Seismic attenuation in Faroe Islands basalts
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ABSTRACT
We analysed vertical seismic profiling (VSP) data from two boreholes at Glyvursnes and Vestmanna on the island of Streymoy, Faroe Islands, to determine the magnitude and causes of seismic attenuation in sequences of basalt flows. The work is part of SeiFaBa, a major project integrating data from vertical and offset VSP, surface seismic surveys, core samples and wireline log data from the two boreholes. Values of effective seismic quality factor (Q) obtained at Glyvursnes and Vestmanna are sufficiently low to significantly degrade the quality of a surface reflection seismic image. This observation is consistent with results from other VSP experiments in the North Atlantic region. We demonstrate that the most likely cause of the low values of effective Q at Glyvursnes and Vestmanna is a combination of 1D scattering and intrinsic attenuation due to seismic wave-induced fluid flow within pores and micro-cracks. Tests involving 3D elastic wave numerical modelling with a hypothetical basalt model based on field observations, indicate that little scattering attenuation is caused by lateral variations in basalt structure.

INTRODUCTION
The poor quality of most seismic reflection images within and beneath basalts is most likely to be due to a combination of factors whose relative importance varies for different survey locations. In the Columbia Plateau, USA, 6 to 120-metre thick basalt flows are interbedded with clay layers which are up to 30 m thick. Pujol, Fuller and Smithson (1989) concluded that there is nothing unusual about the energy transmission characteristics of these basalts but attributed the observed poor seismic reflection data quality to reverberations caused by the large contrast in the elastic properties of the interbedded clay and basalt. Martini and Bean (2002) investigated by numerical modelling the seismic wave scattering due to internal velocity heterogeneity within a basalt sequence and roughness of the boundary surfaces at a scale similar to the seismic wavelength. They concluded that of two possible causes investigated with their models, interface scattering had the most detrimental effect on imaging at depth beneath a basalt succession. Fliedner and White (2001) have drawn attention to the detrimental effect of multiples produced between the sea-surface and the top of a basalt layer on two profiles from the Faroe-Shetland Basin.

Pujol and Smithson (1991) obtained an estimate of 50 for the seismic quality factor (Q) for the Columbia Plateau basalts. This is of the same order as many estimates of Q from localities in sedimentary basins where seismic reflection techniques are usually highly effective. However, other workers have obtained significantly lower values. Rutledge and Winkler (1989) obtained a Q value of approximately 25 for a basalt series beneath the eastern Norwegian Sea. Q values from 15 to 35 were obtained by Maresh and White (2005) and Maresh et al. (2006) for a basalt series in the Rockall Trough. The work of Christie, Gollifer and Cowper (2006) is of particular relevance to our study. They analysed vertical seismic
In this paper, we describe an investigation of the magnitude and possible mechanisms of seismic attenuation in basalts in the Faroe Islands. The work is part of the SeiFaBa project (from 2002–2006) involving the analysis of core samples and wireline log data from two boreholes on the island of Streymoy, combined with VSP, offset VSP surveys and extensive surface reflection and refraction seismic surveys. Preliminary results of this project are described in Japsen et al. (2005). The lavas in the Faroe Islands region are divided into the Lower, Middle and Upper Basalt Formations (Rasmussen and Noe-Nygaard 1970; Waagstein 1988). Individual basalt beds in the Lower and Upper Formations have average thicknesses of about 20 and 10 m, respectively. Sediment layers up to 10 m thick between basalt beds are found within the Lower and Upper Formations, but most interbasaltic sediment layers in these two formations are less than 1 m thick. In the Middle Formation, individual basalt beds are thin and sediments are virtually absent (Rasmussen and Noe-Nygaard 1970). The Glyvursnes-1 hole intersects both the Upper and Middle Formations and the Vestmanna-1 hole intersects the Middle and Lower Formations. In addition, the Lopra-1/1A hole on the island of Suduroy, drilled in 1981 and deepened in 1996, intersects the Lower Formation. The analysis of log and VSP data from this hole is described in Christie et al. (2006). These three holes provide an outstanding source of data for the determination of both the petrophysical and seismic properties of basalts in the Faroe Islands region and how these properties might vary laterally. The studies have been pursued in anticipation of continued interest in petroleum exploration in the Faroe area.

The structure of this paper is as follows: we describe the acquisition of the vertical VSP data acquired from the Glyvursnes-1 and Vestmanna-1 holes; we determine effective Q values from these data; we then attempt to estimate the scattering Q, using both a one-dimensional model and a three-dimensional numerical model; we investigate the possible range of values of intrinsic Q, based on the theory of Pointer, Liu and Hudson (2000); finally, we discuss some implications of our results.

VSP DATA ACQUISITION

The near zero-offset VSP profile at Glyvursnes, southern Streymoy, surveyed approximately 300 m of the Upper Basalt Formation and 300 m of the Middle Basalt Formation at 10 m intervals from 50 m to 600 m depth. A two-metre deep pit was excavated down to bedrock, 14 m from the borehole and a single sleeve Sodera 2.46-litre air gun was suspended in the water at a depth of approximately 1.5 m. The borehole receiver was a custom-made I/O 3-component sensor containing SM-7M 10Hz geophones and a hydraulic bow spring was used to clamp the receiver to the borehole wall at a pressure of...
~2000 psi for each depth recording. A Geometrics ES-3000 seismic recording system was used to record the VSP survey at a sampling interval of 0.125 milliseconds and a record length of 1.95 seconds. No filters other than normal anti-alias filters were applied to the raw data at the time of acquisition. A monitor hydrophone was positioned beneath the source pit within a slim hole at a depth of approximately 10 m.

A second near zero-offset VSP experiment was carried out at Vestmanna Sund, north-western Streymoy, through 500 m of the Middle Basalt Formation at 5 m intervals from 50 m to 500 m depth. Two Halliburton sleeve guns, each with a capacity of 0.65 litres, were chained to flotation buoys and suspended at a depth of approximately 1.5 m in a two-metre diameter natural rock pool situated 18 m from the borehole. The same borehole receivers were used as at Glyvursnes, with a sampling interval of 0.5 milliseconds and a record length of 4.995 seconds. No filters other than normal anti-alias filters were applied to the raw data at the time of acquisition. A monitor hydrophone was placed in the borehole at a constant depth of 20 m throughout the experiment.

**ESTIMATION OF EFFECTIVE Q**

The spectral ratio method is widely used to estimate the attenuation of seismic data. It is applied in the frequency domain and compares the spectrum of a seismic wave entering an attenuative medium with its spectrum as it emerges. Q and the propagation velocity are assumed to be independent of frequency. The amplitude of a seismic signal in relation to the source amplitude is given by:

\[ A = A_0 \exp\left(-\frac{\pi f x}{Qc}\right) \]  

where \( A = \text{amplitude at depth } x, A_0 = \text{source amplitude}, f = \text{frequency}, x = \text{depth}, Q = \text{seismic quality factor} \) and \( c = \text{wave propagation velocity} \). Comparing traces from two depths,

\[ \ln \left( \frac{A_2}{A_1} \right) - \ln \left( \frac{A_{02}}{A_{01}} \right) = -\frac{\pi f \delta t_{2-1}}{Q} \]  

where \( \delta t_{2-1} \) is the difference in travel times. If the source is constant throughout the experiment, \( A_{02} = A_{01} \) and effective Q can be derived from a plot of \( \ln \left( \frac{A_2}{A_1} \right) \) against frequency \( f \), using:

\[ Q = -\frac{\pi \delta t_{2-1}}{\text{slope of plot}} \]  

The VSP data were time-aligned, and a five-point median filter applied (Figs 1 and 2). This step suppressed upcoming waves, which may have interfered with the downgoing waves analysed as shown by Stainsby and Worthington (1985). Two seismic traces of interest, separated by a depth interval of \((x_2 - x_1) \) m, were time-gated with the same double-tapered, half-Hanning window. The length of the time gate was determined by visual inspection with the aim, as far as possible, of separating the first arrival from later arriving multiples. Examples are given in Figs 3 and 4. The amplitude spectra were examined to find the frequency range over which spectral ratios were relatively stable. Once determined, both the length of the time gate and the stable frequency range were kept constant for all trace pairs analysed. Least-squares linear regression was performed within the stable frequency range (Figs 3c and 4c), and the corresponding Q value was calculated from equation (3).

Q values were initially derived for the bulk VSP intervals of Vestmanna and Glyvursnes, by comparing wavelets over
nearly the full depth interval of the surveys (Figs 3 and 4). Specifically, we used the interval 60 to 490 m at Vestmanna and 70 to 580 m at Glyvursnes to avoid any possible edge effects resulting from the median filter. Estimates were then obtained for fixed intervals between trace pairs. A relatively large interval of 250 m was chosen to reduce the variance of the Q estimate, as recommended by White (1992). Analysis was repeated from the top to the base of each borehole, thereby tracking Q with depth.

**Vestmanna**

Good quality VSP data were acquired at the Vestmanna field site due to a constant source signature throughout the experiment and a generally low background noise level. A broad frequency band from 20–500 Hz was achieved due to the excellent source coupling to the bedrock. The Q value derived from comparing the uppermost trace and the lowermost trace is 25, with a range of 24–27 for a 95% confidence interval (Fig. 3c). In the analyses over intervals of 250 m, Q values span 20 to 90 (Fig. 5), with a mean of approximately 30. While there are anomalously high outliers, these occurred at depth levels where short-period multiples created numerous minima in the trace spectra, which increases the error margins in linear regression. Under these circumstances, the high values are...
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Figure 4 (a) Downgoing wavelets at 70 m and 580 m in the Glyvursnes borehole. The light dotted curve is the tapered time gate. (b) Amplitude spectra for the traces in (a). (c) Spectral ratio plot. Least-squares linear regression is performed over a specified frequency range of 40–150 Hz. Q estimates are given for a 95% confidence interval of mean response.

likely to reflect large uncertainty rather than actual physical properties. There was no evidence of any significant variation of Q with depth.

Glyvursnes

Although the source pit was excavated to bedrock, it was of necessity constructed in soft, recent deposits. This caused relatively poor source to bedrock coupling and the frequency bandwidth of the Glyvursnes VSP data was narrower than expected. The monitor hydrophone was positioned at too shallow a depth. The hydrophone signal changed from day to day, partly due to fluctuations in the water table and partly due to positioning errors. The signals bore little relation to the actual downhole signals and consequently were used neither for source correction nor in further analyses. In contrast, first arrivals in the geophone records are very regular and remain consistent throughout the profile (see Fig. 2). This suggests that despite some fluctuations of the water level in the pit due to leakage, the source signature was largely constant throughout the experiment. Over the region where log spectral ratios were stable, 40–150 Hz, Q derived from the uppermost trace and the lowermost trace is 24, with a range of 17–36 for a 95% confidence interval (Fig. 4c). In the analyses over intervals of 250 m, Q values span 5 to 120, with a mean Q value of approximately 25 (Fig. 6). Estimates from the shallow and deep sections of the borehole suggest there is an increase in Q with depth: between 50 and 350 m, mean and modal Q value is ∼15, while between 360 and 600 m, the mean increases to ∼30 (Fig. 7).

Figure 5 Histogram of effective Q measurements in the Vestmanna data, derived using the spectral ratio method, for a depth separation of 250 m between traces.

ESTIMATION OF 1D SCATTERING Q

Rocks in the field area display distinct horizontal layering, and 1D scattering is a candidate mechanism for the low effective Q values observed in real VSP data. Density and P-velocity
Figure 6 Histogram of effective Q measurements in the Glyvursnes data, derived by the spectral ratio method, for a depth separation of 250 m between traces.

Figure 7 (a) Histogram of effective Q measurements in Glyvursnes, 50–320 m. (b) Histogram of effective Q measurements in Glyvursnes, 320–600 m.

The data shown in Fig. 8 display strong fluctuations, which potentially could create much multiple interference. The frequency and magnitude of these fluctuations are noticeably different in the top half of the Glyvursnes hole, which coincides with part of the Upper Basalt Formation, compared to the remainder of the logs. A corresponding difference in 1D scattering attenuation should arise. To test this hypothesis, synthetic seismograms were computed using an algorithm devised by Kennett (1979), for the case of an incident plane wave at normal incidence to the plane layered medium. 1D plane-layered models were constructed from density and P-velocity log data with a constant layer thickness of 0.2 m. Each layer was isotropic and had an intrinsic Q of 1000 (effectively non-attenuative).

P-wave seismograms were generated for the same intervals as the VSP surveys (60–490 m for Vestmanna and 70–580 m for Glyvursnes) using the primary wavelets of the respective uppermost VSP traces as input. The resulting output wavelets were then isolated with the same half-Hanning windows as in previous relevant VSP analyses. In application of the spectral ratio method, linear regression by least squares was also performed over the same stable bandwidth ranges as with the analysis of the real VSP data. The results are shown in Figs 9 and 10. For a 95% confidence interval, the derived 1D scattering Q for Glyvursnes ranges between 19 and 87, with a mean of ∼31. For Vestmanna, Q ranges between 75 and 93, with a mean value of ∼84.

The upper and lower halves of the 1D models were then analysed separately to determine whether the change in character of the Glyvursnes log resulted in a change of scattering attenuation. Isolated primary wavelets from the relevant topmost VSP traces were used as sources. Modelling was also carried out with synthetic zero-phase sources: a wavelet with central frequency of 150 Hz and a broadband wavelet with content up to 800 Hz. The results are summarized in Table 1 and Fig. 11 shows the input and output wavelets of these zero-phase sources. Short lag multiples, which fall within the
analysis window, are observed in the output broadband data. The range of values in Table 1 illustrates the importance of wavelet shape and signal bandwidth in Q estimation. The spectral ratio method provides a linear estimate over a specified bandwidth. Short lag multiples can result in spectral minima, which may skew the Q estimates depending on what bandwidth is chosen. However, the main conclusion from Table 1 is that 1D scattering is a plausible explanation for the observed seismic attenuation in rocks corresponding to the Upper Formation in the Glyvursnes borehole, but is less likely to fully explain our observed data in the other Faroes rocks studied.

**ESTIMATION OF 3D SCATTERING Q**

Having established that predicted 1D scattering cannot fully account for the effective Q values seen in most of the VSP data, we investigated whether significant apparent attenuation might result from seismic wave scattering due to three-dimensional variations in basalt structure. The main shortcoming of our study is that we have no way of determining the true three-dimensional structure within the immediate vicinity of the two boreholes. Any model constructed is thus hypothetical, based on field evidence of basalt structure and which we believe reflects the extent of the lateral variation that is likely to exist.

The Faroe Plateau Lava Group, which erupted in a marginal marine or sub-aerial environment (Ellis et al. 2002), is thought to comprise two main morphological types (S.R. Passey, quoted in Bondre, Duraiswami and Dole 2004) of which there is limited exposure in the field area of the VSP experiments. The first type is a massive tabular lithology with a low spatial frequency, akin to that described by Jerram (2002). This

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**Figure 8** P-velocity and density logs from Vestmanna and Glyvursnes.
Figure 9 (a) Input and output primary wavelets of synthetic seismograms, for a 1D model of the 60–490 m interval of the Vestmanna velocity and density logs. The dotted curve is the tapered time gate. (b) Amplitude spectra for the traces in (a). (c) Spectral ratio plot. Least-squares linear regression is performed over a specified frequency range of 20–400 Hz. Q estimates are given for a 95% confidence interval of mean response.

Figure 10 (a) Input and output primary wavelets of synthetic seismograms, for a 1D model of the 70–580 m interval of the Glyvursnes velocity and density logs. The light dotted curve is the tapered time gate. (b) Amplitude spectra for the traces in (a). (c) Spectral ratio plot. Least-squares linear regression is performed over a specified frequency range of 40–150 Hz. Q estimates are given for a 95% confidence interval of mean response.

type is predominant in the Lower and Upper Formations in the Faroes. The second is composed of hummocky flows between 1 and 5 metres in thickness. These are akin to compound braided basalts described by Jerram (2002) and may be tube-fed (Bondre et al. 2004). In cross-section, they may appear as lobes tens of metres in length and are thought to be the major morphological type in the Middle Formation and thus in the Vestmanna locality. Over a range of wavelengths, these hummocky flows are more internally varied, possibly giving rise to more 3D scattering than the tabular form.

A routine that propagates synthetic basalt flows of a specified thickness and width over a model of existing terrain was developed, in order to build a 3D model based on the hummocky basalt type. In the routine, a multifractal topography with low relief (<5 m) is used as the first surface and successive flows are built up. Each flow starts from the lowest point on the southern edge of the model and moves towards the north.
A 3D discrete numerical elastic lattice method was used to simulate elastic wave propagation through the basalt models. The numerical model consists of particles representing blocks of intact rock, which are arranged on a cubic lattice and which interact with Hooke’s Law forces. The algorithm is described in detail by O’Brien and Bean (2004), with supporting theoretical and practical detail to be found in Monette and Anderson (1994) and Toomey and Bean (2000). A source was input as a force in the z (depth) direction on the particle in the top centre of the basalt model. The peak and centre frequency of the source wavelet was 150 Hz with a main data-carrying bandwidth of 10 to 250 Hz. The time sample interval was 0.1 ms. Both the sample interval and grid spacing (1 m) were chosen to avoid the problems of numerical instability and dispersion. Poisson’s ratio was fixed at 0.25 by setting the bond-bending constant to zero as described in O’Brien and Bean (2004).

Within the restraints of available computing resources, synthetic seismic data were generated for a 280 × 280 × 400 m version of this model. Seismograms were also generated for its 2D cross-sections, and for a homogeneous 3D volume with a velocity of 5250 m/s. These were used to establish that in a vertical VSP, normalized amplitudes from the 3D volume can be approximated by taking the average of normalized amplitudes from their orthogonal 2D cross-sections (Fig. 12). Seismograms for larger 2D cross-sections of the model (1000 × 600 m, in planes parallel and perpendicular to flow direction) were then computed. Their wavefields captured after 20 ms are shown in Figs 13 and 14.

To estimate the effect of moving from a 1D to 2D pahoehoe scattering scenario, vertical velocity profiles were taken through the centre of the 2D models and acoustically modelled as 1D, planar structures. The synthetic seismograms were then compared using the spectral ratio method (Fig. 15). While a more complex 2D structure shifts the positions of minima in the spectra, the associated Q values (calculated within a 95% confidence interval) range from 66 to 109 in the plane parallel to lava flow and 159 to 226 in the plane perpendicular to lava flow. These cannot account for low Q values estimated from the field data.

A simple comparison of amplitude loss from the first arrivals supports the above conclusion. Varying velocity structure perturbs the propagating wavefields (Figs 13b and 14b), but does not induce significantly greater amplitude loss than the homogeneous model after nearly 600 m of propagation (Fig. 16).

### ESTIMATION OF INTRINSIC Q

If the observed data cannot be explained entirely by some form of elastic wave scattering, then an anelastic mechanism is the only alternative explanation. Using the simple inverse relationship between effective, intrinsic and scattering Q mentioned

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**Table 1** Estimated 1D scattering Q values. The range is given by a 95% confidence interval of the mean response. Mean Q values are in bold type

<table>
<thead>
<tr>
<th>Borehole section</th>
<th>Input wavelet</th>
<th>Peak frequency</th>
<th>Broadband</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(frequency range in (Q analysis)</td>
<td>(150 Hz, zero-phase)</td>
<td>(800 Hz, zero-phase)</td>
</tr>
<tr>
<td>Vestmanna 60–490 m</td>
<td>(20–500 Hz)</td>
<td>72–81–96</td>
<td>86–94–102</td>
</tr>
<tr>
<td>Vestmanna 60–275 m</td>
<td>(20–500 Hz)</td>
<td>(10–250 Hz)</td>
<td>(10–800 Hz)</td>
</tr>
<tr>
<td>Vestmanna 275–490 m</td>
<td>(20–500 Hz)</td>
<td>47–52–60</td>
<td>84–92–103</td>
</tr>
<tr>
<td>Glyvursnes 70–580 m</td>
<td>(30–150 Hz)</td>
<td>28–33–46</td>
<td>118–171–302</td>
</tr>
<tr>
<td>Glyvursnes 70–320 m</td>
<td>(30–150 Hz)</td>
<td>(10–250 Hz)</td>
<td>(10–800 Hz)</td>
</tr>
</tbody>
</table>

Due to important morphological and thermodynamic distinctions between ‘ponded flows’ and pahoehoe (Self, Keszthelyi and Thorarinson 1998), the flows are allowed to meander towards lower topography but are not allowed to pool laterally or terminate before reaching the northern side. This simulates pahoehoe ‘tubes’ of fairly regular size but with varying sinuosity and at varying positions, each constrained by a thin crust. The resulting structure is heterogeneous and anisotropic, with greater lateral continuity in the plane parallel to the flow direction. Vestmanna P-velocity log values (between 50 and 600 m depth) were separated into two categories: crust-range (low velocity) and core-range (high velocity). Each flow tube was assigned a core velocity and a crust velocity from these respective lists. This results in a realistic distribution of velocities, although the model has only weak depth correlation with the actual Vestmanna rock volume. Density was set according to:

\[ \rho = 0.15 V_p + 2000 \text{ kg m}^{-3} \]  

based on an approximate linear relationship established from log data.
above, we deduce that required values of intrinsic Q vary from infinity in the top half of the Glyvursnes sequence, where there is little difference between our estimates of effective and scattering Q, to as low as approximately 45 at Vestmanna. These numbers vary greatly within the uncertainty limits of our estimates. For example, it is possible to attribute the observed effective Q of 20 in the shallower sections at Glyvursnes to a combination of scattering Q and intrinsic Q, both with a value of 40. It is necessary to show that values of intrinsic Q as low as 40 are realistic on the basis of existing theory and our knowledge of the petrophysical properties of Faroes basalts.

Fluctuating stresses in a rock caused by the passage of a seismic wave induce pore pressure gradients at the scale of the seismic wavelength and also at the scale of individual grains, pores and micro-cracks. Mavko and Jizba (1991) and Dvorkin, Mavko and Nur (1995) described how seismic attenuation results from the squirt flow of fluid between pores and micro-cracks. However, doubts have been expressed about the magnitude of squirt flow induced attenuation at exploration seismic frequencies (Gist 1994; Pride et al. 2003). Thomsen (1995) proposed the concept of equant porosity in which wave induced fluid diffusion occurs from the cracks into a porous and permeable matrix material. Theories of wave propagation in cracked media developed by Hudson, Liu and Crampin (1996), Pointer et al. (2000) and Chapman (2003) include the equant porosity model. The frequency at which seismic attenuation is at maximum is dependent on the relaxation time of the fluid diffusion. We have used the randomly oriented

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**Figure 11** Results of acoustic modelling of the Faroes borehole logs with synthetic source wavelets. Input primary wavelets are plotted in grey. Output primary wavelets are overlaid in black.
micro-cracks with equant porosity model proposed by Pointer et al. (2000), to calculate values of intrinsic attenuation. Random crack orientation was chosen for our models because we have no means of determining the orientation of in-situ grain scale micro-cracks. We emphasize that the results we obtain are model dependent and non-unique. Our sole objective is to demonstrate that there is no fundamental theoretical objection to the proposition that intrinsic attenuation might be an important mechanism in Faroe Island basalts.

Provisional results from the analysis of 28 core samples and neutron porosity logs from the Glyvursnes and Vestmanna holes by members of the SeiFaBa research group include estimates of porosity and permeability (Japsen and Waagstein, pers. comm.). The basalt sections can be simply classified as consisting of lava core and lava crust or breccia. Very approximately, there are equal proportions of these lithologies in the two holes. Average porosities from the core analysis and the neutron porosity logs calibrated by the core data are 4% for the lava core and 15% for the lava crust. Core sample estimates of permeability for lava core and lava crust are $10^{-17}$ m$^2$ and $5.0 \cdot 10^{-16}$ m$^2$ respectively. For our modelling, we have assumed an average porosity of 10% and a permeability range from $10^{-17}$ to $10^{-16}$ m$^2$. The rocks are assumed to be 100% water saturated. Other known parameters used in the modelling are listed in Table 2. An average $V_p/V_s$ ratio of 1.88 was obtained from an analysis of the VSP data (Shaw 2006). Values of crack radius, crack density and crack aspect ratio are not known. The chosen value of crack density is the approximate mean of values tabulated in a review of experimental data by Crampin (1994), and is consistent with the laboratory data of Peacock et al. (1994).

Figure 17 shows curves of the variation of seismic attenuation with frequency for a range of values of crack radius, permeability and crack aspect ratio. The amplitude of the peaks
is governed by crack density and crack aspect ratio. Curve 1 was obtained with the fixed parameters in Table 2, a permeability of $10^{-17}$ m$^2$, a crack radius of 0.001 m and a crack aspect ratio of 0.01. For curve 2, the crack aspect ratio is increased to 0.03. As might be expected, the rock becomes less compliant as the crack aspect ratio increases and consequently the intrinsic attenuation is reduced. Peak values of $Q$ for curves 1 and 2 are 40 and 55 respectively. In the review of micro-cracks in rocks by Kranz (1983) values of aspect ratio of 0.001 or less are quoted. So our chosen values should produce an underestimate of the possible attenuation. The frequencies corresponding to the peaks in the attenuation curves are equal to the inverse of the relaxation time of the fluid flow, which is proportional to (permeability/crack radius$^2$). So curve 2 could have been obtained with a permeability of $10^{-16}$ m$^2$ and a crack radius of 0.003 m or a permeability of $10^{-17}$ m$^2$ and a crack radius of 0.001 m. Curve 3 was obtained with a permeability of $10^{-17}$ m$^2$, a crack radius of 0.001 m and a crack aspect ratio of 0.03. Values of crack radius have been chosen so that the calculated curves fall within the frequency range of our field data. Additionally, they fall within the range of values quoted by Kranz (1983).

Velocity dispersion of between 100 and 150 m/s is also predicted over a frequency range from 100 Hz to 24 kHz. The latter value is the dominant frequency of our sonic logging tool. In principle, this could be detected by comparing the VSP travel times with integrated log times. Unfortunately, we experienced considerable difficulties with the calibration of our sonic log data and it was not possible to estimate...
DISCUSSION

Our attempt to separate and quantify the effects of scattering and intrinsic attenuation is based on a number of assumptions that should be questioned. Our use of the spectral ratio technique to obtain a single frequency independent effective Q value is common practice but is not entirely satisfactory. Q is known to be a frequency dependent quantity both from theory, as illustrated in Fig. 17, and from field experiments (Sams et al. 1997). Apparent attenuation due to scattering within a plane-layered medium is also frequency dependent (Menke 1983). The usual assumption made is that the frequency dependence is not so strong as to preclude the use of a constant Q when interpreting sufficiently narrow band data. However, our Vestmanna data are relatively broad band. The reason that we see no clear evidence of frequency dependence of Q in our Vestmanna data is partly because of the uncertainty in our Q estimates, as illustrated by the histograms in Figs 5, 6 and 7, but also because our attenuation models are probably too simplistic. For example, it has long been recognized (Liu, Anderson and Kanamori 1976) that constant Q might result from the superposition of a number of relaxation peaks of the type shown in Fig. 17.

Our analysis of 3D scattering attenuation is based on only one hypothetical model. However, we believe that we have constructed an upper limit of likely lateral variability and consequently, our conclusion about the effect of 3D scattering is conservative. Nevertheless, this is not proof and further experimentation could reveal plausible 3D structures that result in significant scattering attenuation.

Our study of intrinsic attenuation demonstrates that this could be a significant mechanism in Faroes basalts but does not prove the point. We can only suggest that our estimated values of crack density, crack aspect ratio and crack radius are realistic by referring to field data reviewed by Crampin (1994), the laboratory data of Peacock et al. (1994) and the extensive review of micro-cracks by Kranz (1983). Our assumption that the cracks are randomly oriented is conservative. Intrinsic attenuation would increase significantly if randomly oriented cracks became aligned, with a resulting P-velocity anisotropy of 1–2%. In other regions where velocity anisotropy is detected, one might reasonably expect relatively low values of intrinsic Q.
CONCLUSIONS

Values of seismic effective Q determined from VSP data acquired in wells at Lopra, Glyvursnes and Vestmanna in the Faroe Islands are all sufficiently low to be a dominant cause of the poor quality of seismic reflection images. However, the possible causes of the low effective Q have been found to vary between the three borehole localities. Christie et al. (2006) conclude that 1D scattering is the dominant attenuation mechanism at Lopra. It is possible to come to the same conclusion for the upper basalt formation in the top half of the Glyvursnes hole. However, within the middle basalt formation in the bottom half of the Glyvursnes hole and at Vestmanna, little attenuation can be attributed to 1D scattering. Results from 3D elastic wave numerical modelling with a hypothetical basalt model constructed on the basis of field observations, indicate that little scattering attenuation is caused by lateral variations in basalt structure. Maresh et al. (2006) came to a similar conclusion that 3D scattering was not significant using a phase-screen modelling technique on data across basalt flows in Rockall trough.

However, the low values of effective Q observed at Glyvursnes and Vestmanna can be explained as resulting from a combination of 1D scattering and intrinsic attenuation due to seismic wave induced fluid flow within pores and microcracks. This study adds to the now substantial number of observations of seismic attenuation within basalt sequences in the North Atlantic region and reinforces the general conclusion that low values of effective Q are to be expected. However, the causes of the high attenuation may vary greatly from one locality to another.

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Figure 17 Modelled values of intrinsic seismic attenuation (1/Q) as a function of frequency, using the randomly oriented cracks with equant porosity model of Pointer et al. (2000). Peak values of Q for curves 1 and 2 are 40 and 55 respectively. Fixed model parameters are listed in Table 2. Values of permeability (m³), crack radius (m) and crack aspect ratio respectively are: (Curve 1) 10⁻¹⁷, 0.001, 0.01, (Curve 2) 10⁻¹⁷, 0.001, 0.03, (Curve 3) 10⁻¹⁸, 0.001, 0.03.
University). We thank Tim Pointer for generously sending us copies of his software. Any possible misuse of this software is entirely our responsibility.

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Shetland Basin with large offset data.


