compound flows with tabular-classic and compound-braided facies architectures, respectively, with volcaniclastic beds ranging in thickness from 0.01–5 m.

Quantitative information on the Faroe Islands Basalt Group comes mainly from 3 deep onshore wells located such that they cover different sections of the stratigraphic sequence (Fig. 1). A brief overview of the velocities is listed in Table 1. Within the flow lobes the velocity is highest in the core, where porosity is lowest, whilst it is lower in the porous crust and basal zone. Volcaniclastic beds generally have the lowest velocities. The \(V_P\)/\(V_S\) ratio is about the same for porous crust, massive core and volcaniclastic beds (Waagstein and Andersen 2003; Boldreel 2006). Both the Beinisvørð and the Enni Formations have high fluctuations in velocities relating to the high abundance of volcaniclastic beds while the Malinstindur Formation, with fewer volcaniclastic beds, has a more stable velocity distribution. As a consequence of this, acoustic impedance in the Beinisvørð and Enni Formations has larger amplitudes than in the Malinstindur. This is reflected in the seismic response of the three formations, where the Malinstindur is apparently transparent to seismic signals, in contrast to the Beinisvørð and Enni Formations (Petersen, Andersen and White 2006).

A second source of basalt velocities is from large-scale refraction seismic experiments in the Faroese area. In agreement with well data (Table 1), these experiments consistently report velocities of \(~5\) km/s (e.g., Richardson et al. 1999; Fliedner and White 2003; Bohnhoff and Makris 2004; Raum et al. 2005; Klingelhofer et al. 2005). However, these experiments give very little detail on the velocity distribution within the basalt sequence and do not identify any of the three main formations (Beinisvørð, Malinstindur or Enni).

Detailed velocity information on the Faroe Islands Basalt Group can be obtained by velocity inversion using densely sampled reflection seismic data, especially in areas where dipping layers intersect the sea-bed. In this paper we report on velocity inversion of a data set acquired across the shoreline at Glyvursnes, Faroe Islands. The 700-m deep Glyvursnes-1 well, drilled as part of the project known as SeiFaBa (Japsen et al. 2006), was located so that the well intersects the Malinstindur-Enni boundary and so as to give optimal conditions for the combination and comparison of different types of data, namely, offset VSP and surface seismic, both onshore and offshore. These data make it possible to assess the performance of velocity inversion on a densely sampled reflection seismic
profile onshore and in shallow waters offshore in an area with high velocity contrasts of the subsurface.

The inversion is based on the WARRP (Wide Angle Reflection/Reflection Profiling) program package from GeoPro GmbH in Hamburg (Ditmar et al. 1999). It allows for any source-receiver configuration and can be applied to any toponography. The method calculates traveltimes from ray tracing. The general tomography algorithm that WARRP is based on is listed in Appendix A. When using ray-theory based inversion schemes, their limitations have to be considered: the ray theory does not account for diffractions; is sensitive to low angular coverage; has less resolution relative to waveform inversion (Williamson and Worthington 1993) and, for highly heterogeneous velocity distributions, areas of the model are bypassed by raypaths (e.g., Washbourne et al. 2008). However, with a reasonable parameter selection, ray-theory inversion is an efficient tool for providing detailed velocity information.

THE DATA

This refraction seismic analysis is based on a reflection seismic profile from the SeiFaBa 2003 acquisition (Andersen et al. 2004; Petersen et al. 2006). The initial profiles, GBX602 and GBXDYN, form a combined profile with certain qualities that make it well suited for shallow refraction seismic analysis. It has relatively long offsets, two-way shooting and is centred at the Glyvursnes-1 well. The total length of the receiver layout of the profile is 1086 m, comprised of a 400-m geophone layout and a 600-m streamer layout. The geophone interval is 5 m, every fourth geophone having 3 components. The hydrophone interval is 6.25 m. The sources are a 160-in³ airgun cluster with source intervals from 15–20 m and 250-g dynamite charges in 3-m deep holes at source intervals from 50–100 m. The maximum source-receiver offset is 1343 m. Both dynamite shots and airgun shots were recorded on both geophones and streamer simultaneously. The recorded length is 3 s, sampled at 0.5 ms. All shots were also recorded on a downhole geophone at 400 m depth in the Glyvursnes-1 well.

During the modelling a third profile, SeiFaBa-01 (dynamite-geophone), was included in order to balance the ray-coverage. The geophone layout of this profile coincides with the GBXDYN shot points and is 600 m long with 5-m receiver intervals starting at the well location. There are 50 shot points along this profile at 10-m intervals starting close to the well location and ending ~500 m from the well. Shots 17–25 are missing due to logistics and shots 14, 36, 37 and 38 due to misfiring.

As part of the SeiFaBa project (Japsen et al. 2006), the Glyvursnes-1 well was drilled to a depth of ~700 m. This well gives velocity information in the form of the zero-offset VSP and velocity logs. The logs from Glyvursnes-1 used for this analysis (Table 2) consist in the GEUS-processed full-waveform velocity logs (Waagstein and Andersen 2003) and the bulk-density log recalibrated to measurements on core samples (Waagstein, unpublished data). The VSP source is a 150-in³ airgun fired in a specially constructed pond with water depth of about 1.5 m. The offset from the pond to the well is 14 m. The VSP receiver is a 3-component downhole geophone, clamped with a hydraulic system, recording at 10-m depth intervals from 50–600 m (Shaw 2006).

The SEG polarity standard for hydrophone or vertical-component geophone data specifies that, in normal polarity, the first break due to a compressional arrival is represented by a negative number (Thigpen, Dalby and Landrum 1975; Brook, Landrum and Sallas 1993), i.e., a downward or leftward deflection on a seismogram. This corresponds to the upward motion of a geophone case or to a pressure increase at a hydrophone. For an explosive source and either a direct arrival or a reflection from an interface with a positive reflection coefficient, this implies a negative deflection for a minimum-phase wavelet (Sheriff 2002, p. 267). But for a zero-phase wavelet this same situation is represented by a positive central peak. However, Sheriff (2002) also noted that in some areas, including the North Sea, the opposite convention is used for zero-phase wavelets. Inspection of our data shows that while the geophone recordings are in accordance with the SEG normal polarity standard the polarity of the hydrophone data had to be reversed to comply with SEG normal polarity.

The start of data recording is triggered by the source. So the data should have proper zero time. However, zero time was re-established from analyses of gathers, this being a case of a small-scale profile very sensitive to small systematic errors. To facilitate the picking of events, the zero time for a positive reflection was established so as to correspond to the trough-peak zero crossing.

| Table 2 | Well logs used in the analysis are the full-waveform velocity logs processed by Geological Survey of Denmark and Greenland (GEUS) and the bulk-density logs recalibrated to measurements on core (Waagstein, unpublished data). |
| Start Depth | Stop Depth | Sampling Interval | File |
| (m) | (m) | (m) |
| ρ | 2.93 | 698.83 | 0.1 | GL1-dens.txt |
| Vₚ | 5.6 | 597.4 | 0.2 | GL1-FWS_GEUS.LAS |
| Vₛ | 5.6 | 597.4 | 0.2 | GL1-FWS_GEUS.LAS |

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Geophone positions in the GBX survey (GBXDYN/GBX602) have an uncertainty on the order of ±0.05 m, whilst in the SeiFaBa-01 survey it is greater – on the order of ±1 m. The uncertainty in the dynamite shot positions of both these surveys is also about ±1 m. The airgun shot points are measured with an uncertainty of about ±2 m.

The hydrophone positions are determined both from the traveltimes of the direct arrivals and from the two tethering points of the streamer, i.e., the jack-up rig and the tugboat. The uncertainty in the hydrophone positions is 0 to +5 m in the in-line or profile direction and ±3 m cross-line.

We specify a single 2D survey line in the x-direction (thin red line in Fig. 2), a combination of the three profiles: GBX602, GBXDYN and SeiFaBa-01. Source positions are projected in the y-direction onto this survey line and projected (best-fit) receiver positions are determined from the actual offsets and projected source positions.

In the present data some actual source positions have a significant cross-line offset (y-direction) from the survey line (Fig. 2). This leads to a significant scatter for some of the projected receiver positions; so, whilst the source-receiver offset is preserved, there can be significant errors in some of the projected receiver positions. In the present modelling we accept receiver position errors of up to ±15 m from the correct values. Traces with larger errors are disregarded when picking first breaks. Such an error in receiver position, ±15 m, may be relatively large but it is a trade-off, since imposing a smaller deviation would lead to the abandonment of too much data.

The initial model is designed from the topographic and bathymetric information of the modelled survey line (Fig. 3). The topography of the land profile is accurately defined due to the high accuracy of the land navigation and because the shot points coincide with the geophone layout. The sail line, i.e., the source line, is offset from the streamer layout, i.e., the receiver line, by a distance of ∼50 m. For parts of the profile there are differences of up to 10 m between the depths at the hydrophone positions and the depths at the corresponding shot positions. In the absence of a better solution, we use the average of these depths.

RESULTS

The results are discussed in view of the performance of inversion parameters and the obtained velocity model. The main consideration during inversion was to arrive at a single model that produces consistent traveltimes for all gathers and is consistent with known geology in the area. The final results of the tomography reflect the outcomes from about 50 different inversion sequences using varying input parameters to optimize the resulting model.

The parameters to adjust during inversion are:
- The velocity-regularization parameter, $\alpha$ in equation (A6)
The objective-function minimization parameter, the maximum desired value of the objective function $\Phi(\Delta m)$ (equation (A6))

- The gridding of the model

The initial model to be used for tomographic inversion is defined as two-layer with the upper layer as the sea and the lower layer below the sea-bed (Fig. 4a). The upper layer (the sea) is considered to be homogeneous with a velocity of 1480
Table 3 The first inversion sequence. RMS: deviation between picked and modelled travel times, VRP: velocity-regularisation parameter and OFMP: objective-function minimisation parameter. e.g., velocity-regularisation parameter = 100 and objective-function minimisation parameter = 0.1 are used to arrive at the first iteration.

<table>
<thead>
<tr>
<th>Model</th>
<th>RMS</th>
<th>VRP</th>
<th>OFMP</th>
</tr>
</thead>
<tbody>
<tr>
<td>Initial model</td>
<td>0.0178</td>
<td>100</td>
<td>0.1</td>
</tr>
<tr>
<td>Iteration 1</td>
<td>0.0080</td>
<td>30</td>
<td>0.03</td>
</tr>
<tr>
<td>Iteration 2</td>
<td>0.0045</td>
<td>10</td>
<td>0.01</td>
</tr>
<tr>
<td>Iteration 3</td>
<td>0.0026</td>
<td>3</td>
<td>0.003</td>
</tr>
<tr>
<td>Iteration 4</td>
<td>0.0022</td>
<td>1</td>
<td>0.001</td>
</tr>
<tr>
<td>Iteration 5</td>
<td>0.0020</td>
<td>0.3</td>
<td>0.0003</td>
</tr>
<tr>
<td>Iteration 6</td>
<td>0.0019</td>
<td>0.1</td>
<td>0.0001</td>
</tr>
<tr>
<td>Iteration 7</td>
<td>0.0025</td>
<td>0.03</td>
<td>0.00003</td>
</tr>
<tr>
<td>Iteration 8</td>
<td>0.0073</td>
<td>0.01</td>
<td>0.00001</td>
</tr>
<tr>
<td>Iteration 9</td>
<td>0.0075</td>
<td>0.003</td>
<td>0.000003</td>
</tr>
<tr>
<td>Iteration 10</td>
<td>0.0075</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

m/s while the layer below the sea-bed is the one considered for inversion. To ensure the presence of turning rays from the substratum, the velocity distribution of the lower layer has a positive vertical velocity gradient. Apart from this the velocity distribution below the surface was chosen with no assumptions of prior knowledge.

The velocity is defined at nodes of a 2D grid in the vertical plane of the survey line. Each velocity layer has a limit of 1000 grid points. A wide range of gridding was tested including non-regular grids. The best performance was with a 50 × 20 (horizontal x vertical) grid, equivalent to approximately 20 m vertical and 30 m horizontal spacing.

The inversion was first done following Ditmar et al. (1999), by starting with a high velocity-regularization parameter and decreasing by a factor of 3–5 for each iteration, keeping the objective-function minimization parameter at no more than 1% of the velocity-regularization parameter. We present an inversion sequence that was followed for 10 iterations (Table 3). At each step the quality of the inverted model was considered. The primary quality parameters supplied by the WARRP code are the ray density distribution (Fig. 4f–j) and the root-mean-square (RMS) deviation between picked and modelled traveltimes (Table 3). The inverted model was also assessed by inspecting raypaths and traveltimes on individual gathers (e.g., Figs 5 and 6).

Further, quality was estimated by relating the inverted model to the known geology in the area. In fact the comparison between the model and the geology was weighted more than the quality estimation based on traveltimes. Based on RMS deviation the desired solution would be at iteration

Figure 5 Traveltimes and raypaths for the final velocity model. Shot gather 4.
4, 5 and 6. But these solutions do not reflect the layering of basalt flows in the area (Passey 2005) and were thus rejected.

After the first iteration, the model starts to develop a low-velocity zone in the middle that was more pronounced for each iteration (closed green contours in Fig. 4c–e). This low-velocity zone was interpreted as the computational expression of a low-velocity layer, as illustrated by the area marked A in Fig. 4(c). The handling of a low-velocity layer puts greater demands on the modelling parameters. Failure to model the low-velocity layer affects the velocities below the low-velocity layer; in other words, the errors propagate to other areas. To produce an overall traveltime corresponding to the picked events, velocities in area A (Fig. 4c) that are too high result in velocities that are too low in other areas of the model. Different velocity-regularization parameters and objective-function minimization parameters for the gradual step-by-step inversion were tested but all failed to model the expected layered structure of successive basalt flows.

The best results were obtained with a quite different approach that allowed us to obtain the final model in only two steps. The first iteration used the same high values of velocity-regularization parameter and objective-function minimization parameter as in the first approach to obtain an initial model, a coarse model that with least details still produces traveltimes close to those of the picked events (iteration 1 in Table 4). The next iteration was performed with the lowest possible values of the velocity-regularization parameter and objective-function minimization parameter that the algorithm could handle, thus minimizing the effect of the regularization condition (final inversion, Table 4).

Whilst this result (Fig. 4e) in terms of the RMS deviation performs worse than most of the stepwise iterations (Table 3), it shows a better modelling of the interpreted low-velocity layer and when regarding the geology in the area the velocity distribution is consistent with the strike and dip of flows (Passey 2005). The contrast between the RMS deviation and quality assessment of the model must also be seen in light of the amount of modelled arrival times, so although iteration 4 has lower RMS deviation than the final inversion it is apparently based on far fewer traveltimes as indicated by the ray density distributions of the models (Fig. 4h and Fig. 4j). Figures 5 and 6 show two seismic shot gathers representing forward and reverse shooting with raypaths and traveltimes based on the final inversion.

<table>
<thead>
<tr>
<th>Model</th>
<th>RMS</th>
<th>VRP</th>
<th>OFMP</th>
</tr>
</thead>
<tbody>
<tr>
<td>Iteration 1</td>
<td>0.0080</td>
<td>0.005</td>
<td>0.00001</td>
</tr>
<tr>
<td>Final inversion</td>
<td>0.0041</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 7 Effects of uncertainty on the final model. a) Random error between ± 5 ms on picks; b) + 5 ms time shift on picks; c) shifting the streamer 5 m to the north. Same colour scale for all velocity models. d), e) and f) Respective velocity ratio relative to the final model (Fig. 4e).

**Error calculation**

Uncertainties can be characterized as dependent or independent. Independent uncertainties are related to picking of events and non-systematic errors in receiver and shot point positions. Effects of independent errors were tested by imposing a random error of ± 5 ms on all picks and then inverting with the same parameters as for the final inversion. Comparison shows a velocity difference of less than ±5% (Fig. 7a and Fig. 7d) relative to the final inversion.

Dependent uncertainties are considered in relation to systematic errors in zero times of the gathers and systematic errors in positions of receivers and shot points. The response to the systematic error in zero time of the gathers was tested by time shifting all traces the same amount, 5 ms downwards. The very shallow velocities are affected as velocities below the sea-bed are as much as 10% lower, velocities onshore as much as 15% lower. The deeper velocities are within ±15% of velocities relative to the final inversion (Fig. 7b and Fig. 7c).

Systematic errors in receiver and shot positions are related to the streamer layout relative to the geophone layout. Whilst there is good constraint on the position of the geophone layout the position of the streamer layout is liable to larger uncertainties in the north-south direction. According to the uncertainty analysis on the positioning of the streamer the streamer was shifted 5 m to the north. The modelling response shows velocities within 5% of those from the original models (Fig. 7c and Fig. 7f).

**Tie to surface mapping**

The correlation to the known geology in the area serves two purposes: the velocity distribution establishes velocity properties of geological structures in the area and it indicates the extent to which it is possible to interpret geological structures directly from the modelled profile.

As part of the SeiFaBa project (Japsen et al. 2006), Passey (2005) performed a geological study of the area surrounding the Glyvursnes-1 well to establish the stratigraphy of the
younger rock units that were not encountered in the well and to evaluate any structural features in the area. The lowest part of the mapped stratigraphic sequence intersects with the modelled profile. Based on the expected seismic properties of the mapped units, a few markers, A, B and C (Fig. 8), are defined for use in the correlation with the modelled profile. The section between A and B consists mainly in a 25–35 m thick tabular basalt flow and is thus expected to be a high-velocity layer. The section between B and C consists mainly of a ∼4 m thick sandstone, an ∼8 m thick tabular basalt flow and a 9–16 m thick volcaniclastic sequence and is thus expected to be a relatively low-velocity layer.

The markers were extrapolated onto the modelled profile according to the geological mapping (Fig. 8). The section between A and B does in fact correlate with a well-defined high-velocity section of the modelled profile and the section between markers B and C correlates with a low-velocity section of the modelled profile. The velocity distribution complies well with the dip of the extrapolated markers. The place where marker C crosses the sea-bed (D in Fig. 8) coincides with the base of a low-velocity zone, while farther to the left the low-velocity zone extends below marker C.

**QUANTITATIVE CORRELATION TO VELOCITY WELL LOGS**

The VSP and velocity logs

Interval velocities were derived from VSP data by picking traveltimes. First breaks were picked on the first maximum gradient of the signal, i.e., the first zero-crossing of the second derivative. A comparison between the VSP-picked traveltimes and the traveltimes calculated from the velocity log showed that although the two are similar, traveltimes from the velocity log were slightly lower (Fig. 9). The difference of ∼2.8% agrees with Shaw (2006).

The delay in VSP traveltimes is also in agreement with a study by Stewart, Huddleston and Kan (1984) who reported that the VSP traveltimes are delayed by 2.0 ms/1000 ft on average. They attributed this to the different travelling pattern of the high-frequency sonic log signal and the low-frequency VSP signal and showed that short-path multiples and velocity dispersion can account for the seismic pulse delay. We take a different approach to explain the time delay of the seismic signal by the rescaling of the sonic log to scales of the seismic signal.

The frequency of the signal used for logging was about 23 kHz (Waagstein and Andersen 2003) while the VSP signal had a centre frequency of ∼33 Hz (Shaw 2006). In other words, logging the data compares to looking at the earth in scales of centimetres whilst the VSP compares to seeing the earth in scales of decametres.

Backus averaging analyses (Backus 1962) of the log data by least RMS traveltme deviation between logged and VSP data for various averaging intervals show best fit for an averaging...
interval of $\sim 25$ m. This can be interpreted as indicating that
the seismic signal used for the VSP in some aspects sees the
earth at a resolution of $\sim 25$ m.

At the 25-m Backus-averaging interval, the traveltimes from
the velocity log and the VSP correspond very well (Fig. 9) and
there is still good consistency between the details of the two.
There is, however, a section in the 100–200 m depth range
where the traveltimes deviate significantly and consistently.
This can be a matter of the horizontal extent covered by the
signal. The full sonic log is affected by properties only in the

Figure 11 Shot gathers 4: traveltimes from full-waveform synthetics versus ray theory. See Fig. 5 for the seismic gather.

Figure 12 Shot gather 46: traveltimes from full-waveform synthetics versus ray theory. See Fig. 6 for the seismic gather.
vicinity of the well location (scales of decimetres), whilst the VSP signal is affected by properties over a much larger distance (scales of decametres).

VSP traveltimes were converted to interval velocities at various averaging intervals. For 30-m intervals, realistic velocity values were obtained and these show good agreement with the logged velocities (Fig. 10).

Comparing the modelled profile to VSP and velocity log

A quantitative comparison between modelled velocities and velocities from VSP and logs is obtained by extracting velocities from the model along a depth profile corresponding to the vertical position of the Glyvurnes-1 well (Fig. 10; see Fig. 8 for location).

Down to 0.1 km depth the modelled velocities resemble the interval velocities from the log, although the modelled velocities are significantly lower, by up to 1 km/s. At 0.1–0.13 km depth the modelled velocities coincide with the log velocities and below this the modelled velocities are higher than the log velocities. Below 0.13 km depth the modelled velocities lose the details of the velocity distribution.

A comparison of the vertical and horizontal P-wave velocities at 25-m Backus averaging shows that throughout the total depth, the horizontal P-wave velocity is higher. While in most sections the difference is small, there are sections with a
A comparison between VSP velocities and modelled refraction velocities is equivalent to a comparison between vertical and horizontal P-wave velocities since the VSP velocities are derived from vertically propagating waves and the modelled velocities are derived from near-horizontal raypaths. However, differences in details between the 30-m interval VSP velocities and the modelled velocities make comparison difficult (Fig. 10). Therefore, the modelled velocities were compared with 100-m interval VSP velocities. In areas with good ray coverage the modelled velocities were lower than the VSP velocities, contrary to results from the Backus averaging (Fig. 10).

**Full-waveform modelling**

When using the ray theory for tomographic inversion on small-scale seismic profiles, as in the present case, it should be considered to what extent the solution is also valid for full-waveform signals. To validate the solution, we compared full-waveform synthetics with traveltimes from the ray theory.

The *ELA2D* finite-difference full-waveform code (RELEASE 1.1 1997) by Joachim Falk was used for the full-waveform modelling. This code is based on the finite-difference method of Virieux (1986). The input for the modelling consists in grids of P- and S-wave velocities and densities. P-wave velocities came from the final velocity model (Fig. 4e); S-wave velocity for the water layer was set to 0 and for the subsurface it was set from the general $V_p/V_s$ ratio of 1.8 for basalts (e.g., Waagstein and Andersen 2003). Densities were assigned as 1000 kg/m$^3$ for the water layer and 2500 kg/m$^3$ for the subsurface as an estimated average from the density log.

The full-waveform synthetic confirmed that for surface seismic data the final velocity model (Fig. 4e) produces valid traveltimes of first breaks for a 30-Hz Ricker wavelet (Figs 11 and 12) except for a small section of the farthest offset of gather 46, where the traveltimes from the ray theory have a delay relative to the full-waveform data.

However, when verifying the model against the data recorded on the downhole geophone at a depth of 400 m in the Glyvursnes-1 well, the solution is invalid in all aspects. Firstly there is an inconsistency between traveltimes by the ray theory and from full-waveform synthetics (Fig. 13 upper) and secondly the model produces incorrect traveltimes relative to seismic data (Fig. 13 middle).

The model from the tomographic inversion (Fig. 4e) shows indications of the layered structure of basalt flows. The horizontal continuation of the layers is, however, incomplete. The inverted model is interpreted to contain three low-velocity layers: I, II and III (Fig. 14). The strike and dip of layers from surface mapping in the area (Passey 2005) were used to establish the suggested location of the low-velocity layers.

A layered model was defined based on this interpretation (Fig. 15). Except for the uppermost layer, the layers were defined as homogeneous with flat interfaces. The velocities for the layers were derived from areas in the final velocity model with high ray density (Figs 4e and Fig. 4j). The velocity distribution in the uppermost layer in the subsurface came directly from the tomographic inversion.

The modelling of interfaces as flat and layers as homogeneous does not reflect the expected structure of flows and flow interfaces (e.g., Waagstein and Andersen 2003; Passey 2007; Shaw et al. 2008; Bean and Martini 2010). These properties are indeed considered important for understanding the seismic properties of successive basalt flows but are beyond the scope of this paper, where we focus on seismic traveltimes.

Some of the small-scale detail of the velocity distribution in Fig. 4(e) could be interpreted as real properties of the basalt...
flows; however, when considering that these, to some degree, are artefacts of the method, we decided to use the large-scale velocity distribution of the model as the basis for the layered model. Each layer in the model could very well represent the combined property of many basalt flows.

The seismic response from the layered model (Fig. 15) agrees with the seismic data and the downhole-geophone data – considering both traveltimes from the ray theory and full-waveform synthetics (Figs 16–18). It should however be noted that the full-waveform synthetics show high amplitudes for interface waves and post-critical-angle reflections that are not present, or that have much smaller amplitude, in the seismic data. This is attributed to the simplified modelling of interfaces as flat and the layers as homogeneous (Fig. 16).

Traveltimes from reflections at the two deepest interfaces tie to events on the gathers (interfaces 5 and 6 in Fig. 17). Reflections from interface 5, with a zero-offset time of 0.11 s, tie to a zero crossing from above that correctly represents a positive reflection; and reflections from interface 6, with zero-offset time at 0.16 s, tie to a zero crossing from below that correctly represents a negative reflection. At the Glyvursnes-1 well location, the depth of interface 6 is ~300 m, which thus represents the depth at which the character of the sonic log changes from large alterations in velocities to a more stable...
velocity trend. This represents a marker close to the base of the Upper Basalt Formation at 355 m depth (Waagstein and Andersen 2003).

**DISCUSSION AND CONCLUSION**

The principal objective of this work was to gain an understanding of the applicability of shallow surface-seismic traveltime tomography in basalt-covered areas. The tomogram (Fig. 4e) gives velocity information about and correlates to geology in the area, although clearly the velocity distribution does not reflect the expected basalt-flow structure. The *a priori* geological mapping of the area (Passey 2005) played a significant role in the interpretation of the solution. Whereas Ditmar et al. (1999) recommended a gradual decrease of the smoothing constraint over several iterations, we conclude that, in the present case, first obtaining a smoothly varying starting model that produces traveltimes close to picked events and then doing the inversion in a single iteration with minimized smoothing constraint gives the best result.

The better result from two-step inversion relative to the gradual approach is attributed to the current case of multiple velocity inversions (high-velocity layer overlying low-velocity layer) having a detrimental effect on the ray coverage in the form of shallow penetration and uneven ray density distribution. The smooth starting model has a relatively even ray...
density distribution and the traveltime residuals contain information that is utilized in the two-step inversion sequence. The gradual inversion sequence, however, develops a more uneven ray density distribution for each iteration and loses thus some of the initial information.

All considered, stand-alone traveltime tomography based on the ray theory must be seen as an inadequate method for imaging the current profile, although the tomogram gives usable velocity information for the near-surface geology. However, based on the interpretation of the tomogram and geological mapping, a model was defined (Fig. 15) that produces correct traveltimes for first arrivals and is consistent in comparison with full-waveform synthetics for all gathers, including the downhole-geophone gather. This model is well documented and is a valid, although simplified, velocity distribution of the profile.

Comparing velocity log and VSP with the tomogram shows that in the uppermost 100 m the velocities from tomographic inversion are significantly lower (Fig. 10). This could be indicative of anisotropy, as log and VSP represent vertical velocities whilst the tomogram resembles horizontal velocities. This is also in line with Kiørboe and Petersen (1995) who reported a 10% lower horizontal P-wave velocity in the uppermost 800 m around the Lopra well. However, the apparent anisotropy...
could also be an artefact of the method, as an overall positive vertical velocity gradient is needed for turning rays from the substratum for surface seismic acquisition geometry.

ACKNOWLEDGEMENTS

This study is part of a project (contract number: C46–49-01) sponsored by the Sindri Group. The data for the analysis are from the SeiFaBa project – also sponsored by the Sindri group. Thanks to the Department of Earth Sciences, Faculty of Science, at Aarhus University for their efforts in maintaining equipment that makes small-scale seismic acquisition feasible and thanks to Per Trinhammer for, with high dedication and technical skill, making the very best out of the acquisition. Also thanks to Michael Worthington for help with the VSP and downhole-recorded data. Thanks to the farmer Sjúrður Patursson and the fish-farming company Gulin for help and goodwill during all acquisition at Glyvursnes. Thanks to Landsverk (the Faroese Office of Public Works) for assisting with high-accuracy position measurement. And finally, thanks to all the driving forces behind the SeiFaBa group.

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APPENDIX A

THE TOMOGRAPHY ALGORITHM

The tomographic inversion of refraction seismic data is a matter of inverting traveltimes to a velocity model. This means: what velocity distribution does a model need to have in order to produce traveltimes that reproduce the observed traveltimes?

A condition for finding a solution is that the traveltimes can be related to a parametrized model. One way to parametrize the model is to divide it into cells – or to grid the model. The total traveltime for a certain source-receiver pair is then a summation over the time spent in each cell

\[ t_i = \sum_{j=1}^{M} \Delta s_j p_j, \]  

(A1)

where \( t_i \) is the traveltime for a certain source-receiver pair and \( M \) is the number of cells in the gridded model. \( \Delta s_j \) is the travelled length in cell \( j \) of the model and \( \Delta p_i \) is the slowness in cell \( j \) of the model.

The inversion of traveltime data is a matter of finding \( p_j \). But the distances travelled in each cell, \( \Delta s_j \), is unknown. By imposing modelled traveltimes from an initial model a traveltime residual can be expressed by

\[ \Delta t_i = \sum_{j=1}^{M} \Delta s_j \Delta p_j, \]  

(A2)

where \( \Delta t_i \) is the difference between the observed and modelled traveltimes and \( \Delta p_j \) is the difference between the slowness of the initial model and that of the desired model. In matrix form this is

\[ A \Delta m = \Delta t. \]  

(A3)

\( \Delta t \) is a vector containing the traveltime residuals for all source-receiver pairs, \( \Delta m \) is the model residual consisting in all \( \Delta p_j \), and \( A \) is a \( (i_{max} \times j_{max}) \) matrix of \( \Delta s \). The matrix \( A \) is determined from the initial model. Subtracting \( \Delta m \) from the initial model \( m_0 \) will then give a model more consistent with the observed traveltimes. The desired vector, \( \Delta m \), is not found from matrix inversion due to the high computational cost. Instead an objective function is designed:

\[ \Phi(\Delta m) = \| A \Delta m - \Delta t \|^1, \]  

(A4)

which is to be minimized. This is the general formulation of the problem. The inputs for equation (A4) are the wave paths calculated from an initial model and the traveltimes from the first breaks.

It is common practice to calculate the wave propagation based on the ray theory. This offers a computationally cost-efficient method for doing the job. However, the ray theory is a high-frequency approximation and does not account for diffraction effects. The ray theory is in fact only accurate for modelling anomalies down to the scales of the first Fresnel zone, i.e., a few wavelengths (Williamson 1991; Williamson and Worthington 1993). Alternative methods of obtaining wave paths and traveltimes have been suggested by various authors (e.g. Červený and Soares 1992, Vasco, Peterson and Majer 1995).

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The WARRP (version 3.0.17) program package used for the current project is based on the inversion scheme outlined in equation (A4). The following is an outline of the inversion algorithm. For details on the inversion algorithm we refer to Ditmar et al. (1999).

The objective function in equation (A4) is sensitive to noisy data, as the model parameters will be unrealistically high or low in order to satisfy erroneous travel times. Therefore some kind of smoothing is used to lessen these effects (Phillips and Fehler 1991). In the WARRP algorithm the smoothing is applied with a regularization condition (equation (A5)):

\[ R = \int_{\Omega_1} \left[ 10 \left( \frac{\partial n(x, z)}{\partial x} \right)^2 + \left( \frac{\partial n(x, z)}{\partial z} \right)^2 \right] dx dz. \]  

(A5)

\( \Omega_1 \) is the area covered by the velocity grid and \( n \) is the model vector \( \Delta m \). It should be noted that for purposes of inversion the model vector, here defined as change in slowness, is defined as the fractional change of slowness in the description by Ditmar (1999).

The regularization condition involves an expression for the magnitude of the change of the model vertically and horizontally. The factor 10 in the first term accounts for the assumption that vertical variations are more probable than horizontal ones. The intensity of the regularization is controlled by a velocity-regularization parameter, \( \alpha \), which appears in:

\[ \Phi(\Delta m) = \| A\Delta m - \Delta t \|^3 + \alpha R. \]  

(A6)

The raypaths and travel times necessary for the inversion are calculated from the ray theory with the SEIS83 code (Červený and Pšeničk 1984). The current project represents a shallow, small-scale profile, where the section under investigation is expected to have large heterogeneities over small distances. So the limitations of the ray theory for refraction tomography are definitely an issue here. The shortcomings of the ray theory are expected to have their greatest effect in areas with low ray coverage, while areas with large angular coverage are expected to be less influenced.

Finally, it should be emphasized that the results of the inversion are non-unique. This implies that the solution must be viewed in relation to expected geology and other \textit{a priori} information in order to disqualify unrealistic models.