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Crustal structure of a super-slow spreading centre:a seismic refraction study of Mohns Ridge, 72° N

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Abstract: A series of eight high-resolution seismic refraction profiles from the ultra-slow spreading (16 mm yr^{-1} full spreading rate) Mohns Ridge in the Norwegian–Greenland Sea has been treated with modern inversion methods. The profiles were shot parallel to the ridge at an off-axis distance of 0–135 km corresponding to crustal ages of 0–22 Ma. The resulting models are constrained by synthetic seismograms and gravity modelling.

The crustal thickness in all profiles is well below the global average for typical oceanic crust, and shows a high variability with a mean thickness of 4.0 ± 0.5 km. This is mainly due to a very thin and variable lower crustal layer (Layer 3). Generally, the crust is thicker beneath basement highs and thinner beneath basins, implying local isostatic compensation. The top of the basement (Layer 2a) consists of a zone with low P-wave velocities (2.5–3.0 km s^{-1}). The mean thickness of this layer decreases with distance from the ridge. Beneath it lies a layer with slightly higher velocities (Layer 2b). Its thickness shows less variability along a given profile and an overall increase with age. The combined average thickness of the upper two layers remains nearly constant, indicating that the boundary between Layer 2a and 2b may represent an alteration front.

Upper mantle velocities are generally slow, around 7.5 km s^{-1}. For the profile directly within the rift valley, a model without a third layer, incorporating a constant gradient up to upper mantle velocities, and a model with a Moho depth inferred from neighbouring profiles and upper mantle velocity as slow as 7.2 km s^{-1} fit the seismic and gravity data equally well. The crustal structure is not mature below the ridge. These observations support previous models suggesting the presence of low densities and velocities at about 2 km below the rift axis. Poisson's ratios determined from converted S-wave modelling are incompatible with a Layer 3 consisting of purely serpentinized peridotite. However, a volume fraction of 10–40 per cent serpentinite cannot be ruled out.

Keywords: seismic, refraction, Greenland Sea, crustal, structure

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1   Introduction

Early refraction studies in the 1960's first demonstrated that mature oceanic crust has a rather uniform thickness (ca. 6.5 km) and is composed of three layers (Raitt, 1963), while subsequent studies in the 1970's subdivided these layers (Houtz & Ewing, 1976) and demonstrated a gradient in seismic velocities within each of these layers (Kennett & Orcutt, 1976; Whitmarsh, 1978; Spudich & Orcutt, 1980). Significant advances in the understanding of oceanic crust formation have been made since then, one of the most important being the dependence of axial morphology and ridge segmentation on spreading rate (Tapponier & Francheteau, 1978; Chen & Morgan, 1990; Lin & Phipps Morgan, 1992; Phipps Morgan & Chen, 1993; Schouten, Klitgord & Whitehead, 1985). Most detailed seismic studies however have been conducted on crust formed at spreading rates spanning the range from the superfast values of the southern East Pacific Rise (Bazin et al., 1998; Grevemeyer, Weigel & Jennrich, 1998; Forsyth & the MELT Seismic Team, 1998) down to the slower rates of the Mid-Atlantic Ridge (Fowler, 1976; Detrick et al., 1990; Tolstoy, Harding & Orcutt, 1993) and very few experiments have examined the crust formed at ultra-slow spreading rates (Muller et al., 1997).

Studies of super slow spreading centres, with half-spreading rates less than 10 mm/y, are critical for understanding ocean crust formation processes, since they include major portions of the mid-ocean ridge (MOR) system, such as the > 6500 km long South West Indian Ridge. Seismic refraction data and geochemical analyses of rocks acquired on this ridge indicate, that the crust is anomalously thin (4 ± 1 km) and the lower crust might contain up to 35 ± 10 vol% serpentine (Minshull & White, 1996; Muller et al., 1997). Furthermore, super slow spreading centres are worth studying since they represent end member structures.

In order to study crustal accretion at an oblique super slow spreading centre, a two-cruise geophysical survey was conducted at the Mohns Ridge, in the Norwegian-Greenland Sea (Figure 1). Gravimetric, magnetic and swath bathymetry data were gathered in addition to single channel seismic reflection and refraction data, for the purpose of constraining the crustal structure and its evolution. From these data, the general characteristics of the ridge, and the scale of the along-axis segmentation were determined (Géli, 1993; Géli, Renard & Rommevaux, 1994). The Mohns Ridge is characterized by a segmentation at the scale of 30 - 50 km and both the axial graben as well as the bounding rift flanks are marked by a series of oblique *en-echelon* ridges or basement highs with intervening valleys. These *en-echelon* ridges are oriented perpendicular to the 120° spreading direction indicated by the NUVEL-1 global plate motion model (De Mets et al., 1990). Based on these observations and the presence of 030° oriented magnetic anomalies Géli, Renard & Rommevaux, 1994, interpreted these ridges to be volcanic edifices representing en-echelon spreading segments. Alternatively, analog modelling of oblique extension suggests that such oblique ridges can form tectonically and that the Mohns Ridge crust may consist of transtensional fault bounded blocks (Dauteuil & Brun, 1993; Dauteuil & Brun, 1996).

2   Structure of oceanic crust

Although it is now widely accepted that oceanic crust is comprised of layers with different velocity gradients, rather than the constant velocity layers first suggested by Raitt, 1963, we will use the terms Layer 2a, 2b and 3 in the discussion below (White, McKenzie & O'Nions, 1992). In our modelling velocity steps have been included between layers. In reality this might be a very steep velocity gradient, which cannot be resolved by the data. There is, for example, strong evidence that directly above the Moho a steep velocity gradient can be found (Anderson, 1989).

Layer 2, the uppermost part of the oceanic crust, is comprised of lava flows and pillows, and part or all of the underlying sheeted dike complex. Seismically, it shows a steep velocity gradient with, on average, velocity increasing from 2.5 - 6.6 kms^{-1} and an average thickness of 2.11 ± 0.55 km (White, McKenzie & O'Nions, 1992). Talwani, Windish & Langseth, 1971 were the first to report a low velocity Layer 2a on the Reykjanes Ridge. Numerous geophysical studies confirm that Layer 2a exhibits a strong lateral increase in seismic velocity away from the ridge, going from < 3 kms^{-1} at the ridge axis to > 5 kms^{-1} for 10 Ma oceanic crust (Houtz & Ewing, 1976; Purdy, 1987; Rohr, 1985). Carlson (1998) and Grevemeyer & Weigel (1996) both show that the

seismic velocity of Layer 2a increases mainly during the first 10 Ma, as hydrothermal circulation ceases, cracks close and overall porosity decreases (Fisher, 1998 and references therein), and only slightly thereafter. Layer 2b typically has a seismic velocity of 5.07 ± 0.4 kms^{-1} (Houtz & Ewing, 1976).

The boundary between Layer 2 and Layer 3 is often interpreted as the transition from sheeted dikes to gabbros. While this model may hold true for fast and super-fast spreading (e.g. East Pacific Rise), comparison of ODP drilling results to seismic experiments for the moderate spreading Nazca-Cocos Ridge seem to place the transition within the sheeted dike complex and associate it with changes in the crustal porosity (Detrick et al., 1994). Seismically, Layer 3 is characterized by a higher seismic velocity (6.69 ± 0.26 kms^{-1}) and a smaller vertical velocity gradient (< 1 s^{-1}) than Layer 2 of the oceanic crust (White, McKenzie & O'Nions, 1992). For most parts of the oceanic crust, Layer 3 is believed to be composed mainly of gabbros, which are also found in typical ophiolite sections (Christensen & Smewing, 1981). However on slow spreading ridges, extension of the lithosphere and lack of melt might lead to emplacement of mantle rocks within the gabbroic lower crust (Karson et al., 1987; Cannat, 1993).

Normal upper mantle velocities range from 7.9-8.3 kms^{-1} (White, McKenzie & O'Nions, 1992). Various experiments on the slow spreading Mid-Atlantic Ridge have yielded reduced velocities from 7.2-7.6 kms^{-1} at the ridge crest (Fowler, 1976), and experiments at the Reykjanes Ridge revealed upper mantle velocities from 7.1 kms^{-1} (Bunch & Kennett, 1980) to 7.8 kms^{-1} (Smallwood & White, 1998).

3   Data acquisition and modelling

In July 1988, the R/V Jean Charcot shot a series of 8 refraction seismic profiles using ocean bottom hydrophones (OBHs) and four 9-litre Bolt airguns producing a source frequency centred at 10 Hz. The profiles are between 55-70 km in length and arrayed parallel to the ridge axis. The shot interval of 40s translates to a trace spacing of 100 m. To compensate for the loss in penetration due to backscattering from the exposed rough basement near the ridge, where no sedimentary cover exists, Profile 10 was shot using a larger airgun array of six 16.4-litre airguns operating at lower frequencies (5-10 Hz) and thus a longer shot interval (120s). On each line except Profile 10, three OBHs were deployed, located in the middle and ca. 9 km from both ends of the profile. All OBH data were digitized using a sampling rate of 6 ms and bandpass filtered. When necessary, they were corrected for the OBH offset from the profile determined from the navigation file and for the inline OBH offset, which was determined by fitting the water primary and first multiple arrivals. The largest offset from the profile was found for OBS 1 on Profile 2 (850m), and inline offsets were generally < 200-300m. A time shift of 80 ms, probably due to clock errors, was found in the arrivals of all OBHs and corrected.

The crustal structure of all refraction profiles was obtained using the joint inversion and raytracing algorithm of Zelt & Smith (1992). The profiles were modelled downwards starting with the sediments. Upper layers, where not constrained from the arrivals from this layer, were adjusted to improve the fit of lower layers. We modelled

first the profile furthest away from the ridge, because of its better quality. Going toward the ridge, we used the adjacent profile as a starting model, adjusting the velocities and depths only when necessary. The thicknesses and velocities of the different layers were confirmed by synthetic seismogram computations using the software of Zelt & Ellis (1988). The code is based on assymptotic ray theory (Cerveny, Molotkov & Psencik, 1977). The bathymetry, the sedimentary layers and the basement topography were digitized from the single channel reflection profiles, except for Profile 10, directly along the ridge, where no reflection data were available. For this particular line, the sediment thickness is negligible and the bathymetry was derived from the SeaBeam data. To demonstrate the quality of the single channel reflection profiles Figure 3 shows the single channel reflection data of Profile 3. The horizontal spacing between the nodes at which the depth of a boundary was defined was on average 1 km, and closer when necessary for the basement, because of its steep topography. For deeper horizons it was increased to 5 km. In our models the water velocity is 1460 m/s in the upper 300 m and increases continously to 1490 m/s at the seafloor. These velocities are consistent with water temperatures between (1 and 4°C) found in the Norwegian-Greenland Sea (Dietrich et al., 1975). The water depths in the refraction profiles match the sonar depths closely (within 1%). The outermost four profiles (2, 3, 4 and 6) with ages between 7.9 and 22.4 Ma have a 0.5 -1.0 km thick sediment cover. The sediments are divided into two layers with near water velocities at the ocean bottom increasing up to 2.0 kms^{-1} directly above the basement. In these profiles sediment velocities were determined from the modelling. Profiles 7, 8, 9 and 10 with ages between 0 and 4.4 Ma show little or no sediment cover and only few refracted arrivals from the sediments can be found in the sections. For the few sedimentary basins we used similar velocities as for the upper sedimentary layers of the other profiles.

The picking error in all profiles for the P-wave arrivals was not larger than ± 50 ms, except for Profile 10, where the picking error was ± 60 ms. The multiple arrivals are generally earlier than predicted from the model, probably due to an out of plane raypath. Both the multiples and the converted S-arrivals were assigned a larger error of 90 ms. The multiple arrivals on Profile 10 show the largest misfit (150 ms), because of the steep topography in the rift valley. The fit between the model and the travel time picks is given by the normalized χ^2 function for each profile and phase respectively and should not exceed 1.0 for any arrival. Perturbation of the velocity and the depth of a single horizon allows the assessment of the velocity and depth uncertainties. Figure 4 shows the variation of the χ^2 parameter for all relevant phases for a velocity perturbation and a depth perturbation of Layer 2a in Profile 8 and of the depth of the Moho and velocity of Layer 3 in Profile 7, which are representative for all profiles. Using the $\chi^2 = 1$ contour, a reasonable estimate of the depth uncertainty of the Moho is ± 1 km. The number of picks, the picking error, the values for the χ^2 parameter and the rms-misfit are given in Table 1 for each profile and phase, respectively.

4 Results

All models derived from the inversion of the refraction data are shown in Figure 5. The isovelocity contours (0.5 kms^{-1} spacing) are included as well as the layer boundaries used during the inversion. For Profile 10 directly on the ridge, two models fit the data equally well, one incorporating a change in gradient at a depth where the Moho might be expected from the neighboring profiles (model A) and a second model with a gradual increase of velocity from the base of Layer 2b to the bottom of the model (model B). Crustal velocity-depth functions averaged for velocity gradient and layer thickness from the refraction modelling are shown in Figure 6 a, b. These are superimposed on the velocity-depth relationships from White, McKenzie & O'Nions, 1992 for Atlantic crust of ages from 0-127 Ma and 0-7 Ma, respectively.

The overall crustal thickness in the study area is 4.0 ± 0.5 km and thus well below the global average for normal oceanic crust. The crustal thickness shows a high degree of variability, a consequence of the abundance of oblique ridges and intervening troughs. The velocities and thicknesses of the Layers 2a, 2b, 3 and their changes with crustal age will be described in greater detail in the following subsections.

4.1 Upper crust

First arrivals from Layer 2a were only observed on two profiles (8 and 9), for all other profiles its velocity was determined from delays of other arrivals. At the top of Layer 2a we found unusually low velocities atypical of basalt. In all profiles, Layer 2a exhibits a high degree of variability with appreciable thickening at topographic highs (see Figure 5). The velocity gradient in this layer is high with 1.63 ± 0.29 s^{-1}. Profile 10 and Profile 9 directly above the ridge both display a thick continuous Layer 2a with velocities ranging from 2.5-3.0 kms^{-1} (see Figure 7), which is lower than the velocity of Atlantic crust of the same age (for comparison see Figure 6). This layer velocity increases away from the ridge as sediment cover thickens, reaching 3.3-3.5 kms^{-1} at the furthest profile. The average thickness of this layer drops from 0.45 to 0.2 km in the same direction (Figure 8). In Profiles 2, 3 and 4 furthest away from the ridge it is not found continuously, and often disappears beneath local basins. As also found in previous refraction experiments, the main velocity increase occurs during the first 10 Ma after formation of the crust (Carlson, 1998; Grevemeyer & Weigel, 1996).

The structure of Layer 2b is more uniform and the variability of the thickness is lower than that of Layer 2a along each profile as well as between different profiles. The velocity of this layer also increases slightly away from the ridge. The velocity is 3.5-4.5 kms^{-1} on the ridge axis and reaches an average of 4.5kms^{-1} for older crust. As in Layer 2a, the main increase can be found for the first 10 Ma, and only slight changes afterwards. The velocity gradient in this layer is 0.74 ± 0.1 s^{-1}, less than the velocity gradient in Layer 2a. The combined mean layer thickness along each given profile of Layer 2a and 2b remains nearly constant (see Figure 8).

4.2 Lower crust and mantle

The average thickness of Layer 3 is 2.45 $\pm$ 0.50 km, which is extremely thin compared to typical oceanic crust, where Layer 3 has a thickness of 4.97 $\pm$ 0.90 km (White, McKenzie & O'Nions, 1992). The thickness of Layer 3 remains nearly constant for all profiles, except for model (A) of Profile 10, which has an even thinner Layer 3 (2.0 km). The variability within each profile is larger than for Layer 2b, mainly due to thickening crust beneath topographic highs and thinning crust beneath basins. The velocity displays almost no increase from 5.8-6.8 kms^{-1} at the ridge to 6.2-6.8 kms^{-1} for Profile 2 furthest away from the ridge (22.4 Ma). The average velocity gradient for all profiles is 0.38 s^{-1} and therefore substantially lower than in the overlying units, as has been found in other regions (Mutter & Mutter, 1993). On all profiles of our study, the upper mantle velocity is anomalously low around 7.6 kms^{-1}, with the lowest values of 7.0 and 7.4 kms^{-1} in Profile 10 directly on the ridge (see Figure 7).

4.3 Synthetic modelling

We constrained the thickness of the crust for all profiles by synthetic seismogram modelling using software developed by Zelt & Ellis (1988). Figures 9 and 10 show the results of the amplitude modelling for, respectively, OBH 2 located in the middle of Profile 2 and OBH 1 in the middle of Profile 6. Included in the figures are the ray paths for that part of the model covered by the data from the OBH and the data section. The synthetic seismograms are spaced at a 200 m interval and the same offset-dependent gain has been applied as to the seismic data. The wavelet used to compute the synthetic seismograms has been extracted from the original data. The onset of the P_n refraction can clearly be seen and constrains the crustal thickness estimates. Part of the P_mP reflection is also visible and has been used in the modelling. The parts of a raypath which the rays traversed as S-waves are shown dashed.

Figure 11 shows the synthetic seismograms, raypath and data for an example from line 10. The lower frequencies recorded do not resolve fine details as effectively as in the other profiles. On the other hand, it enabled deeper structures beneath the ridge to be modelled successfully.

4.4 Gravity modelling

A further constraint on the refraction models is provided by two dimensional gravity modelling. Figure 12 shows the results of the gravity calculations along the profiles. The modelling was performed with the 2-D gravity inversion program Saki (Webring, 1980) and with the gravity module of the software of Zelt and Smith (1992). Density for the water layer is assumed to be 1000 $\mathrm{kg/m}^3$, the density of the other layers was calculated from the P-wave velocity after Ludwig, Nafe & Drake, 1970. For the Profiles 2, 3, and 4 we used the gravity data from Sandwell & Smith (1995), which covers latitudes up to 72°N. Generally the results of the model fit the data reasonably well, with local isostatic compensation by crustal thickening beneath topographic highs. An exception is Profile 2, where a Moho subparallel to the seafloor topography provides

the best fit to the seismic and gravity. This might be the result of a fracture in the crust, however it is difficult to trace this fault to the MOR or to identify it in the single channel reflection data.

For Profiles 6 to 10, extending to latitudes higher than 72°N, satellite-derived gravity data from Anderson and Knudsen, 1995 were used. The overall fit is less satisfactory than in the southern profiles. One reason is clearly 3-D effects, since the topography directly at the ridge is rough. The unsatisfactory fit on the eastern side of Profile 6 is probably due to edge effects. Profile 7 clearly cuts a ridge in an oblique angle leading to a calculated anomaly of the right size but its maximum shifted to the east. It was not possible to fit the gravity data on Profile 10 using two dimensional models, due to the steep topography in the rift valley.

Therefore, a 3-D gravity model was calculated in the area covered by the SeaBeam data using the method of Parker (1972). The effect of the seafloor topography down to 3500 m depth was subtracted from the free air anomaly using a constant density of 2700 kgm^{-3} for this layer. Both models of Profile 10 fit the observed gravity equally well, it is not possible to determine which is more realistic. The fit obtained in both cases is better than that of the 2-D modelling. This implies that some of the misfits on the off-axis profiles may be caused by 3-D effects. However these calculations cannot be extended to the rest of the profiles, because of the lack of high resolution bathymetric data.

To verify if the crustal thickening beneath the highs is sufficient to compensate those highs, we calculated the load anomaly for all profiles. Figure 13 shows the resulting anomaly for Profile 3. The out-of-balance stresses do not exceed 50 bar, which can be sustained by the crust (Whitmarsh et al., 1996).

4.5   S-wave velocity

The outermost Profiles 2, 3 and 4 (22.4 to 15.4 Ma) show a thick sedimentary layer and clear S-wave arrivals. In the middle Profiles 6,7 and 8, S-wave arrivals can be found, but they are less energetic. No S-wave arrivals could be detected in the two on-axis Profiles 9 and 10. One reason for the absence of S-wave arrivals in these profiles might be sidescattering from steep walls in the rift valley, as shown for a reflection profile at the Mohns ridge by Eiken (1991). Most S-wave arrivals result from diving rays into Layer 2b and Layer 3. In some profiles S-wave arrivals from diving rays into the mantle were also found. Observable S-wave arrivals from Layer 2a are scarce, mainly because they interfere with refracted arrivals from the sediments. Figure 14 shows OBH 1 located in the middle of line 3. The reduction velocity of 4 kms^{-1} gives a good image of the S-wave arrivals. The raypath is shown as a solid line for the part that the wave travels as a P-wave.

Water-saturated sediments exhibit a Poisson's ratio close to water, between 0.49 at the seabottom and 0.41 at a depth of 1 km (Hamilton, 1979). The conversion from P- to S-wave takes place at the interface between sediments and basement (Vidmar, 1980). Poisson's ratio structure of the upper oceanic crust can be determined from ophiolites, boreholes or by modelling of seismic data (Collier & Singh, 1998 and references therein). Because of the paucity of converted arrivals from Layer 2a, we assigned a Poisson's

ratio of 0.27 to this layer. The Poisson's ratio of this layer doesn't influence our results substantially, since the profiles exhibiting the best S-wave arrivals show a thin and discontinuous Layer 2a. Through forward modelling of the S-wave arrival times, the Poisson's ratio could be determined for Layer 2b, Layer 3 and for the upper mantle. For these layers a Poisson's ratio of 0.28 gave the best results. For comparison, Christensen & Smewing (1981) found a Poisson's ratio of 0.27 for the upper Layer 2 consisting of pillow lavas in the case of the Oman ophiolite. Similar values have been found from seismic data and compared to data from ophiolites by Spudich & Orcutt, 1980. The gabbroic layers have Poisson's ratios between 0.28 and 0.30 with one exception of 0.34 for a gabbro. Analyses of the upper mantle peridotite resulted in Poisson's ratios of 0.28 to 0.33 deeper within the mantle. Our results are consistent with these experimental data.

5 Discussion

5.1 Crustal thickness

While for full spreading rates > 15 mm/a, there is no observed dependence of the crustal thickness on spreading rate, crust formed at ultra slow rates is often thin and highly variable (Chen, 1992; White, McKenzie & O'Nions; Bown & White, 1994). In the study area on Mohns Ridge we found an average crustal thickness of 4.0 ± 0.5 km, well below the global average for oceanic crust. The crustal thickness along a given profile displays a higher degree of variability than is typical for oceanic crust. Figure 15 shows a compilation of crustal thicknesses at various ultra slow spreading centres from Bown & White, 1994 compared to their numerical melting model. Since the model is two dimensional we take the spreading rate perpendicular to the ridge axis (14 mm/a). The crustal thickness of Mohns Ridge falls within the range of predicted values, for a mantle potential temperature of ca. 1280° C.

5.2 Structure of the upper crust

The averaged thickness of the upper crust (Layer 2) for all profiles it is 1.57 ± 0.16 km, which is only 0.55 km less than average crust (White, McKenzie & O'Nions, 1992). The modelling shows, that Layer 2a exhibits an overall thinning off axis from values of ca. 500m at the axis to <200 m for 10 Ma old oceanic crust. This trend is anti-correlated with the thickness of Layer 2b, which shows a mean thickness of 1.3-1.4 km at the ridge axis, increasing to 1.6 km for 10 Ma crust. These observations tend to confirm the prevailing model that Layer 2a is a temporal phenomenon (Houtz & Ewing, 1976; Grevemeyer & Weigel, 1996; Carlson, 1998) and that once hydrothermal circulation ceases and pore space closes, this layer in a sense disappears. The seismic velocities in Layer 2a show an increase from 2.5 kms^{-1} at the ridge axis to values > 3.0 kms^{-1} beyond 20 km from the ridge. Similar velocities have been found on Reykjanes Ridge (Smallwood & White, 1998) and on the Mid-Atlantic Ridge (Purdy, 1987). Since the ray coverage for the profiles further away from the ridge is not high, this increase might not be statistically significant. It would however be in agreement with models

indicating that changes in the physical properties in the oceanic crust occur in the first few Ma directly at the ridge and that little to no variation occurs thereafter. The fact that modelling of Layer 2a requires considerable thickening at topographic highs might also indicate that the thickness of this seismic layer for young crust is locally linked to the volcanic activity and therefore the amount of extrusives. Through modelling of the magnetic anomalies and comparison to seismic reflection and refraction data Tivey & Johnson (1993) proposed that in young crust the base of Layer 2a represents the extrusive to dike transition, while in older crust it corresponds to an alteration front. Both models are consistent with our data and the question might be resolved for Mohns Ridge through modelling of the magnetic data.

5.3   Structure of the lower crust and mantle

Layer 3 shows an average thickness of 2.45 ± 0.5 km, which is 2.53 km less than that of average oceanic crust. Thus Layer 3 is mainly responsible for the thin crust we observe at Mohns Ridge, while Layer 2 shows near-normal thicknesses. This is in good agreement with the work of Mutter & Mutter, 1993, who found that the paleodepth to the Layer2/3 boundary in zero-age restored crust is relatively invariant throughout the worlds oceans, and conclude that variations in thickness of Layer 3 control the oceanic crustal thicknesses. Isostatic compensation of the crust at Mohns Ridge is mainly due to variations in the thickness of Layer 3. Most basement highs correspond to local thickening of Layer 3, while troughs correspond to local crustal thinning. The only exception was found in Profile 2, where a crust of nearly constant thickness fits the seismic and gravity data best. This might be the result of a fault in the crust, however it is difficult to trace it to the MOR or to identify it in the single channel reflection data.

Attempts have been made to correlate refraction velocities to the lithology of the lower crust and upper mantle using laboratory data. However, P-wave velocities as low as 5.8 kms^{-1} and as high as 6.8 kms^{-1} (comparable to the Layer 3 velocities reported here) can correspond to either gabbros or serpentinized peridotites (Christensen, 1966; Christensen & Smewing, 1981; Horen, Zamora & Dubuisson, 1996). Based on laboratory measurements, Christensen (1966) found a dependency between the Poisson's ratio and the degree of serpentinization in peridotites. Gabbroic rocks from ODP Leg 153 on the Mid-Atlantic Ridge south of the Kane Fracture Zone show P-wave velocities between 6.5 and 7.4 kms^{-1} and a Poisson's ratio between 0.25 and 0.28, while serpentinized peridotites from the same Leg have a higher Poisson's ratio (0.30 to 0.40) at lower velocities between 4.5 and 6.0 kms^{-1} (Miller & Christensen, 1997). On the other hand, a compilation of data for different oceanic rocks Horen, Zamora & Dubuisson (1996) shows that for P-wave velocities of 6.1 - 7.5 kms^{-1}, pure gabbros and peridotites with 10 - 40 % serpentinite, have similar Poisson's ratios, between 0.25 and 0.35. Since this is the velocity range found in our study it is not possible to distinguish between 100 % gabbro or a mixture of peridotite and 10 to 40 % serpentinite. The most one can definitively say is that a Layer 3 consisting purely of serpentinized peridotite as proposed by Hess (1955) is not consistent with the Poisson's ratio found from S-wave and P-wave modelling in this work.

In our study, the average upper mantle velocity (ca. 7.5 kms^{-1}) is anomalously slow. Similarly low velocities are typically only encountered beneath very young oceanic crust at slow spreading centres (Fowler, 1976; Bunch & Kennett, 1980; Smallwood & White, 1998). One possible explanation might be, that fracturing of the crust, which is more dominant at slow and ultra-slow spreading rates (Morris et al., 1993), opens pathways for hydrothermal fluids into the upper mantle (Francis, 1981; White, McKenzie & O'Nions, 1992). The serpentinization resulting from the circulation of fluids in the upper mantle would thus be responsible for the anomalous low seismic velocities (Horen, Zamora & Dubuisson 1996; Miller & Christensen, 1997). A different explanation for the anomalously low velocities for ridge parallel profiles might be upper mantle anisotropy, caused by the preferential alignment of olivine crystals in the flow direction away from the ridge and which can cause around 5% difference between the velocities measured parallel and perpendicular to the ridge (Anderson, 1989).

5.4 Structure beneath the spreading axis

Beneath the centre of the rift valley (Profile 10), two different velocity models fit the data equally well: below Layer 2, there may be either a distinct Layer 3 overlying an upper mantle characterized by a seismic velocity of about 7.2 kms^{-1} (model A), or a continuous velocity structure characterized by a smooth velocity gradient (model B). However, only 15 km off axis, Profile 9 (at the edge of the rift valley) reveals a seismic structure similar to all other off-axis profiles, which systematically exhibit an upper mantle velocity of 7.5 - 7.7 kms^{-1} and a well defined Layer 3. This result recalls early observations from the Mid-Atlantic Ridge (Whitmarsh, 1975; Fowler, 1978; Fowler and Keen, 1979) indicating that "normal" seismic structure already exists as close as 10 - 20 km away from the spreading axis. Beneath the spreading centre, the final seismic structure of the ocean crust is not yet mature: Layer 3 may not yet be completely formed (model B), and/or (the two hypotheses are not incompatible) slow velocities are present within the upper mantle and at the base of the crust (model A).

In both models, the structure is different, on and off-axis. In model A, low densities may be present at the base of a crustal layer that is not yet fossilized below the spreading centre; in model B, the upper mantle velocity is slower, thus the density is expected to be lower below the spreading axis. Whatever the model, the across-axis median valley relief is not isostatically compensated, but supported by stresses within the lithosphere or asthenophere, as shown elsewhere by numerous previous studies at the crest of the Mid-Atlantic Ridge (e.g. Cochran, 1979; Watts, 1982; Prince & Forsyth, 1988; Dalloubeix, Fleitout & Diament, 1988). At the Mohns Ridge, moreover, the density variations on and off axis that can be inferred from our seismic data increase the departure from isostatic compensation instead of compensating for the observed seafloor relief.

In order to explain the so-called "paradox of the axial profile", Neumann and Forsyth (1993) have proposed that the amplitude of the dynamic mechanism creating the median valley topography is correlated with the thickness of the relatively strong upper mantle layer located above the 750°C isotherm. The extreme lack of compensation of the axial relief that we infer here for the Mohns Ridge suggests that the stiff

upper mantle layer is particularly thick at ultra-slow spreading centres: as explained by Neumann and Forsyth (1993), thinning the crust deepens the 750°C isotherm and thickens the strong layer in the lithospheric mantle. Therefore, the anomalously slow upper mantle velocities (between 7.2 to 7.6 kms^{-1}) that we find in all profiles at Mohns Ridge cannot be due to higher upper mantle temperature, compared to the Mid-Atlantic Ridge. Possible alternative causes may be a higher degree of fracturing and/or serpentinization within the crust and the upper mantle near the axial zone, resulting in differences in the depth of the brittle-ductile transition (Dauteuil and Brun, 1996).

Using Bouguer anomalies, obtained by subtracting from the measured free air gravity anomalies the effect of seafloor topography referred to an arbitrary flat datum level located 3.5 km below sea level, Géli, Renard & Rommevaux (1994) indicated the possible presence of a three dimensional body with densities lower than the surrounding off-axis rocks lying at about 2 km below the surface, as well as probable local effects associated with the presence of eruptive centres. This interpretation appears to be consistent with the presence of an anomalous velocity/density structure in the lower crust near the spreading axis at the Mohns Ridge (model A) and with other observations at the ultra-slow spreading South-West Indian Ridge indicating that the magma supply is likely to be controlled by near surface processes such as 3-D melt migration within the crust (Grindlay et al., 1998). However, in the absence of additional information - (e.g. the velocity vs. petrology relationship), it is impossible to discriminate between the two seismic models A and B and ascribe a geological significance to the low density bodies inferred from our interpretation of the gravity signal, after correcting for bathymetry.

6   Conclusions

The present study shows that at the ultra-slow spreading Mohns Ridge the seismic velocity structure of the oceanic crust for ages between 1 and 22 Ma is well described by Layers 2a, 2b and 3 as found elsewhere. However, we verified that the measured mean crustal thickness of about 4 km is approximately 1 to 2 km thinner than the average crust produced at the Mid-Atlantic Ridge where the spreading rate is 30 to 40 % faster. The observed small crustal thickness is due to the 1 to 2 km smaller than average thickness of Layer 3 whereas the thickness of Layer 2 (between 1.5 and 2 km) is similar to typical crust produced at the MAR. The importance of Layer 3 variation in accounting for the observed variations in crustal thickness was previously recognised by Mutter & Mutter (1993) from a compilation of oceanic crustal structure produced in a wide range of environments. If we take into consideration the spreading velocity normal to the ridge instead of the oblique spreading velocity then the anomalous thickness of the crust is well explained by the melt generation model of Bown & White (1994) for a mantle potential temperature of 1280°C (see figure 15). No anomalous mantle temperature is needed to explain the reduced crustal thickness.

Along the profiles that closely follow isochrons, Layer 3 is generally thicker below topographic highs, so that an overall correlation is observed between crustal thickness

and topography at a scale of about 20 to 40 km. This correlation justifies the assertion that isostatic compensation along isochrons at Mohns Ridge is attained by the variable thickness of Layer 3. Locally the gravity modelling showed that some areas remain uncompensated, probably due to 3-D tectonic features that are not properly imaged by the 2-D velocity structure obtained.

The thickness of Layer 2 remains approximately constant but as crustal age increases the thickness of Layer 2b increases at the expense of a thinning layer 2a. This evolution is accompanied by an increase in Layer 2 velocity up to 10 Ma. Layer 3 velocity has a much smaller but similar variation. Layer 2 velocities are also found to be smaller than average for typical oceanic crust produced at the MAR. All these observations support the view that the seismic structure of the upper crust is strongly influenced by hydrothermal processes for young crust but their effect ceases at 10 Ma probably due to pore sealing and reduced thermal gradient. The transition is likely to be associated with an alteration front, the depth of which tends to decrease with age. The local thickening of Layer 2a at topographic highs might be correlated to the thickness of the extrusives.

In most profiles converted S-waves at the sediment/basement interface were observed and studied. The measured Poisson's ratio of 0.28 for Layer 3 does not allow a clear distinction between gabbros and peridotites with a degree of serpentinization between 10 and 40 %, both of which have been proposed as the major component for oceanic Layer 3 generated at slow spreading ridges. Based on the data analysed, we cannot rule out a more complex model where gabbros and serpentinites alternate in the lower crust at a small scale.

Off-axis, the average upper mantle velocity (about equal to 7.5 kms^{-1}) is slower than the average observed for other Mid-Atlantic Ridge spreading centres. If fractures resulting from the mechanical extension of the crust reach the upper mantle, then fluid migration is facilitated and a small average serpentinization of the upper mantle might explain the measured values. Such lower than average values have been also found in other slow spreading centres (Fowler, 1976; Bunch & Kennett, 1980; Smallwood & White, 1998). Below the rift valley axis, the presence of low velocities suggests that either Layer 3 is not completely formed (and the crustal structure, not completely "imprinted") below the rift axis, or low velocities are present within the upper mantle and at the base of the crust.

The changes in density structure on and off axis that can be inferred from our seismic data *tend to increase the lack of isostatic compensation* instead of compensating for the across-axis median valley relief. The stiff upper mantle layer below the rift axis is thus probably thicker than at the MAR, due to a higher degree of fracturation and serpentinisation within the crust and the upper mantle near the axial zone. We think that this may help explain why the upper mantle velocities are slower at the Mohns Ridge than the Mid-Atlantic Ridge.

7 Acknowledgements

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9 Figure captions

Figure 1: The study area at the ultra-slow spreading Mohns ridge in the Norwegian-Greenland Sea at 72ºN. The box outlines the area shown in Figure 2.

Figure 2: Shaded seafloor relief from SeaBeam data acquired during the cruise (Illumination from the Northwest. Refraction profiles and OBH positions are shown (Black lines and inverted triangles). Contoured in the background is the gravimetric data from Anderson and Knudsen, 1995.

Figure 3: Single channel reflection data of Profile 3.

Figure 4: (a) The variation of the χ^2 parameter for all relevant phases for a velocity depth perturbation of Layer 2a in Profile 8, (b) same for the depth of the Moho and the velocity of Layer 3 in Profile 7. Upper and lower layer velocities were varied keeping the gradient constant. All depth and velocity nodes for a given layer were varied simoultaneously.

Figure 5: Final models of all refraction profiles, including the model boundaries used during inversion (solid lines) and isovelocity contours with an $0.5\,\mathrm{kms}^{-1}$ contour interval (dashed lines).

Figure 6: (a) Velocity-depth relationship averaged for gradients and layer thicknesses of Profiles 2, 3 and 4 representing crust older than 12 Ma. Shaded gray area is from White, McKenzie & O'Nions, 1992 for Atlantic crust between 0-127 Ma. (b) velocity-depth relationship averaged for gradients and layer thicknesses of Profiles 6, 7, 8, 9 and 10 situated closer to the ridge and representing the younger crust. Shaded gray area is from White, McKenzie & O'Nions, 1992 for Atlantic crust between 0-7 Ma.

Figure 7: Velocity of Layer 2a, 2b and Layer 3 and the upper mantle versus basement age for the refraction lines. Dots indicate minimum and maximum velocities.

Figure 8: Crustal thickness and thicknesses of Layer 2a, 2b and (2a + 2b) combined, and Layer 3 versus basement age for the refraction lines. Bars represent the standard deviation calculated from each 2D profile.

Figure 9: (a) Refraction seismic data of OBH 2 in the middle of profile 2. An offset dependent gain and a bandpass filter (6-18 Hz) have been applied. (b) Synthetic seismogram using the same gain and plotting parameters as the data section. (c) Two dimensional model of the refraction data from the same OBH. The upper part shows the raypaths, while in the lower part the observed travel time picks and the calculated travel times are shown.

Figure 10: (a) Same as Figure 9 only for OBH 2 on profile 6

Figure 11: (a) Same as Figure 9 only for OBH 2 the second OBH from the west on Profile 10

Figure 12: Observed (satellite altimetry data) and calculated free air gravity for all profiles. Profiles 2, 3, and 4 were modelled using the data of Sandwell & Smith (1997). Profiles 6, 7, 8, 9, and 10 using the data of Anderson & Knudsen, 1995. Profile 10 has been corrected for 3-D topographic effects. The bad fit for Profiles 6, 7 and 10 is likely due to 3-D effects.

Figure 13: Load anomaly of Profile 3. The out-of-balance stresses do not exceed 50 kPa, which can be sustained by the crust.

Figure 14: (a) Refraction seismic data of OBH 1 in the middle of Profile 3, reduction velocity is 4.0 km/s. (b) Synthetic seismogram using the same gain and plotting parameters as the data section. (c) Two dimensional model of the refraction data

from the same OBH, including the Poisson's ratios. The upper part shows the rays, while in the lower part the observed travel time picks and the calculated travel times are shown. Parts of the raypath on which the wave travels as an S-wave are shown dashed, parts which are travelled as a P-wave are shown as solid line.

Figure 15: Crustal thickness vs. apparent full spreading rate for various ultra-slow spreading ridges. Model curves of the predicted crustal thickness as a function of full spreading rate from Bown & White, 1994, are superimposed. Assumed viscosities: 10^{18} Pa s (dashed lines), 10^{20} Pa s (solid lines), for mantle potential temperatures of 1280, 1300 and 1320° C. [1] Jackson, Reid & Falconer, 1982; [2] Whitmarsh et al, 1993; [3] Osler & Louden, 1992; [4] Horsefield et al, 1994; [5] Pinheiro, Whitmarsh & Miles, 1992; [6] White, Muller & Minshull, 1998.

	Phase	No. of picks	Picking error [s]	Rms misfit [s]	χ^2
Profile 10 mod (a)		988		0.099	0.755
Profile 10 mod (b)		988		0.104	0.824
	Water	326	0.06	0.045	0.568
	Basement refraction	41	0.06	0.042	0.503
	Layer 3	126	0.06	0.064	1.153
	Moho	102	0.06	0.032	0.284
	Mult. base., Moho	121	0.15	0.195	1.705
Profile 9		1025		0.047	0.471
	Water	350	0.05	0.043	0.727
	Basement refraction	53	0.05	0.035	0.509
	Layer 3	84	0.05	0.032	0.421
	Moho	53	0.05	0.038	0.578
	Mult. base., Moho	94	0.10	0.071	0.516
	Converted	12	0.10	0.033	0.116
Profile 8		1186		0.048	0.391
	Water	348	0.05	0.029	0.340
	Basement refraction	101	0.05	0.035	0.500
	Layer 3	139	0.05	0.039	0.616
	Mult. base., Moho	156	0.10	0.095	0.905
	Converted	31	0.10	0.045	0.208
Profile 7		1327		0.056	0.550
	Water	374	0.05	0.046	0.834
	Basement refraction	83	0.05	0.035	0.486
	Layer 3	103	0.05	0.030	0.356
	Moho	93	0.05	0.028	0.316
	Mult. base., Moho	184	0.10	0.093	0.864
	Converted	58	0.10	0.073	0.538
Profile 6		804		0.073	0.891
	Water	126	0.05	0.039	0.604
	Basement refraction	80	0.05	0.039	0.632
	Layer 3	108	0.05	0.047	0.893
	Moho	213	0.05	0.044	0.776
	Mult. base., Moho	129	0.10	0.082	0.684
	Converted	74	0.10	0.160	1.003
Profile 4		1824		0.058	0.479
	Water	536	0.05	0.018	0.126
	Basement refraction	119	0.05	0.030	0.363
	Layer 3	153	0.05	0.046	0.855
	Moho	69	0.05	0.042	0.730
	Mult. base., Moho	179	0.10	0.135	0.774
	Converted	151	0.10	0.073	0.540
Profile 3		1003		0.037	0.243
	Water	220	0.05	0.009	0.032
	Basement refraction	62	0.05	0.033	0.441
	Layer 3	51	0.05	0.047	0.899
	Moho	81	0.05	0.034	0.477
	Mult. base., Moho	105	0.10	0.055	0.305
	Converted	69	0.10	0.056	0.321
Profile 2		1499		0.063	0.612
	Water	370	0.05	0.020	0.158
	Basement refraction	80	0.05	0.023	0.212
	Layer 3	180	0.05	0.048	0.946
	Moho	72	0.05	0.041	0.678
	Mult. base., Moho	197	0.10	0.071	0.505
	Converted	116	0.10	0.172	1.249

Table 1: Number of picks, picking error, rms-misfit and χ^2-parameter for each phase identified on each profile.

Depth (m)

PROFILE 3

TWTT (seconds)

7°47.71 N
04°53.74 E

7°30.06 N
03°18.11 E

5nm

#3

(a)

Depth perturbation [km]

Velocity perturbation [km/s]

(b)

Depth perturbation [km]

Velocity perturbation [km/s]

Distance along the profile [km]

Distance along the profile [km]

S₂

S_n S₃

T-D/8 (s)

T-D/8 (s)

Depth [km]

T-D/8 [s]

σ=0.5
σ=0.27
σ=0.28
σ=0.28

Distance along the profile [km]