Transitional continental–oceanic structure beneath the Norwegian Sea from inversion of surface wave group velocity data

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SUMMARY

We have analysed the fundamental mode of Love and Rayleigh waves generated by 12 earthquakes located in the mid-Atlantic ridge and Jan Mayen fracture zone. Using the multiple filter analysis technique, we isolated the Rayleigh and Love wave group velocities for periods between 10 and 50 s. The surface wave propagation paths were divided into five groups, and average group velocities calculated for each group. The average group velocities were inverted and produced shear wave velocity models that correspond to a quasi-continental oceanic structure in the Greenland–Norwegian Sea region. Although resolution is poor at shallow depth, we obtained crustal thickness values of about 18 km in the Norwegian Sea area and 9 km in the region between Svalbard and Iceland. The abnormally thick crust in the Norwegian Sea area is ascribed to magmatic underplating and the thermal blanketing effect of sedimentary layers. Maximum crustal shear velocities vary between 3.5 and 3.9 km s\(^{-1}\) for most paths. An average lithospheric thickness of 60 km was observed, which is lower than expected for oceanic-type structure of similar age. We also observed low shear wave velocities in the lower crust and upper mantle. We suggest that high heat flow extending to depths of about 30 km beneath the surface can account for the thin lithosphere and observed low velocities. Anisotropy coefficients of 1–5 per cent in the shallow layers and >7 per cent in the upper mantle point to the existence of polarization anisotropy in the region.

Key words: anisotropy, crust, lithosphere, Norwegian Sea, seismic resolution, surface waves.

INTRODUCTION

The Norwegian–Greenland Sea region (Fig. 1) is tectonically complex. Whilst the seafloor spreading process along the ridges north of the Jan Mayen fracture zone proceeded symmetrically, the process south of the zone has been characterized by westward shifts of the spreading axis (Eldholm et al. 1990). This has resulted in extinct ridges scattered through the southern part of the study area. The Jan Mayen, Greenland and Senja fracture zones intersect the area (Fig. 1), and the Jan Mayen Ridge micro-continent further complicates the tectonics. Oceanic basins containing thick layers of sedimentary rocks are found east and west of the mid-Atlantic ridge. Those east of the mid-Atlantic ridge are bounded on their eastern margins by the Voring plateau and Faeroe–Shetland escarpments. The Voring and Greenland–Faeroe plateaux were formed by the Iceland hot spot (Vink 1984).

Reflections observed beneath the smooth acoustic basement of the Voring plateau and in the Lofoten basin (Eldholm & Sundvor 1980) provide evidence of further complications to the crustal structure of the region. Deep sea drilling projects have found that the reflections correlate with the top of basalts and that these basalts were extruded at very high rates during the initial stage of spreading in the area (Eldholm & Sundvor 1980). Although extensive studies have been carried out to determine the crustal thickness and velocity structure of the Norwegian continental margin (e.g. Chroston & Brooks 1989; Planke et al. 1991; Mjelde et al. 1992; Kodaira et al. 1995, 1998), very little work appears to have been undertaken on the lithospheric structure (especially the upper mantle) of the margin and Norwegian Sea area. Until now, surface wave studies have been impractical since most of the earthquake events that occur in the vicinity of this region are small (M < 5.0). Published studies concentrating on the lithospheric structure are of regions outside our area of interest (e.g. Chan & Mitchell 1985; Clark 1983; Evans & Sacks 1979, 1980; Léveque 1980; Calcagnile & Panza 1978). This study attempts...
to estimate the crustal and upper mantle structure of this region using group velocities of the fundamental mode of Love and Rayleigh waves. This is achieved by simultaneously inverting Rayleigh and Love wave group velocities. Owing to its simplicity, this method has been widely used to estimate crust and upper mantle structure. We did not take into account the effect of anisotropy, but performed a simple test to assess the extent to which polarization anisotropy exists in the region of study. We believe that the results obtained in this study constitute a valuable basis for future velocity structure studies in this region.

SEISMOLOGICAL DATA

The region shows moderate seismicity, with the highest concentration of epicentres along the mid-Atlantic ridge (MAR) and associated with fracture zones (e.g. the Jan Mayen fracture zone). High-quality long-period and broad-band records of these earthquakes are available from seismic stations operated by the University of Bergen (UiB) and the Norwegian Seismic Array (NORSAR).

We selected 12 earthquakes (Table 1) recorded at five seismic stations (Fig. 2). Of the five stations, Kongsberg (KONO) and Kings Bay (KBS) are operated by the seismology group at UiB, whilst the other three are part of the NORESS (NRS), ARCESS (ARC) and Svalbard (SPB) arrays operated by the Norwegian Seismic Array group (NORSAR).

We obtained three-component long-period data from all five stations. In addition, KONO station and the three arrays operate three-component broad-band stations. The magnitude of the earthquakes selected is such that the surface waves produced at the recording stations are not very large. However,
Figure 2. Locations of the recording seismic stations and earthquake events used in the study. The straight lines indicate the approximate paths between the events and stations. (Units are in degrees.)

Table 1. Parameters of earthquakes used in the study.

| No. | Date (UTC) | Epicentre Lat. | Long. | Depth (Km) | Magnitude |
|-----|------------|----------------|-------|------------|-----------|
| 1   | 1992 0909 1308 57.3 | 76.370 | 10.790 | 02.00 | 3.7 |
| 2   | 1994 0208 0327 54.4 | 66.240 | -19.330 | 05.50 | 4.9 |
| 3   | 1995 1004 0917 32.7 | 75.948 | 10.289 | 09.90 | 4.2 |
| 4   | 1996 0315 0944 53.4 | 55.290 | -19.400 | 20.50 | 5.7 |
| 5   | 1996 0407 0432 16.2 | 54.680 | -23.080 | 10.00 | 5.2 |
| 6   | 1996 0411 1051 17.0 | 56.980 | -33.750 | 10.00 | 5.1 |
| 7   | 1996 0511 0438 42.2 | 80.680 | -90.090 | 29.00 | 4.5 |
| 8   | 1996 0820 0011 02.9 | 78.070 | 07.550 | 10.00 | 4.2 |
| 9   | 1996 0929 1048 00.2 | 63.790 | -20.920 | 10.00 | 4.9 |
| 10  | 1996 1225 1735 27.9 | 71.100 | -98.950 | 15.19 | 5.0 |
| 11  | 1997 0516 1431 26.6 | 71.790 | -01.730 | 10.00 | 3.9 |
| 12  | 1997 0722 1621 48.5 | 66.650 | -17.640 | 10.00 | 4.7 |

they are recorded with a good signal-to-noise ratio, and show dispersion.

Fig. 3 shows a typical example of the spectrum obtained from the transverse component of the earthquake event 12 (Table 1) recorded at KONO. The signal-to-noise ratio is good between 0.01 and 0.2 Hz. This record is typical of those used in the dispersion analysis.

DISPERSION ANALYSIS

Fundamental mode Love and Rayleigh waves were extracted using the multiple filter technique (MFT) to isolate the group velocities (Dziewonski et al. 1969; Herrmann 1973). The Rayleigh wave group velocities were obtained from the radial and vertical components, and the Love wave velocities from the transverse component. Instrument responses were removed using the SEISAN software package (Havskov 1997). The MFT was then applied to the corrected amplitude spectra yielding a plot of amplitude as a function of group velocity and period (Fig. 4). Fig. 4 shows a typical example obtained from the long-period vertical component of an event recorded at the
Figure 3. Original seismogram (a) and velocity spectrum (b) obtained from the transverse component of event 12 (Table 1) recorded at station Kongsberg (KONO), showing a good signal-to-noise ratio in the range 0.01–0.2 Hz. The upper curve in (b) represents the signal and the lower curve, noise.
Transitional shear structure beneath the Norwegian Sea

Because of the tectonic complexity of the region and to reduce the effects of lateral structural differences, we grouped together paths passing through similar structures and travelling in approximately the same direction. Thus, the propagation paths shown in Fig. 2 are divided into five groups: W, W1, W2, W3 and W4 (Table 2). The propagation paths of group W traverse the region between Iceland and Svalbard, crossing the Iceland plateau. Group W1 paths cover a region including the Lofoten and Greenland basins, and also cross the Senja fracture zone and the Mohrs and Knipovich ridges. W2 propagation paths cross the region between Iceland and the Norwegian mainland including the Iceland–Faeroe ridge and Faeroe–Shetland escarpment. The paths of groups W3 and W4 traverse regions with similar structure, the difference between the groups being in the orientation of the paths as shown in Fig. 2. W3 paths cross the Lofoten basin as well as

Table 2. The propagation paths between epicentres and stations, divided into the five groups W, W1, W2, W3 and W4. (The numbers correspond to the epicentre order as in Table 1.)

| Groups | Propagation Paths |
|--------|-------------------|
| W      | 9 - SPB           |
| W1     | 1 - ARC           |
| W2     | 2 - NRS           |
| W3     | 3 - ARC           |
| W4     | 4 - NRS           |

Table 3. Layer parameters of the three initial models used in the inversion for the shear wave velocities.

| Layer thickness (km) | P wave velocity (km s⁻¹) | S wave velocity (km s⁻¹) | Density (gm cm⁻³) |
|----------------------|--------------------------|--------------------------|------------------|
| 1                    | 1.50                     | 0.00                     | 1.00             |
| 2                    | 3.00                     | 2.10                     | 2.74             |
| 3                    | 6.20                     | 3.50                     | 3.75             |
| 4                    | 7.10                     | 3.80                     | 2.85             |
| 5                    | 8.15                     | 4.44                     | 3.40             |
| 6                    | 8.25                     | 4.64                     | 3.50             |
| 7                    | 8.35                     | 4.84                     | 3.50             |
| 8                    | 8.40                     | 4.94                     | 3.50             |
| 9                    | 8.50                     | 4.94                     | 3.60             |
| 10                   | 7.60                     | 4.37                     | 3.60             |
| 11                   | 7.70                     | 4.43                     | 3.60             |
| 12                   | 7.80                     | 4.50                     | 3.60             |
| 13                   | 7.90                     | 4.56                     | 3.60             |
| 14                   | 8.00                     | 4.62                     | 3.60             |
| 15                   | 8.10                     | 4.68                     | 3.60             |
| 16                   | 8.20                     | 4.75                     | 3.60             |
| 17                   | 8.30                     | 4.81                     | 3.60             |
| 18                   | 8.40                     | 4.87                     | 3.60             |
| 19                   | 8.50                     | 4.93                     | 3.60             |
| 20                   | 8.60                     | 5.00                     | 3.60             |
the sheared ridges in the region, whilst those from group W4 cross the Norway basin, Jan Mayen fracture zone and the Iceland–Faeroe ridge. Henceforth, the regions will be referred to by the name of the group of paths crossing them. For example, the region traversed by paths making up group W will be referred to as region W.

As a result of the different structures across which all these paths travel, the estimated group velocities are an average of the velocities along each path. For each group of paths (W–W4), we calculated the average of the group velocities for each period. For each mean velocity, we calculated the standard deviation using the standard relation.

Figure 5. Theoretical and observed group velocities for the best-fitting model from Love and Rayleigh waves (Love: Love wave group velocities; Ray: Rayleigh wave group velocities), for the regions W, W1, W2, W3 and W4. (Continuous and broken lines: computed Rayleigh and Love wave group (model) velocities respectively; vertical bars: standard deviations.)

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INVERSION OF GROUP VELOCITY DATA

The shear wave velocity models were estimated by simultaneously inverting Love and Rayleigh wave group velocities using the Herrmann (1987) software package. The inversion subroutines given by Lawson & Hanson (1974) and used in this package are based on a general form of the stochastic inversion procedure which is equivalent to the damped least-squares method (Herrmann 1987). Details of this method are discussed in Jackson (1972) and Wiggins (1972). A basic outline is given below.

A starting model is selected and is used to compute the dispersion curves and partial derivatives of the group velocities with respect to the unknowns. The dispersion curves are compared with the observed data, and model corrections are computed. After applying first-order corrections to the starting model, the whole process is repeated until the changes in the model are small. In our calculations, the inversion was terminated once the model rms error was small (e.g. 0.0001) and had stopped improving.

A simple oceanic model was selected as the principal starting model. To test for stability, two further models (Table 3) were also used as initial models. We inverted the group velocities for both shear wave velocity and depth of interface. Initially, the layer thicknesses were held constant whilst inverting for velocity. Inversion for interface depth was only undertaken at the final stage once the computed dispersion curves for both Love and Rayleigh waves fitted well within the data uncertainties of the observed dispersion values (Fig. 5).

The standard deviation for the shear wave velocity in each layer of the model was calculated using the relation given by Mitchell (1976) for the stochastic least-squares approach to inversion:

$$\sigma^2 = \left[ \sigma^2 W_i \sum_{j=1}^{n} \frac{V_j^2}{\lambda_j^2 + \sigma^2} \right]^{1/2},$$

where $W_i$ is the component of the weighting matrix, inversely proportional to the layer thicknesses in the model (Mitchell 1976), $\lambda_j$ is the damping factor, $\sigma^2$ is an adjustable parameter called the problem variance, and $V_j$ are the eigenvectors associated with the rows of $A$, where $A$ is the matrix containing the partial derivatives of group velocities with respect to the shear wave velocity (Rodi et al. 1975). The reliability of the estimated models is shown by resolution kernels obtained from the rows of the resolution matrix (Jackson 1972; Wiggins 1972).

RESULTS AND DISCUSSION

Computed and observed group velocities for both Love and Rayleigh waves are shown in Fig. 5. The computed group velocities generally fall within the standard deviation of the observed velocities. Since anisotropic effects are not accounted for, standard deviation values of the observed velocities in all the groups except group W are large. Even in group W, the standard deviations of the velocities at short periods are large. These, however, decrease with increasing period, reflecting the homogeneity of the velocity structure with depth. Because of the small size of the earthquakes used (Table 1), the longer periods are less reliable since the amplitudes at these periods are small. Christensen et al. (1980) observed a similar effect when they used events of similar magnitudes to study the deep structure beneath the Atlantic Ocean. Therefore, to reduce the impact of this effect, only periods up to 50 s were used in this study.

The group velocities observed over the whole region are generally lower than the North Atlantic Ocean values obtained by Christensen et al. (1980), but are similar to those obtained for the Iceland plateau by Evans & Sacks (1979). The exception is W1 group velocities, which are very low compared with velocities observed in the other four groups. The maximum velocities are less than 3.5 km s$^{-1}$. The low velocities, especially at short periods, may be a reflection of the thick sedimentary layers (up to 10 km) in this part of the region.

The group velocities were inverted to obtain the shear wave velocity models shown in Fig. 6. The top layer, which represents the water depth, is given an average thickness of 1.5 km for the whole region. Although we also inverted for depth, most of the layer thicknesses do not show much change from those of the initial models. Most of the observed changes are in the shear velocities. An examination of the models along with their resolving kernels (Fig. 7) enables one to judge where features in the models are artefacts of the inversion process. The resolution kernels plotted in Fig. 7 correspond to the velocity models for the regions traversed by the paths making up groups W, W2, W3 and W4. The kernels shown in the plots are quite similar, implying that resolution is uniform across the whole region. The maxima of the curves coincide with the great depths (< 20 km) and at great depths (> 100 km) is wider than the layer thicknesses at the same depths, implying that the resolution of layer structure at those depths is poor. Thus the results obtained here are most reliable for the sub-Moho velocity structure. The model layer standard deviations (horizontal bars on model layers in Fig. 7) seem to reflect the standard deviations observed in the group velocity data. This is illustrated by the low group velocity standard deviation for group W, which corresponds to the low model layer standard deviation shown in Fig. 7.
Figure 6. Shear wave velocity models resulting from the simultaneous inversion of Love and Rayleigh wave group velocity data for the four different regions traversed by paths in groups W, W2, W3, W4. The thin lines represent the three models obtained by inverting from the three initial models for each region, and the thick line is an average of the three models in each group of paths.

believe that the scatter is due to azimuthal anisotropy, which is not accounted for here. The effect is low in the region traversed by paths making up group W, as shown by the low standard deviations.

The shear velocity model for region W is very similar to young oceanic-type structure, especially as regards the shear velocity values in the lithosphere. The thickness of the crust is approximately 10 km. This value must be regarded as approximate since resolution is poor at this depth. The lithospheric thickness value of about 60 km is, however, larger than expected when compared to results obtained by Evans & Sacks (1980). Since this is an average value depending on the structures crossed by paths making up group W, the thick lithosphere beneath the Iceland plateau (Evans & Sacks 1979) may well have influenced the results for this group. The final model for region W1 appears to be quite unrealistic and thus it was not included in the final results. The very low group velocity values obtained are reflected in the unrealistically low shear wave velocities in the model.

The models for regions W2, W3, and W4 differ significantly in upper lithospheric structure from a typical oceanic structure. The maximum shear wave velocities in the crust vary between...
3.5 and 3.9 km s\(^{-1}\). These are quite low compared with expected oceanic values (i.e. greater than 4.0 km s\(^{-1}\)). The crustal thicknesses for regions W2, W3 and W4 were estimated at about 20 km. This value is higher than typical oceanic-type structure, but comparable to values obtained by other studies. By inverting surface wave group velocities, Clark (1983) found crustal thickness values of about 20–24 km in the Iceland–Faeroe region. Other geophysical studies by Chroston & Brooks (1989), Planke et al. (1991) and Mjelde et al. (1992) suggest values of 25–30 km in the Lofoten basin, whilst Evans & Sacks (1979) obtained crustal thickness values of 15–20 km for the Iceland plateau.

Results from several studies on the evolution of the Norwegian Sea seem to explain these abnormal crustal thickness values. Vink (1984) proposed the hot-spot-fed rise model, with the Iceland hot spot producing vast quantities of basalt, which cooled to form areas of thick crust and plateaux in the region. Færseth et al. (1995) and Talwani & Eldholm (1977), among others, attribute the thicker than normal crust to extension processes which resulted in the creation of the Norwegian Sea.

The presence of deep post-Jurassic sedimentary basins and thick sedimentary layers on highs as well as in the Barents Sea area is reflected in the low crustal velocities and can also explain the thick crust. High heat flow in the region (Cermak & Hurtig 1979) and the thermal blanketing effect of sedimentary layers result in the metamorphism of existing sedimentary rocks, thereby increasing the thickness of the crust (Singh 1988).

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**Figure 7.** Resolving kernels (right side) for the best-fitting models (left side) for regions traversed by paths in groups W, W2, W3 and W4.
Lithospheric thickness and upper mantle shear wave velocities are much better resolved than crustal values. All the models exhibit a lithospheric thickness of about 60 km. Between approximately 10 and 30 km below the surface, the shear velocities observed in the models for regions W2, W3 and W4 are low, varying between 3.5 and 4.5 km s\(^{-1}\). Below the lid is a low-velocity zone, the bottom of which is not resolved accurately.

In Fig 8, the shear velocity models obtained in this study are compared with three other shear velocity models, namely the Arabian Sea model (Singh 1988), a 45 Myr old oceanic crust (Evans & Sykes 1980) and the Baltic Shield (Calcagnile & Panza 1978) model. The Arabian Sea area studied by Singh (1988) is a continent–ocean transition zone, with the oceanic part of similar age to the Norwegian Sea area studied here, and Singh’s models were derived using surface wave dispersion data. Crustal thicknesses beneath the Arabian Sea vary between 16 and 28 km, similar to values obtained in this study. Singh (1988) concluded that thermal blanketing effects played a major role in generating the abnormally high thicknesses. Thus, considering the tectonic similarities of the two environments, our crustal thicknesses are reasonable, despite the poor resolution of the upper part of the lithosphere in our study.

The results from this study were also compared with a 45 Myr old oceanic crust model (Evans & Sykes 1980) and with the Baltic Shield structure as observed by Calcagnile & Panza (1978). As expected, the average lithospheric depth is smaller than the Shield value. Kanamori & Press (1970) found an ocean basin lithospheric thickness value of 70 km, which...
Figure 8. Comparison of the shear velocity models estimated in the present study using initial model 1 (Thick black lines, (a) W; (b) W2; (c) W3; (d) W4) with velocity models of the Arabian Sea (thin broken line, Singh 1988), 40 Myr old oceanic crust (thin continuous line, Evans & Sykes 1980), and the Baltic Shield (thick broken line, Calcagnile & Panza 1978).

compares well to the value of 65 km obtained by Evans & Sacks (1980) for oceanic structure of age 45 Myr. Compared with these, our value of 60 km is low.

The problem of crust and upper mantle elastic anisotropy is not addressed in this study. However, the large standard deviations shown in Fig. 5 point to the existence of azimuthal anisotropy. We investigated the presence of anisotropy by performing a simple test for polarization anisotropy. Love and Rayleigh wave group velocities were inverted separately for each of the regions W, W2, W3 and W4. The results are shown in Fig. 9. Rayleigh and Love wave models for region W3 are not very different from the model obtained after the joint inversion of Love and Rayleigh wave group velocities. The Love wave model has larger velocities at a depth of about 30 km below the surface. There is a clear mismatch between the models in the other three regions. This difference in the models points to the existence of polarization anisotropy in the whole region. Having attributed the difference to anisotropy, the anisotropy coefficient $\gamma$ for all regions is estimated by

$$\gamma = \frac{SH - SV}{S_{\text{mean}}}.$$  

The shallow layer coefficients are 1–5 per cent for all regions except region W2, where the value is extremely large (>10 per cent). The coefficients between the crust and lithosphere lid are also large (>7 per cent).

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Figure 9. Comparison of the shear wave velocity models obtained after separate inversions of the Rayleigh (thin black line) and Love wave (broken line) group velocities. The thick black line is the model obtained after inverting Rayleigh and Love wave group velocities simultaneously.

CONCLUSIONS

We have presented the results of a study of Rayleigh and Love wave dispersion across the Norwegian–Greenland sea region. The models obtained for the region covered by the paths making up group W (between Svalbard and Iceland) are similar to those of typical young oceanic-type structures; however, the observed lithospheric thickness of about 60 km is greater than expected for young oceanic-type structure. This can be attributed to the contribution of the Iceland plateau velocity structure to the final average structure. The low crustal thickness value of 9 km is reasonable for this area and is probably due to the influence of the young oceanic region near the ridge segments crossed by the group W paths. The continent–ocean transition covered by the paths making up groups W2, W3 and W4 has been shown to have a crustal thickness of between 18 and 20 km, which compares well with previous geophysical studies in the region. However, lower crustal and upper mantle shear wave velocities are much lower than expected, possibly reflecting high heat flow values extending to depths of about 30 km beneath the surface. Average lithospheric thicknesses of 60 km are observed. Evans & Sacks (1980) found that ocean of crustal age 45 Myr has a lithospheric thickness of about 65 km. Thus our value is lower than expected, especially considering that the region under study is a continent–ocean transition zone. It is possible that the thick sedimentary layers in the region have reduced the normal cooling rate of the lithosphere, resulting in lower shear velocities.
velocities than normal. Wang & Nur (1992) observed that the shear velocity decreases significantly with increasing temperature.

Although resolution at shallow depths is poor, crustal thicknesses and velocities are reasonable enough to give confidence in the results. The anomalies observed such as the low shear wave velocities in the lower crust and upper part of the upper mantle, as well as the 'thin' lithospheric thickness of about 60 km, point to the need for more comprehensive studies of the structure. One way forward would be to address some of the weaknesses observed in this study. Nearly all the paths cut across seafloor spreading anomalies (Eldholm & Sundvor 1980), and so the velocity structures determined are average structures across the region. Determining pure-path velocities, as shown by Forsyth (1975), would help in obtaining individual velocities of the various tectonic units crossed by the paths.

Since it has been shown that anisotropy exists in this region, future studies should account for its effects on the velocities of the tectonic units, especially the normal oceanic lithosphere. To improve and increase the reliability of the results, tomographic methods to estimate velocity structure whilst accounting for anisotropy (Montagner & Tanimoto 1990; Yanovskaya et al. 1998, Lévêque et al. 1998), should be employed.

The results obtained in this study provide important information to be utilized in future studies on the inversion for shear wave velocity structure of the Norwegian-Greenland sea region.

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