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Crust and uppermost mantle structure of the Iceland–Faroes region from Rayleigh wave group velocity dispersion

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Summary. Fundamental mode Rayleigh wave group velocities over the period range 13–85 s have been measured at stations in the British Isles from earthquakes in the Iceland–Jan Meyen region. Published models for the British Isles upper mantle have been used to correct the observations for the continental portion of their propagation path, and thus isolate the group velocity dispersion curve associated with the Iceland–Faroes Ridge area alone. This has been inverted to shear-velocity depth profiles to a depth of *c.* 150 km using a combined Monte-Carlo/Hedgehog technique.

The results show that the Iceland–Faroes Ridge has an anomalous crustal thickness, for oceanic areas, of 20–24 km, supporting the conclusion of other published geophysical studies. Upper mantle structure, which has been determined for the first time, appears less atypical of oceanic regions, with a lithospheric thickness of some 60 km and 'lid' velocities of 4.5–4.55 km s^{-1}.

Introduction

The bathymetry and crustal structure in parts of the North Atlantic Ocean bordering the British Isles are atypical of most ocean basins due to the complex continental breakup and migration of spreading axes which has characterized the evolution of the North Atlantic. This short communication reports shear velocity–depth models for the uppermost 150 km of the Earth on three paths between the Iceland–Jan Meyen region and mainland Scotland (Fig. 1), obtained from inversion of fundamental mode Rayleigh wave dispersion velocities. These models represent the first estimates of upper mantle structure to such depths in this area, and complement the seismic refraction crustal studies of Bott, Browitt & Stacey (1971), Bott *et al.* (1974), and Bott, Nielsen & Sunderland (1976). Surface wave studies, on both local and regional scales, have been carried out for the North Atlantic by Ossing (1964), Bravo & Udias (1974), Searle (1975), Evans & Sacks (1979), Christensen, Kimball & Mauk (1980) and most recently Canas & Mitchell (1981).

Tectonic setting

As well as spanning the Mid-Atlantic Ridge (MAR), Iceland is thought to be the location of at least one 'hotspot' (Morgan 1971; Burke & Wilson 1976), with a lateral expression as an

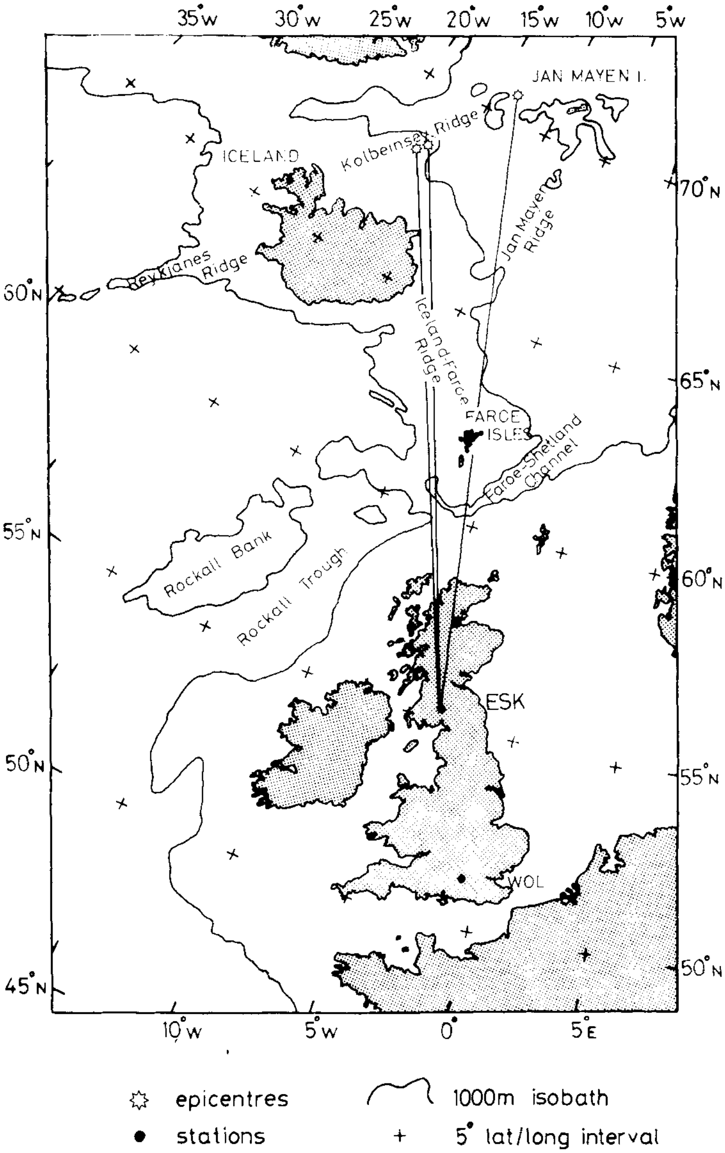

Figure 1. Epicentres, seismograph stations, and physiographic setting of the paths between the Iceland–Jan Meyen region and the British Isles.

aseismic ridge trending SE from Iceland terminating after *c.* 500 km at the Faroe Islands (Fig. 1). Gravity, magnetic and seismic refraction studies (Bott *et al.* 1971; Fleischer 1971; Fleischer *et al.* 1974; Bott & Gunnarsson 1980) have revealed a crustal structure and Moho depth of 27 km which contrast with each of the low velocities of the Icelandic crust and upper mantle (Tryggvason 1962, 1964; Long & Mitchell 1970; Pálmason 1971), typical MAR structures (e.g. Lilwall 1980), and the Faroes Block. The lattermost, from its crustal structure (Bott *et al.* 1974; Bott *et al.* 1976), is thought to be part of a continental micro-fragment system comprising the Rockall Bank, the Faroes Block, and the Jan Meyen Ridge (Scrutton 1972; Roberts, Ardus & Dearnley 1973; Bott *et al.* 1974) as may be seen Fig. 1.

Table 1. USCGS PDE details of the earthquakes utilized in this study. Depths constrained to 33 km during epicentre location are indicated by N. Epicentral distance Δ and azimuth θ are with respect to station ESK (Fig. 1).

Date	Origin time UTC	Latitude (°N)	Longitude (°W)	Depth (km)	m_b	M_s	Δ	θ
1971 April 25	17–48–03.9	68.249	18.139	N	4.9	–	14.70	337.8
1972 March 18	15–00–59.7	68.829	17.331	N	4.9	4.5	15.02	340.0
1974 March 22	19–10–27.6	70.738	14.722	22	5.0	–	16.28	346.3

Data processing and group velocity observations

The earthquakes utilized in this study are listed in Table 1, together with their USCGS PDE details. Source-station group velocities were measured at the WWSSN station ESK (Eskdalemuir, South Scotland) and the MOD(PE) station WOL (Wolverton, SE England). The long-period vertical component seismograms from ESK are presented as Fig. 2. These and the WOL seismograms were digitized at 2 s intervals, either directly from analogue magnetic tape records or manually, in which case the time series were produced using the methods described by Clark & Stuart (1981). These were designed to minimize the systematic and random errors in spectral amplitude and phase introduced by distortion of the seismogram and uncorrected baseline trends. The effects of the seismograph systems were deconvolved from the time series prior to determination of group arrival times using the Multiple-Filter Method (Dziewonski, Bloch & Landisman 1969; Burton & Blamey 1972). No correction was made for source rise time, since for earthquakes of these magnitudes such a term would be insignificant. The maximum likely epicentre mislocations and processing errors (Clark & Stuart 1981) would together incur uncertainties of $\pm 0.075 \text{ km s}^{-1}$ or so in the group velocity estimates.

The stations ESK and WOL lie within 4° of a common great circle path with the three sources. In order to isolate the dispersion characteristics of the Iceland–Faroes Ridge area (IFR) alone, we set up a simplified model of the source-station great circle, consisting of

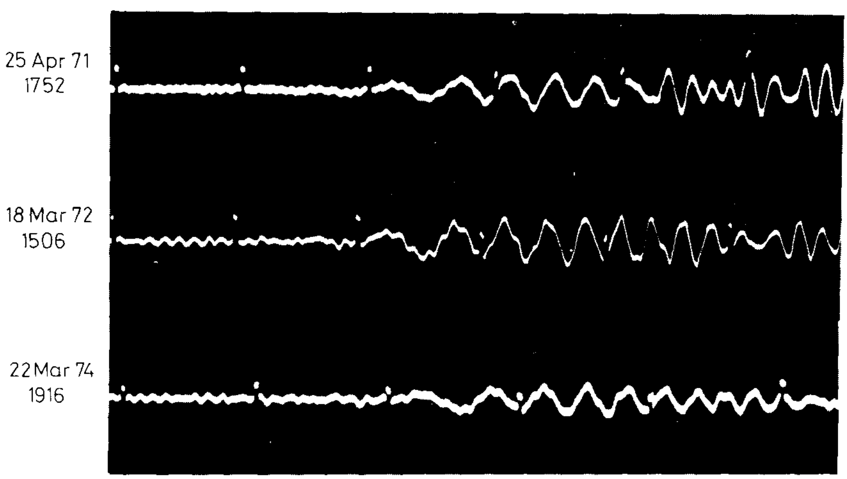

Figure 2. WWSSN LPZ seismograms from the three earthquakes of Table 1 recorded at station ESK. The time of the first minute mark along the record is indicated. The direct Rayleigh wave arrivals are visible by eye at periods up to 60 s or so, and the multiple-filter process has recognized energy up to periods of 75–85 s.

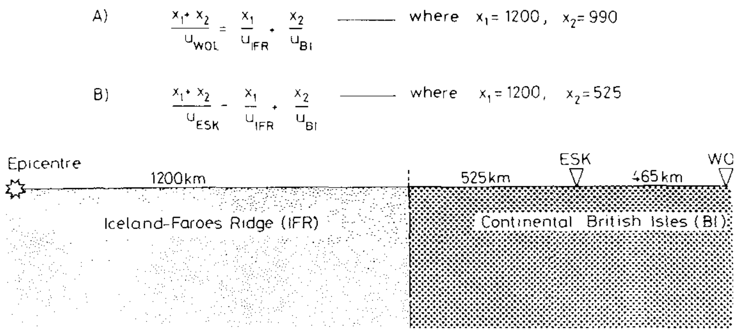

Figure 3. The regionalization scheme adopted to isolate the dispersion characteristics of the Iceland–Faroes Ridge area. Also shown is the distance-weighted average slowness formula of Knopoff (1969), as applied to observations at: (a) ESK and (b) WOL for the two-structure model shown in the lower half of the figure.

Table 2. Group velocity data in km s^{-1} and their associated period datums. Column (1) gives the mean and standard deviation determined from the three earthquakes of Table 1, and column (2) tabulates the mean from column (1) after correction for the continental portion of the propagation path.

Period (s)	(1) Observed, Iceland – ESK mean	± one sd	(2) Regionalized Iceland–Faroes Ridge
85.33	3.543	0.150	(3.381)
73.14	3.859	0.122	3.795
64.00	3.826	0.079	3.746
56.89	3.814	0.080	3.733
51.20	3.866	0.047	3.814
46.54	3.895	0.039	3.868
42.67	3.889	0.031	3.876
39.38	3.854	0.041	3.846
36.57	3.824	0.051	3.825
34.13	3.798	0.043	3.812
32.00	3.732	0.102	3.743
30.12	3.667	0.137	3.677
28.44	3.609	0.136	3.623
26.95	3.564	0.133	3.588
25.60	3.533	0.132	3.574
24.38	3.508	0.121	3.571
23.27	3.474	0.097	3.553
22.26	3.431	0.076	3.521
21.33	3.384	0.070	3.481
20.48	3.341	0.073	3.444
19.69	3.299	0.081	3.407
18.96	3.257	0.092	3.365
18.29	3.218	0.108	3.325
17.65	3.186	0.118	3.293
17.07	3.156	0.119	3.260
16.52	3.128	0.116	3.228
16.00	3.095	0.110	3.186
15.52	3.057	0.104	3.134
15.06	3.014	0.097	3.074
14.63	2.969	0.091	3.010
14.22	2.929	0.094	2.952
13.84	2.900	0.106	2.911
13.47	2.880	0.117	2.882
13.13	2.865	0.131	2.859

just two structures (Fig. 3); the IFR and the continental British Isles (BI). Ideally we would wish to determine the two unknown group velocities u_{IFR} and u_{BI} independently, operating on both ESK and WOL data with the distance-weighted average slowness formula (Knopoff 1969) as shown in Fig. 3. However, such a process is a very sensitive to scatter in the observations. With only three estimates of u_{ESK} and u_{WOL} available at each period datum, the resulting 'pure-path' group velocity curves were too irregular to be of value.

Instead, the shear velocity–depth profiles given by Clark & Stuart (1981) for the British Isles continental platform were assumed to hold for the continental portions of these paths. The model selected was one which satisfied interstation Rayleigh wave phase velocities on paths traversing both the Caledonides and the Midlands Craton. The associated group velocity curve was computed, and the observations at ESK only utilized to produce the group velocity curve due to the IFR alone.

Two of the three earthquakes analysed here were also used in the British Isles study of Clark & Stuart (1981, table 1). The reliability of the interstation phase velocity values obtained demonstrate that the Rayleigh wave train is travelling sufficiently close to its predicted great circle path that the group velocities measured here are not significantly in error due to lateral refraction.

Both the source-ESK and regionalized IFR group velocities are listed in Table 2 and displayed in Fig. 4, together with various published data. The regionalization exercise described above have in fact only adjusted the group velocities by a small amount (some 0.1 km s^{-1} at most). At periods (T) less than 40 s, the regionalized values are slightly increased relative to the observations, but the similarity of all three dispersion curves (observed at ESK, computed for the British Isles, and the IFR) in this period range suggest at once that the crustal structure of the Iceland–Faroes Ridge area is not typically oceanic.

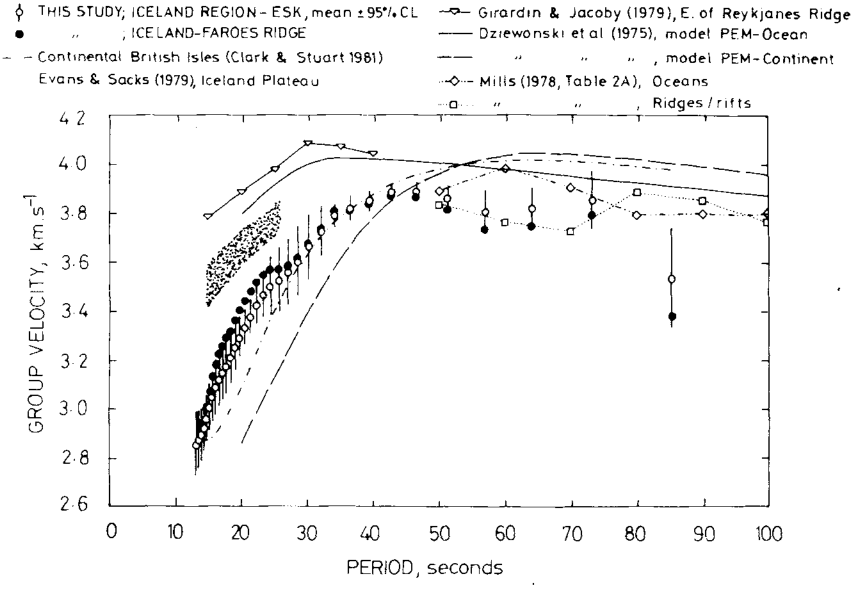

Figure 4. Observed source-ESK group velocities, continental British Isles group velocities, and the regionalized IFR dispersion curve. These are compared to observations of Mills (1978) for global-average ocean basins and rifts, Girardin & Jacoby (1979) for the Reykjanes Ridge, Evans & Sacks (1979) for the Iceland Plateau, and the predicted group velocities for the Parametric Earth Models (PEM) of Dziewonski *et al.* (1975). Uncertainty bounds indicated are 95 per cent confidence limits.

At longer periods, regionalization reduces the velocities slightly. Observational errors in a poorly-defined point at $T = 85$ s have been exaggerated by the regionalization, and it is therefore discarded from the IFR group velocity curve. The inversion to be described below will be carried out for observations with $T \leqslant 73$ s.

The following comments apply equally well to both observed and regionalized data sets. Comparison with the PEM models of Dziewonski, Hales & Lapwood (1975) at periods below *c.* 45 s suggests that lithospheric structure in the Iceland–Faroes region is intermediate between continental and oceanic, although at longer periods the observations are not distinguishable from those of Mills (1978) for global-average ocean basins and spreading centres. In Fig. 4, uncertainty bounds have been omitted from the data of Mills (1978) for clarity, but they encompass all Iceland–British Isles observations at $T > 45$ s. Group velocities determined at Icelandic stations by Girardin & Jacoby (1979) pertaining to 10–20 Ma ocean crust off the Reykjanes Ridge are typically oceanic, as also are the values reported by Evans & Sacks (1979) for the 'Iceland Plateau', between Iceland and Jan Meyen (Fig. 1). Both of these sets of observations (shown in Fig. 4) are significantly faster than those of this study, due most probably to their proximity to the MAR and thus to a younger, thinner lithosphere. Additionally, Christensen *et al.* (1980) have measured group velocities over paths to the south of the area studied here, and tabulated their values regionalized into age ranges. None of these are compatible with the observations obtained here.

Inversion procedure

The regionalized IFR data were inverted to a set of acceptable shear velocity–depth profiles using a systematic 'Hedgehog' search (Keilis-Borok & Yanovskaya 1967), the starting model for which was determined objectively using a preliminary random Monte-Carlo search. The size of the solution region in parameter space is controlled by the uncertainty bounds placed on the data and the non-uniqueness inherent in dispersion velocity inversion (Der & Landisman 1972). Knopoff & Chang (1977) note that at a given level of error in observations, group velocity offers improved inversion compared to phase velocities.

Trial dispersion curves were computed for isotropic plane-layered earth models using the Thompson–Haskell matrix formulation (Thompson 1950; Haskell 1953), then transformed to values for a spherical earth (North & Dziewonski 1976). To be considered acceptable, a model under test must generate a dispersion curve which: (1) falls entirely within the error envelope around the observations, and (2) has an rms deviation from the observations which is exceeded by the rms deviation of observations from their mean at each period datum (Stuart 1978). Ninety-five per cent confidence limits on the means (Table 2) were used to define the error envelope.

A water depth of 0.75 km was adopted as an average for all paths. An upper crustal layer (P, S velocities 5.7, 3.29 km s^{-1}; density 2.7 g cm^{-3}) from seafloor to a depth of 6 km was also assumed (after Bott & Gunnarsson 1980), and held fixed throughout the inversion. Lower crustal thickness and shear velocity were unconstrained. The shear velocities and thicknesses of a lid and low velocity zone (LVZ) made up the six independently varying unknowns. The remainder of the layer parameters were calculated from fixed interface depths of the P velocity: S velocity: density relationship of Ludwig, Nafe & Drake (1970). The layered model was bounded by a 5.2 km s^{-1} sub-LVZ layer, and a half-space beneath the 420 km discontinuity identified by England, Kennett & Worthington (1978).

The initial search was undertaken between sufficiently extreme bounds on each unknown that any geophysically feasible model would be tested, and subsequent more detailed Hedgehog searches used to define the solution region more accurately.

Results and discussion

The solutions located by the Hedgehog search are displayed as velocity–depth profiles and tabulated in Fig. 5 and Table 3 respectively.

The thickness of the lower crustal layer is restricted to between 14 km (when its shear velocity, $\beta_{LC} = 3.55\ \mathrm{km\,s^{-1}}$) and 18 km (when $\beta_{LC} = 3.7\ \mathrm{km\,s^{-1}}$). The equivalent range of P velocities is 6.15–6.45 km $\mathrm{s^{-1}}$, and the total crustal thickness is 20–24 km. The sub-Moho S velocity is determined as 4.50–4.55 km $\mathrm{s^{-1}}$ (or 7.98–8.07 km $\mathrm{s^{-1}}$ for P). Deeper structure is less well defined. Most solutions give lid thicknesses of 38–50 km or a corresponding lithospheric thickness of *c.* 60–75 km. The thicker lithospheres are associated with an LVZ of S velocity 4.35–4.4 km $\mathrm{s^{-1}}$. The Hedgehog process has, however, recovered solutions with apparently no LVZ ($\beta_{\mathrm{lid}} = \beta_{\mathrm{LVZ}}$). Noting that the longest period group velocity reliably measured is at $T = 73$ s (Table 2), resolution cannot be expected to be good below depths of 100–150 km or so. Thus these apparently shield-like profiles are not considered representative. Noting the trade-off between layer velocity and layer thickness (Table 3), it is suggested that these solutions are instead symptomatic of a slightly thicker lithosphere and/or a greater velocity contrast with the LVZ. The lower boundary of the LVZ is effectively unresolved.

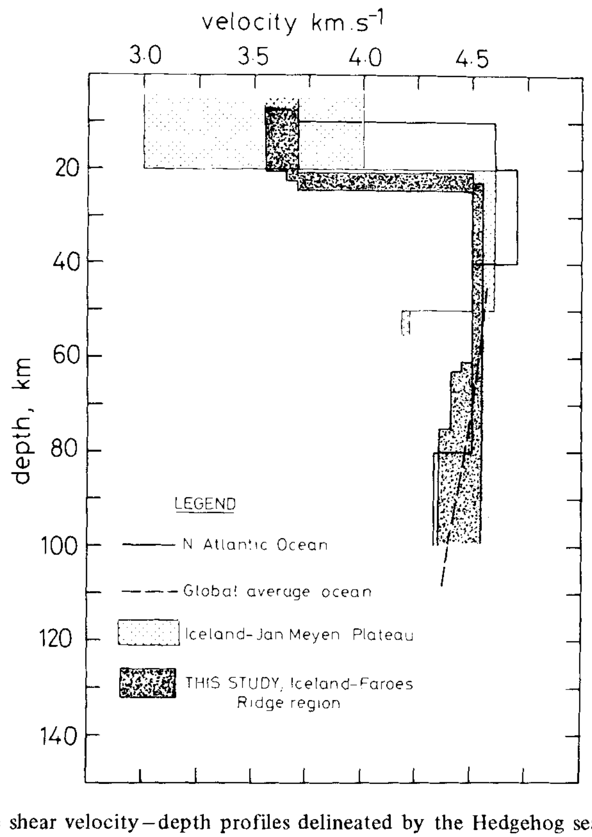

Figure 5. Acceptable shear velocity–depth profiles delineated by the Hedgehog search. The set of solutions is shown for clarity as a continuous region, but the tradeoff between individual pairs of variables may be seen in Table 3, where the parameters of these models are tabulated in full. Also included are the models of Weidner (1974), Evans & Sacks (1979), and Lévêque (1980) (denoted 'N. Atlantic Ocean', 'Iceland–Jan Mayen Plateau', and 'global average ocean' respectively).

Table 3. Details of the independent variables determined by inversion of the regionalized Iceland–Faroes data. The relationship of unlisted layer parameters (e.g. density) to those tabulated here is discussed in the text. Also shown, in parentheses at the head of each column, is the search increment associated with each variable (in the Hedgehog program). All combinations of adjacent values have been tested and rejected. Thicknesses, h, are in km; velocities, β, are in km s^{-1}. LC = lower crust, LVZ = low velocity zone.

	h_{LC} (2.0)	h_{lid} (4.0)	h_{LVZ} (25.0)	β_{LC} (0.05)	β_{lid} (0.05)	β_{LVZ} (0.05)
1	16.0	42.0	125.0	3.65	4.50	4.45
2	16.0	46.0	125.0	3.65	4.50	4.45
3	16.0	42.0	125.0	3.70	4.50	4.45
4	16.0	42.0	125.0	3.65	4.50	4.40
5	16.0	42.0	125.0	3.65	4.50	4.50
6	16.0	38.0	125.0	3.70	4.50	4.45
7	16.0	46.0	125.0	3.65	4.50	4.40
8	16.0	50.0	125.0	3.65	4.50	4.40
9	18.0	42.0	125.0	3.70	4.50	4.40
10	16.0	38.0	125.0	3.70	4.50	4.50
11	14.0	42.0	125.0	3.60	4.50	4.50
12	16.0	42.0	100.0	3.65	4.55	4.50
13	16.0	34.0	125.0	3.70	4.50	4.50
14	18.0	46.0	125.0	3.70	4.50	4.40
15	18.0	38.0	125.0	3.70	4.55	4.40
16	14.0	38.0	125.0	3.60	4.50	4.50
17	14.0	42.0	125.0	3.55	4.50	4.50
18	14.0	46.0	125.0	3.55	4.50	4.50
19	16.0	42.0	100.0	3.65	4.55	4.55
20	16.0	46.0	100.0	3.65	4.55	4.55
21	18.0	50.0	125.0	3.70	4.50	4.35
22	14.0	46.0	125.0	3.55	4.50	4.45
23	14.0	50.0	125.0	3.55	4.50	4.45

These solutions highlight the anomalous nature of this region of ocean. The Moho depth of 20–24 km appears more continental than oceanic. It agrees well with the figure of 22–31 km obtained previously for the Rockall Bank (Scrutton 1972). Crustal models reported for the IFR have varied; the results given here are adequately consistent with a Moho depth of 27 km suggested by Bott *et al.* (1974), but are somewhat shallow compared to a more recent interpretation of 28–35 km (Bott & Gunnarsson 1980). Assuming a Poisson's ratio of 0.25, the lower crustal velocities are also low relative to published body wave results (6.15–6.45 km s^{-1}, cf. 6.7 km s^{-1} from Bott & Gunnarsson 1980).

Lithospheric structure is less atypical of oceanic areas. The range of lid velocities agrees very well with the P_n velocity of 7.8 km s^{-1} measured on the IFR from refraction profiles (Bott & Gunnarsson 1980), and with the S_n velocities determined by Hart & Press (1973: 4.58 ± 0.02 km s^{-1} for areas of the N. Atlantic younger than 50 Ma). In surface wave studies in adjacent areas, Girardin & Jacoby (1979) obtained β_{lid} = 4.5–4.7 km s^{-1} and lid thickness of at least 30 km below a typical oceanic crust, off the Reykjanes Ridge. The models of Weidner (1974) for the North Atlantic Basin (included in Fig. 5) show reasonable similarity of lower lithosphere structure. The models reported by Evans & Sacks (1979) for the Iceland Plateau region, and by Lévêque (1980) for a global-average 'old' ocean, also feature good argument with lid velocity determined here. Canas & Mitchell (1981) present models for the North Atlantic regionalized in two alternative fashions (0–65 Ma/> 65 Ma: and 0–23 Ma/> 23 Ma). All of these models feature lithospheric thicknesses of 60–100 km, but they are too poorly resolved to warrant any detailed comparisons.

Conclusions

Group velocities of Rayleigh waves observed in the UK (Fig. 2) from sources in the Iceland–Jan Meyen area (Fig. 1) have been corrected for the continental portion of their path using published models for the British Isles (Figs 3, 4). Inversion to S velocity–depth profiles has yielded a well-defined Moho depth of 20–24 km and lower crustal velocity of 3.55–3.7 km s^{-1}, overlying an uppermost mantle of velocity 4.5–4.55 km s^{-1} (Fig. 5). This result supports the conclusion of Bott & Gunnarsson (1980) and earlier authors concerning the anomalous nature of the crust in the Iceland–Faroes Ridge region. The sub-crustal lithosphere appears 'transitional' in the sense that it has a similar velocity but greater thickness than that of the Iceland Plateau to the north, yet conversely has a similar thickness but reduced velocity than the North Atlantic Basin to the south (Figs 1 and 5).

As suggested by Knopoff & Chang (1977), inversion of group velocities with small uncertainty bounds is able to provide acceptable resolution in shear velocity–depth models. The results obtained from this preliminary study of the Iceland–Faroes region, and also for Azores–British Isles paths (R. A. Clark, unpublished data), suggest that the depth extent of lateral variations in crust and upper mantle seismic structure associated with anomalous areas such as continental fragments may be delineated by regionalization and inversion of a more extensive dataset.

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