Geophys. J. R. astr. Soc. (1972) 27, 259–275.

The Crustal Structure of Rockall Plateau Microcontinent

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(Received 1971 November 1)

Summary

Interpretation of seismic refraction measurements and gravity profiles from the Rockall Plateau area of the North-east Atlantic proves that the Plateau is a microcontinent. The depth to the Moho beneath the Plateau varies between 22 and 31 km. P wave velocities of about 6·4 and 7·1 km s^{-1} were obtained for the main crustal layers; these velocities may be typical of microcontinents. The Plateau is approximately in isostatic equilibrium, as too is Rockall Trough. The Trough is underlain by a thin crust, possibly of oceanic origin.

Introduction

Rockall Plateau is a large bathymetric feature of the North-east Atlantic. It comprises Hatton Bank, Hatton–Rockall Basin, Rockall Bank and George Bligh Bank. Rockall Trough is the deep water area separating the Plateau from the continental shelf west of Scotland (Fig. 1).

Current theories on the evolution of the North Atlantic Ocean basin (Le Pichon, Hyndman & Pautot 1971; Laughton 1971; Vogt *et al.* 1969) consider Rockall Plateau to be a microcontinent. Geophysical evidence in support of a continental origin for the Plateau is given by the reconstruction of the continents around the North Atlantic by Bullard Everett & Smith (1965), by seismic interpretation (Scrutton & Roberts 1971) and by preliminary gravity and magnetic interpretation (Roberts 1971). Geological evidence includes the strontium and lead isotope measurements made on the aegirine granite of Rockall Islet by Moorbath & Welke (1969). These show that at least part of the material comprising Rockall Bank has continental affinities.

In order to obtain more knowledge of the variation in crustal structure beneath Rockall Plateau long gravity traverses across the area have been interpreted using seismic refraction and seismic reflection control. The interpretation, presented in this paper, indicates that Rockall Plateau is indeed a microcontinent.

Data

Most of the seismic refraction data and all of the seismic reflection and gravity data were collected by *R.R.S. Discovery* in 1969, during Cruise 29 jointly run by the National Institute of Oceanography and the Department of Geodesy and Geophysics, Cambridge University. Other seismic refraction data were collected by M.V. *Surveyor* in 1970. Fig. 2 shows the location of the gravity profiles and the seismic control. The major gravity profile, to be interpreted in detail, is YY′ from Bloody Foreland to the flanks of the Reykjanes Ridge; XX′ and ZZ′ are supplementary profiles.

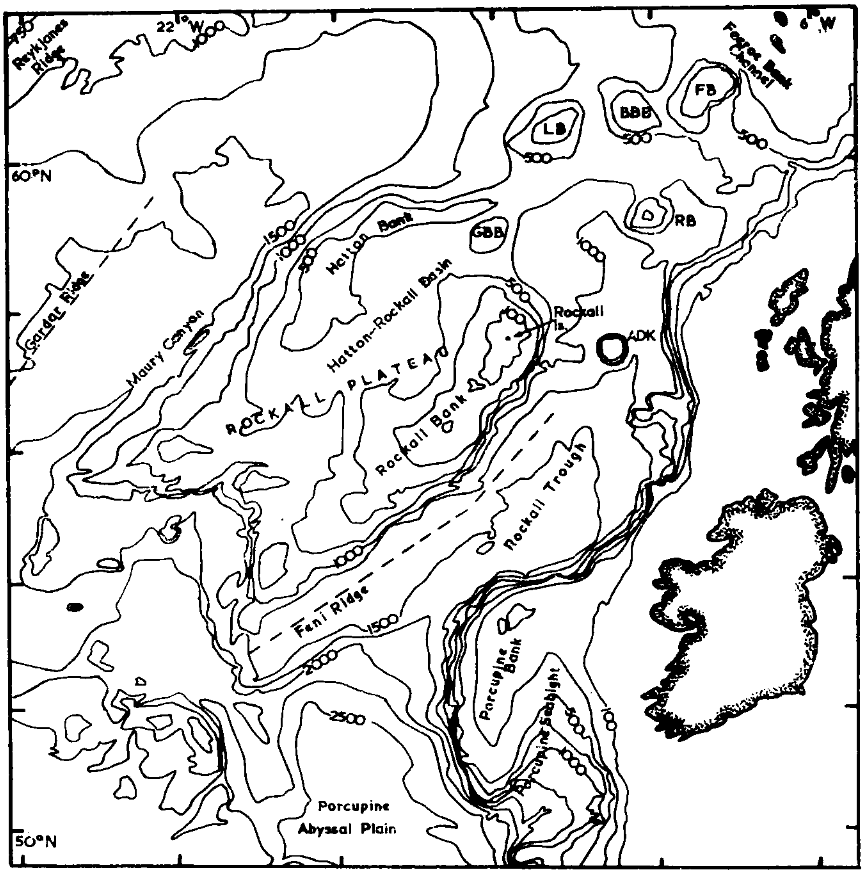

FIG. 1. Bathymetry of the Rockall Plateau area. Isobaths in fathoms. ADK—Anton Dohrn Kuppe, RB—Rosemary Bank, GBB—George Bligh Bank, LB—Lousy Bank, BBB—Bill Baileys Bank, FB—Faeroe Bank.

A Graf-Askania Gss 2 sea gravimeter mounted on an Anschutz gyro-stabilized platform was used to obtain the gravity data. Free-air gravity anomalies were computed assuming the value of 'g' at four base stations: 981,204·8 at No. 1 Dock, Barry, Wales, 981, 541·0 at the Naval wharf in Londonderry, N. Ireland, 981, 606·1 outside the Port Authority building in Greenock, Scotland and 981,630·3 over US/UK gravity range No. 2 north of Ireland. The anomalies have been corrected for the cross-coupling effect which was measured with a cross-coupling computer built by Haworth (1968).

The seismic refraction data were obtained using the single ship sono-radio buoy technique described by Hill (1963). The Cambridge sono-buoys have been developed to include a radio telemetry system together with an internal magnetic tape recording system (Gray & Owen 1969). Triple play tape provides up to 12 hr recording and the tape recorder can be delayed to switch on just before shooting begins. The travel times read from the radio and tape records have been corrected for the slant range between the shot and the ship's hull-geophone, for the depth of the shot and receiver below the sea-surface, and for the bottom topography assuming the first seismically observable sub-bottom layer to be planar. The theory behind these corrections is

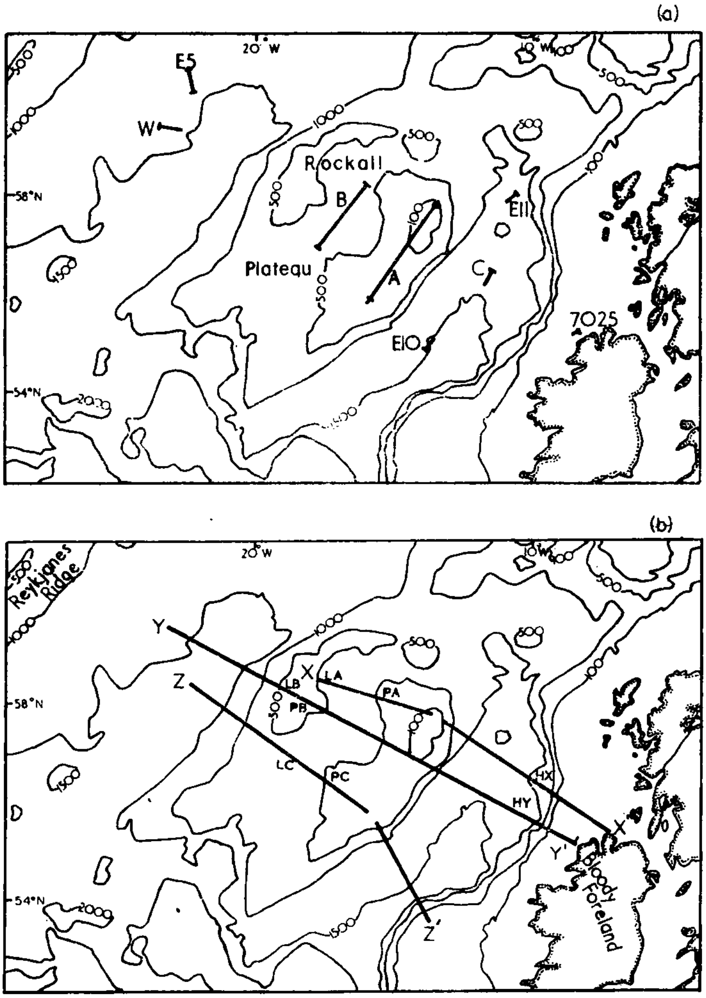

Fig. 2(a). Location of seismic refraction stations referred to in the text. Those prefixed E—Ewing & Ewing (1959). (b) Location of profiles XX', YY', ZZ'. PA, PB, PC, HX, HY are the positions of gravity highs; LA, LB, LC are the positions of gravity lows.

described by Gaskell, Hill & Swallow (1958) and Swallow (1954). Straight lines have been fitted to the data by the least squares method. The depths and dips of the refracting layers were calculated using the formulae given by Ewing, Woollard & Vine (1939).

Seismic refraction interpretation

Five refraction stations lie astride the major profile, YY'. All but station B were occupied during *Discovery* Cruise 29. Station A (Scrutton 1970; Scrutton & Roberts 1971) shows that Rockall Plateau is devoid of unconsolidated sediment cover and that it has a complex shallow structure. There are two main crustal layers beneath the Bank possessing velocities of 6·36 and 7·02 km s^{-1}, and the Moho lies at a depth of about 31 km. This structure is depicted in Fig. 3 where it is compared with that of the Seychelles Bank (Davies & Francis 1964).

Station W (at 59° 21′ N, 23° 29′ W) is an unreversed line established by Whitmarsh (1971) using a pop-up bottom seismometer. The crustal structure revealed is oceanic. The depth to the Moho of 9·6 km was obtained by assuming the upper mantle velocity of 7·83 km s^{-1} calculated for an anisotropy experiment at this locality. In Fig. 3 station W is compared with station E5 (Ewing & Ewing, 1959) which lies 50 miles to the north. There are marked differences in velocities and layer thicknesses between the stations suggesting a variable crustal structure in this area. Vogt *et al.* (1969) have already shown that a complex zone of transform faults exists here. However, the differences may be more apparent than real. It is possible Ewing and Ewing's widely-spaced data prevented them from defining layers 2 and 3, in which case station W shows the true crustal structure.

In an attempt to find the depth to the Moho beneath Rockall Trough, and thus provide valuable seismic control for the gravity interpretation, a 100 nautical miles

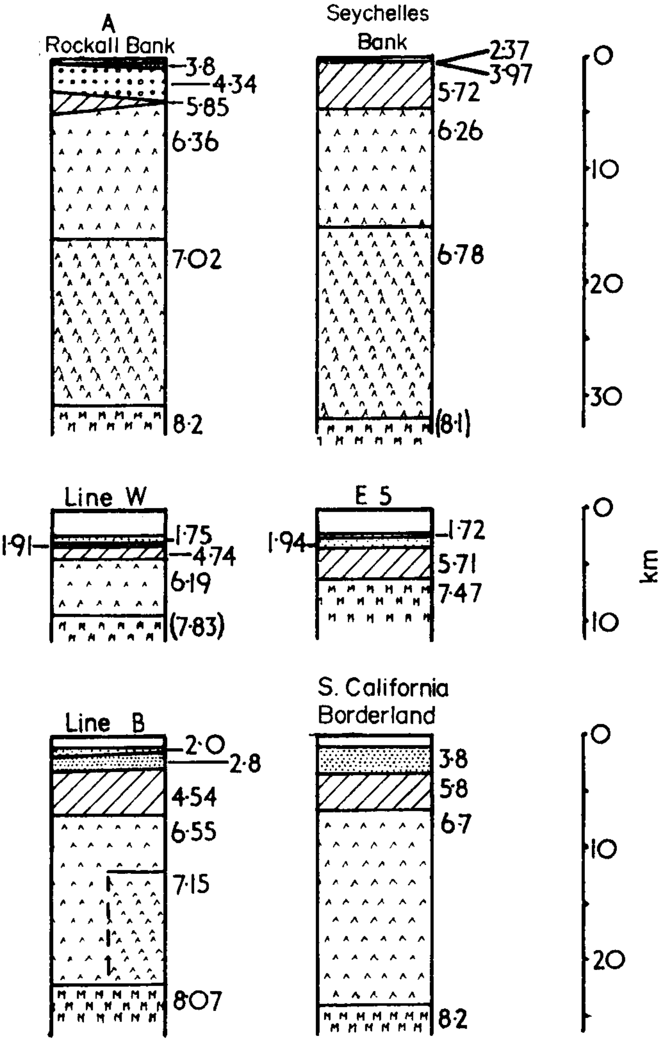

FIG. 3. Crustal structures from the Rockall area (left-hand column) compared with structures of similar type. Seychelles Bank after Davies & Francis (1964), E5 after Ewing & Ewing (1959), southern California continental borderland after Shor & Raitt (1958). Velocities in km s^{-1} and assumed velocities in parenthesis.

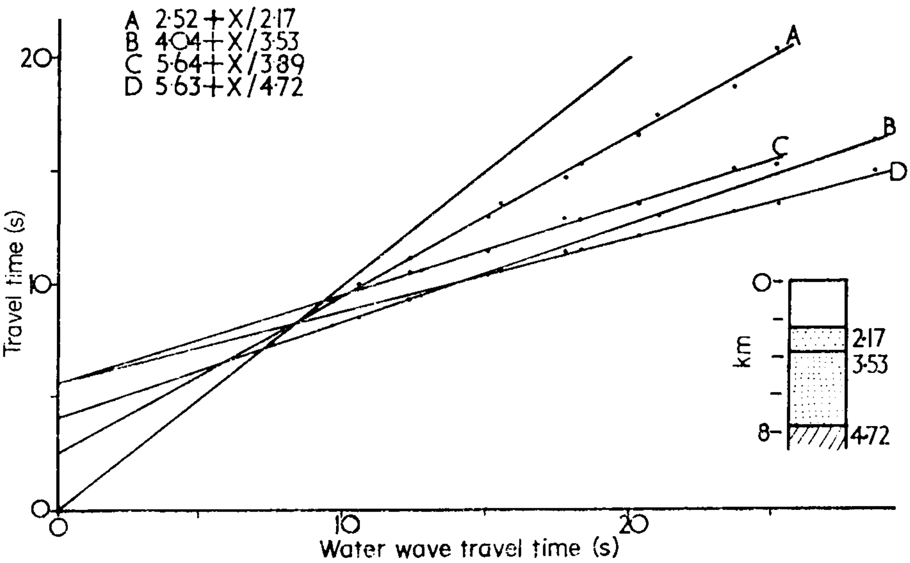

FIG. 4. Travel-time plot and crustal section for refraction line C. Equations to the straight lines on the plot are given in the form:

intercept time (s) + X/velocity (km s^{-1}).

long reversed refraction line was planned for *Discovery* station 7031. After shooting the line (denoted C) from the northern end it was found that beyond 22 nm no ground wave signal had been obtained, the reversal was therefore abandoned. A travel-time plot of the 22 nm of data is shown in Fig. 4 together with the calculated structure. The plot shows a line of weak arrivals representing a velocity of 3·89 km s^{-1}. They can be interpreted as arriving from the 3·53 km s^{-1} refractor after having been reflected once within the 2·17 km s^{-1} layer. The discrepancy in velocity between 3·89 and 3·53 km s^{-1} can be explained by a thinning of the 2·17 km s^{-1} layer towards the south. A variation in sediment layer thickness along the axis of Rockall Trough may also explain the lack of correlation between the refraction results and the seismic reflection results obtained to the south of this station (see ‘ 7031 ’ on Fig. 6). The deepest observed relector may, however, correlate with the 4·72 km s^{-1} refractor at about 5 km below the sea bed.

The structure beneath station C, together with that from stations E10 and E11 of Ewing & Ewing (1959), shows that a considerable thickness of sediment, 5 km or more, exists within Rockall Trough. The velocity of 4·72 km s^{-1} may represent consolidated sedimentary rock, but it also lies within the range of velocities for oceanic layer 2 (Ewing 1969) and could provide some support for the theory that Rockall Trough is underlain by oceanic crust (Vine 1966). Whatever its composition, sedimentary or igneous, the 4·72 km s^{-1} horizon appears to cause high attenuation of seismic signals. If this horizon represents a velocity inversion, due to a low velocity salt layer for example (Pautot, Auzende & Le Pichon 1970), ground wave arrivals would be expected to reappear on the records of the more distant shots, but none were observed. A more likely explanation is that the loss of signal was due to scattering and absorption resulting from complex structures and inhomogeneities in the rock (Ewing, Antoine & Ewing 1960).

Refraction station B, situated over Hatton–Rockall Basin, was established by M.V. *Surveyor* in 1970. The travel-time plot and the calculated structure profile are shown in Fig. 5(a) and 5(b) respectively.

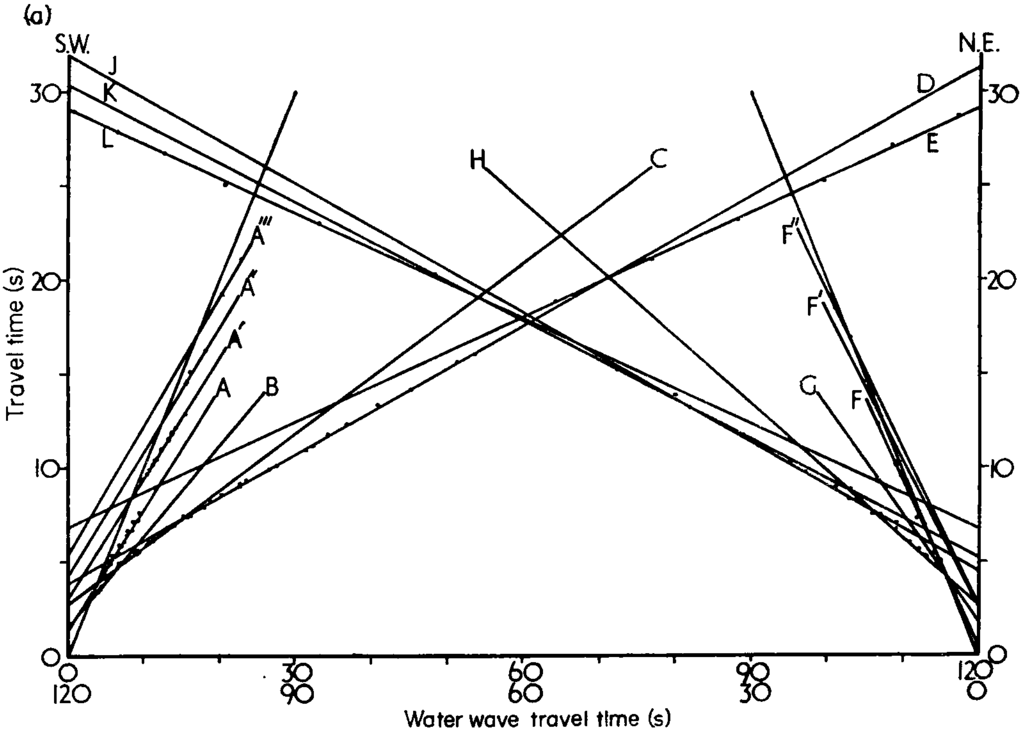

FIG. 5(a). Travel-time plot for refraction line B. Equations to the straight lines are given in the form:

intercept time (s)+X/velocity (km s^{-1}).

A 1·28±0·23+X/2·28±0·08
A' 2·92±0·15+X/2·32±0·03
A'' 4·16±0·28+X/2·25±0·02
A''' 5·26 +X/2·16 (only two points on the line)
B 1·53±0·07+X/3·06±0·07
C 2·76±0·05+X/4·95±0·04
D 3·82±0·04+X/6·55±0·03
E 6·81±0·16+X/8·07±0·04
F 0·53±0·09+X/1·70±0·02
F' 2·26±0·31+X/1·85±0·05
F'' 2·79 +X/1·80 (only two points on the line)
G 1·73±0·03+X/2·61±0·03
H 2·76±0·07+X/4·23±0·05
J 4·52±0·08+X/6·55±0·12
K 5·23±0·10+X/7·15±0·06
L 6·76±0·38+X/8·07±0·13

The shallow structure of the sediments in Hatton–Rockall Basin is well defined by the short range arrivals at the south-west end of the line, but not so well defined at the north-east end where there were fewer shots. The sub-bottom layer of unconsolidated sediments shows a marked velocity difference between the two ends of the line, 2·38–1·70 km s^{-1}. The difference may indicate that this sub-bottom layer varies in

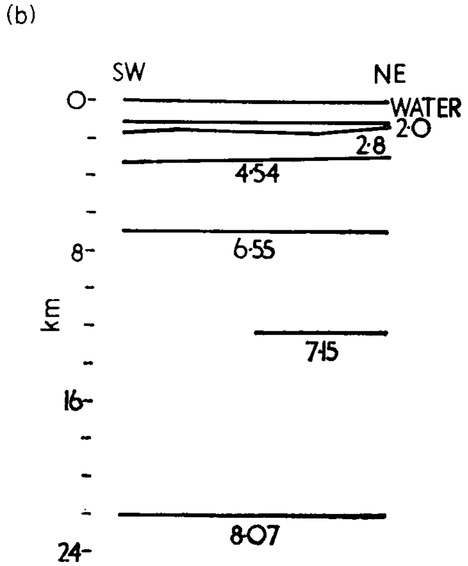

FIG. 5(b). Crustal structure revealed by line B over Hatton–Rockall Basin. Velocities in km s^{-1}.

physical properties over the Basin. The thickness of 500–700 m calculated for the unconsolidated sediments corresponds to the thickness of Miocene to recent deposits sampled at hole 116 on Leg XII of the Deep Sea Drilling Project (DSDP Scientific Staff 1970).

At both ends of line B refraction multiples (Hill 1952) which have been once, twice and three times reflected between the sea surface and the sea floor are observed. Similar arrivals were recorded by Francis (1964) near the Seychelles and less obvious ones by Gaskell *et al.* (1958) at their refraction stations at the south-west end of Hatton–Rockall Basin. These multiples suggest there is a strong acoustic contrast between the sea water and the unconsolidated sediment.

The 2·61 and 4·23 km s^{-1} velocities at the north-east end of line B correlate well with the 3·06 and 4·95 km s^{-1} at the south-west end. A true velocity of 2·80 km s^{-1} is obtained for the shallower layer, almost certainly representing semi-consolidated sediments. These sediments probably lie between reflectors 4 and 5 (see Fig. 6) of Roberts *et al.* (1970); their age was shown to be Eocene by D.S.D.P. holes 116 and 117. The deeper layer has a true velocity of 4·54 km s^{-1} and lies at a depth of 2 km beneath the sea bed. As at station C, it is not possible to decide whether 4·54 km s^{-1} represents lavas or consolidated sediments.

The deep structure beneath station B, like the shallow structure, is more clearly defined at the south-west end of the line. The 4·54 km s^{-1} layer is about 3 km thick at this end and is underlain by a single main crustal layer of velocity 6·55 km s^{-1}. At the north-east end of the line the 6·55 km s^{-1} layer is observed again, but at 34 s the travel-time curve breaks over into a 7·15 km s^{-1} velocity. This is not a clearly-defined break but it produces a crustal structure which correlates well with that observed beneath station A on Rockall Bank. The mid-crust discontinuity may be identified with the Conrad of continental areas. The consistently high crustal velocities obtained for Rockall Plateau are atypical of the continents (Steinhart & Meyer 1961; James & Steinhart 1966). Perhaps they are more representative of the transition zone

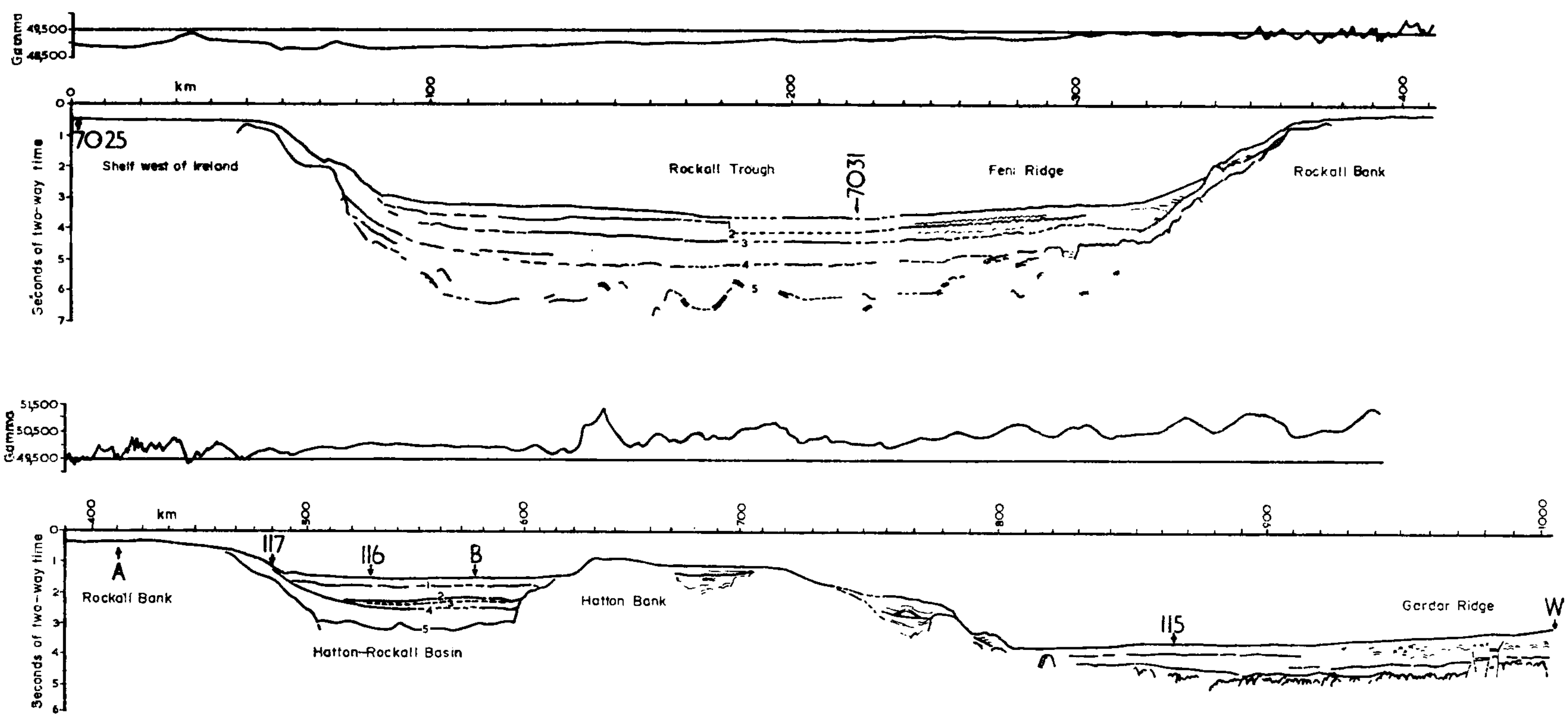

FIG. 6. Line diagram of the seismic reflection profile along YY′ (location shown in Fig. 2(b)). The positions of seismic refraction stations and Deep Sea Drilling Project drilling sites are indicated. Total magnetic field is shown above the profile. Redrawn from Scrutton & Roberts (1971).

between continent and ocean (Drake & Nafe 1968) or of a subsiding continental margin as documented by Sheridan (1969). It is interesting to note that high *P* wave velocities have also been obtained beneath the Seychelles (Davies & Francis 1964) and beneath the Faeroes Bank (Bott, private communication) which are, like Rockall, considered to be continental fragments. The Faeroes Bank may be a north-easterly continuation of the Rockall fragment.

Line B produced good first arrivals from the Moho. They give an upper mantle velocity of 8·07 km s^{-1} in both directions and intercepts that differ by only 0·06 s. The Moho appears to be a well-defined horizon and lies at a depth of 22 km b.s.l. With the result from line A we now have good seismic control on the depth to the Moho beneath Rockall Plateau.

In Fig. 3 the structure for Hatton-Rockall Basin is compared with that calculated for parts of the southern California continental borderland (Shor & Raitt 1958). The line B structure also bears close resemblance to Francis & Raitt's (1967) results from the quasi-continental Broken Ridge south-west of Australia, and to parts of the Gulf of Mexico (Ewing *et al.* 1960). These similarities suggest that Hatton-Rockall Basin is underlain by crust of continental origin.

Discovery station 7025 was a short unreversed refraction line shot on the continental shelf near Bloody Foreland to test a newly-built sono-buoy. The test provided ample data to establish a sub-bottom velocity of 6·15 km s^{-1}. The velocity undoubtedly represents the 'granitic' layer of the Earth's crust. Sediments must be very thin or absent in this area.

The results of refraction stations W, B, A, 7031 (C) and 7025 provide seismic control for the gravity interpretation which follows. They are summarized on a bathymetric profile of YY′ in Fig. 7. Also in Fig. 7 is an outline of the control available from a seismic reflection profile along YY′. The reflection profile, shown in line diagram form in Fig. 6, was obtained on *Discovery* Cruise 29 and has been described by Scrutton & Roberts (1971) and Roberts *et al*, (1970). The profile is particularly useful for defining the shape of the sedimentary basins.

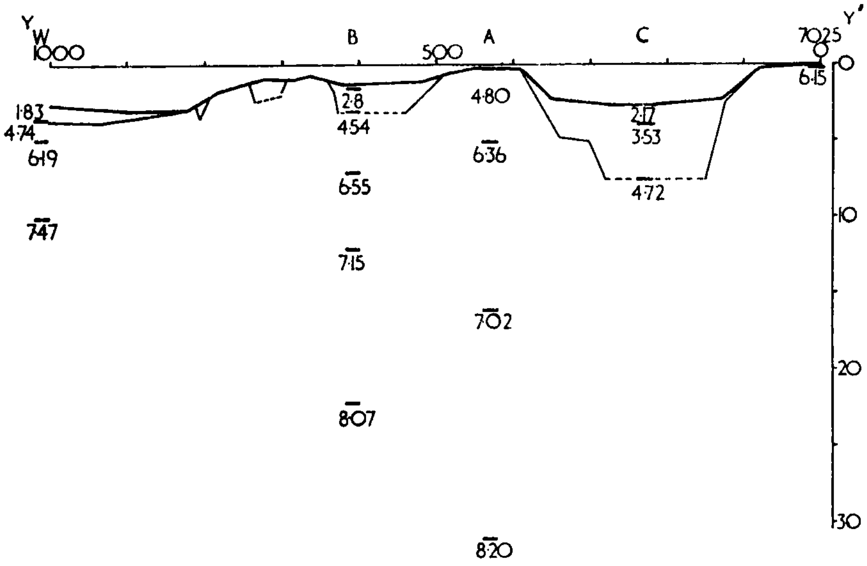

FIG. 7. Summary of available seismic data for control of the gravity interpretation along YY′ (locations shown on Fig. 2). Thin lines represent reflection data, thick lines refraction data. Velocities in km s^{-1} and scales in kilometres.

Gravity interpretation

We now proceed to study the free-air gravity profiles XX′, YY′ and ZZ′. When making a gravity interpretation it is important to keep the accuracy of the gravity data in mind. For *Discovery* Cruise 29 there are 12 track cross-overs where gravity was measured. The mean crossover error without regard to sign is 7 mgal. In the following interpretation no attempt has been made to fit the data to better than a few milligals. Furthermore profiles XX′, YY′ and ZZ′, being up to 1000 km long,will probably be affected by a regional component of gravity. Kaula (1970) reports that the Earth's gravitational field, defined up to the 16th harmonic, may decrease by as much as 20 mgal, between Iceland and the British Isles, that is a gradient of 0·015 mgal km^{-1}. An inspection of the three profiles suggested that such a gradient would fit their regional trend very well. Consequently all three profiles are corrected for a gravity gradient of $-0{\cdot}015$ mgal km^{-1} towards the south-east.

For the first step in the gravity interpretation the observed free-air anomaly for each of XX′, YY′ and ZZ′ is compared with the calculated free-air anomaly due to a two-dimensional isostatic model of the crust. The isostatic models assume Airy's hypothesis of compensation and are based on a depth of 31 km to the Moho beneath Rockall Bank. The densities of 2·84 and 3·33 $\mathrm{g\,cm}^{-3}$ used for the crust and upper mantle respectively have been obtained by averaging available seismic velocities (shown in Fig. 7), then using the extended velocity–density relationship described by Ludwig, Nafe & Drake (1970) to find the corresponding density.

Figs 8, 9 and 10 show the isostatic models and their associated gravity profiles. The observed and calculated free-air anomalies are in general agreement for each of the profiles. This suggests that the Rockall area, including Rockall Trough with its large thickness of sediments, is approximately in isostatic equilibrium. This is the conclusion Browne & Cooper (1950) came to using their pendulum gravity readings. However, marked departures from equilibrium do occur over small uncompensated structures. These may tell us something about the geology of the area.

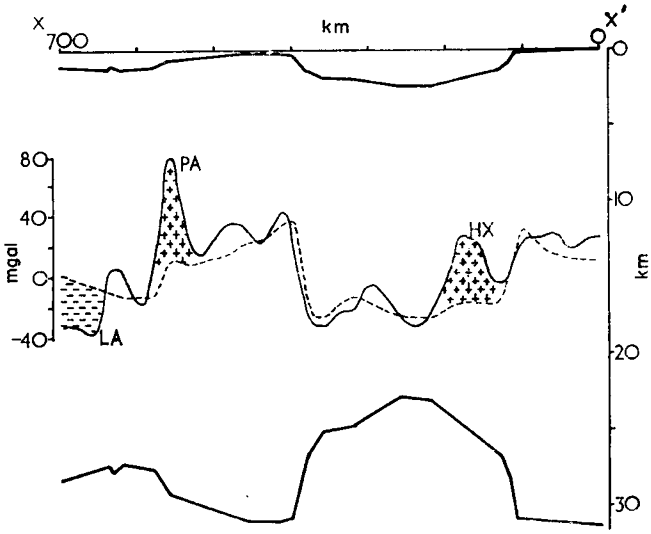

FIG. 8. Profile XX′ (location shown in Fig. 2(b)) showing the observed free-air gravity anomaly ——, and the free-air anomaly calculated for the Airy isostatic model of the crust - - - -. Densities of 2·84 and 3·33 $\mathrm{g\,cm}^{-3}$ were used for the crust and mantle respectively.

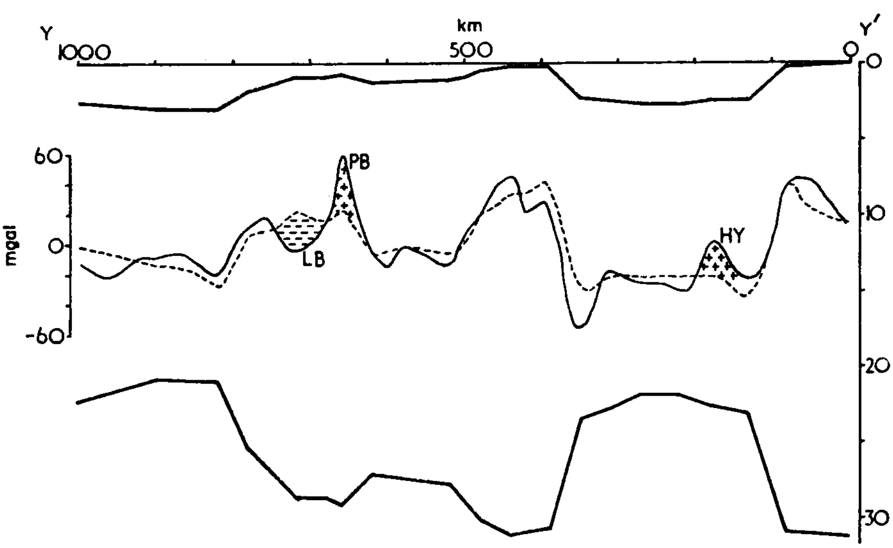

FIG. 9. Profile YY' (location shown in Fig. 2(b)) showing the observed free-air gravity anomaly ——, and the free-air anomaly calculated for the Airy isostatic model of the crust - - - -. Densities of 2·84 and 3·33 g cm^{-3} were used for the crust and mantle respectively.

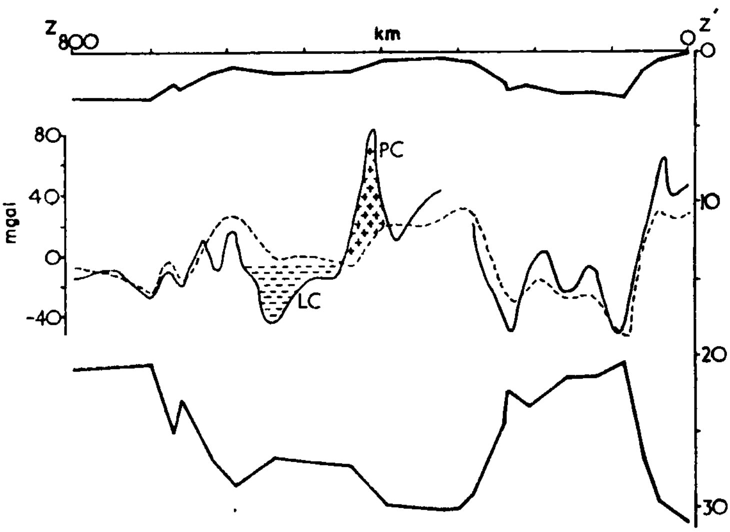

FIG. 10. Profile ZZ' (location shown in Fig. 2(b)) showing the observed free-air gravity anomaly ——, and the free-air anomaly calculated for the Airy isostatic model of the crust - - - -. Densities of 2·84 and 3·33 g cm^{-3} were used for the crust and mantle respectively.

Positive anomalies PA, PB and PC (Figs 8, 9, 10 and 2(b)) are remarkably similar in shape and they all lie above prominent bathymetric features. Their most likely cause is high density basic intrusions at shallow depth. Gravity high HX is almost certainly due to the Hebrides Terrace lava plateau, while high HY is probably due to a structure related to the Terrace, perhaps the magma chamber. All the prominent gravity deficiencies, LA, LB and LC, occur above sedimentary basins and probably indicate incomplete compensation of the sediments. LA occurs where the sediments of Hatton–Rockall Basin are known to be thick (Roberts *et al.* 1970). LB may represent the south-westward continuation of the thick sediments.

Having gained a general impression of the way in which gravity varies across the Rockall area a more detailed interpretation of the free-air gravity along profile YY′ can be made. To design a more complex, layered model of the Earth's crust the available seismic data (Fig. 7) have been correlated along the profile. The densities of the component layers have been obtained by averaging the velocity data as shown in Table 1, and then reading the corresponding density from the extended velocity–density curve. The density of 3·16 g cm^{-3} for the upper mantle west of Hatton Bank corresponds to the velocity of 7·47 km s^{-1} obtained at refraction station E5 by Ewing & Ewing (1959).

The final model of the Earth's crust is depicted in Fig. 11 together with the associated gravity anomalies. The model was obtained by adjusting the shape of the component layers until the calculated free-air anomaly was in general agreement with the observed. It must be remembered that it is not possible to arrive at a unique model by this method, a different one may fit the observed gravity data equally well.

The final model reveals the variation in deep crustal structure across Rockall Plateau and Rockall Trough. The structures established by refraction lines A and B, over Rockall Bank and Hatton–Rockall Basin respectively, produce a calculated anomaly which is in good agreement with the observed gravity. This is most encouraging and suggests that the interpretation is meaningful. Apart from confirming the

Table 1

Velocity (km s^{-1})	Source station	Mean velocity (km s^{-1})	Density (g cm^{-3})
(2·0) 2·17 2·08 1·83	 C B W	2·02	1·93
2·80	B		2·12
3·53	C		2·26
4·72 4·54 4·80 4·74	C B A W	4·70	2·52
6·15 6·36 6·55 6·19	7025 A B W	6·31	2·81
7·02 7·15	A B	7·09	3·07
(7·47)			3·16
8·20 8·07	A B	8·14	3·33

(7·47) assumed velocity

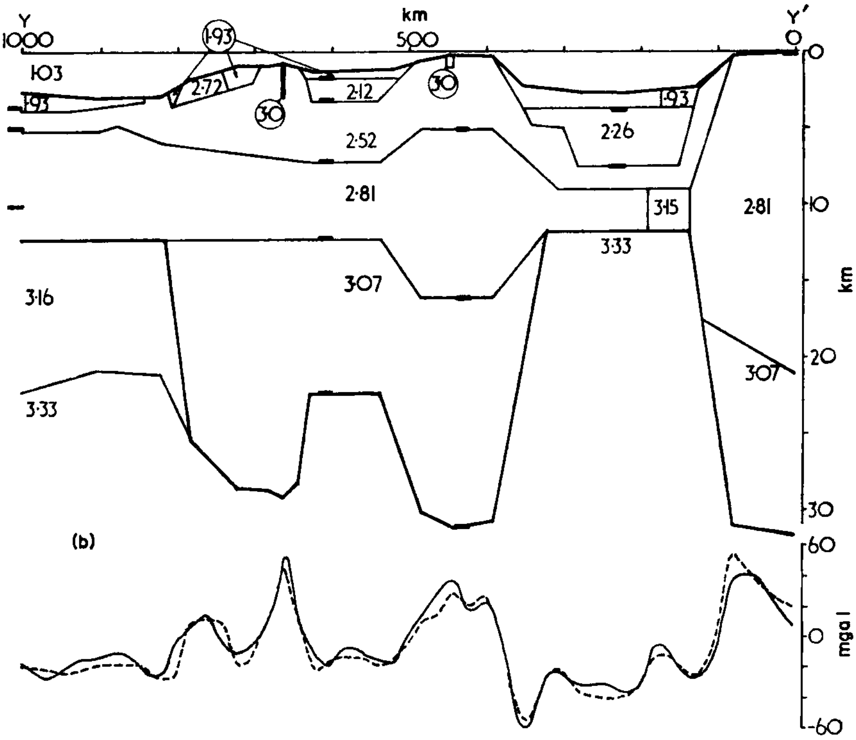

FIG. 11(a). Final model of the Earth's crust along profile YY′ (location shown in Fig. 2(b)). Very thick lines are seismic refraction control. Densities in g cm^{-3}. (b) Observed free-air gravity anomaly along YY′ ——, and the free-air anomaly calculated for the model above – – – –.

continental nature of the crust beneath Rockall Plateau, a notable feature of the model is the calculated structure beneath Rockall Trough. To maintain isostatic equilibrium here the Moho has been brought up to about 12 km below sea level. This is strong evidence in favour of Rockall Trough being underlain by oceanic crust (Vine 1966). However, equilibrium could also be maintained with a thickened crust and a high density upper mantle. The 5 km of sediment that have accumulated above layer 2 of the Trough suggests that some sinking of the Trough has taken place, and crustal thickening is normally associated with the sinking of small ocean basins (Menard 1967). It is unfortunate we have no deep seismic control beneath the Trough, since this would enable us to determine the thickness of the crust and the density of the upper mantle.

Some other features of the crustal model are worthy of note. The block of density 3·15 g cm^{-3} beneath Rockall Trough accounts for gravity high HY (Fig. 9) which is considered to be connected with high HX situated over the Hebrides Terrace lava plateau. The block may represent part of the magma reservoir for the Terrace. On Rockall and Hatton Banks high density bodies are further evidence of igneous activity in the Rockall area. West of Hatton Bank the observed gravity profile does not show a strong continental edge effect. To account for the effect's absence the Moho calculated for the isostatic model of the crust (Fig. 9) has been preserved as a boundary between normal mantle and low density mantle. The Moho is therefore placed at about 12 km, below a thick oceanic crust. This combination of anomalous mantle and thick crust is similar to the Mid-Atlantic Ridge structure established by

Talwani, Le Pichon & Ewing (1965). Profile YY′ is about 400 km from the ridge crust, placing it at the limit of Talwani *et al.'s* anomalous zone. A partial explanation for the density difference in the mantle may lie in Green & Ringwood's (1963) classification of mineral assemblages in the upper mantle. They suggest that the assemblages 'plagioclase pyrolite' or 'ampholite' (mean density 3·25 g cm^{-3}) lie beneath the oceanic Moho while 'pyrolite' (density 3·31 g cm^{-3}) occurs under the continents.

Conclusions and discussion

The interpretation of gravity data using the control provided by seismic refraction and seismic reflection measurements reveals that Rockall Plateau is underlain by crust of continental type. *P* wave velocities of 6·4 and 7·1 km s^{-1} were measured for the crust, and the Moho was calculated to lie between 22 and 31 km depth. The shallow structure of the Plateau is complex and typical of a continental block: restricted gravity highs may be associated with igneous activity while lows may represent local thickening of sediments. The gravity interpretation also indicates that Rockall Trough is underlain by a thin crust, possibly oceanic in origin. This is an important conclusion since other seismic and magnetic data from Rockall Trough (Roberts 1971; Scrutton & Roberts 1971) can be equally well interpreted in terms of either a continental or oceanic basement. The Trough separates Rockall microcontinent from the European continental shelf.

Throughout the world's oceans there are numerous examples of submerged blocks like Rockall that are known, or thought, to be of continental origin. The best example is the Seychelles–Mascarene Plateau (Matthews & Davies 1966); other examples include Broken Ridge (Francis & Raitt 1967), Kerguelen Plateau, Agulhas Plateau (Graham & Hales 1965), Galicia Bank (Black *et al.* 1964), Flemish Cap, Porcupine Bank (Gray & Stacey 1970) and Fareoes Bank (Bott & Watts 1971). All these features are separated from the continental shelf by thinned continental crust or true oceanic crust. Where seismic velocities are available for these continental fragments, or microcontinents, they are anomalously high, being about 6·4 and 7·0 km s^{-1} for the upper and lower crust respectively. These velocities may be typical of small, sunken continental blocks. Where the depth to the Moho has been calculated it is usually between 20 and 30 km below sea level. Upper mantle velocities appear to be typical, about 8·1 km s^{-1}.

The gravity results presented in this paper show that Rockall Plateau is approximately in isostatic equilibrium according to the Airy hypothesis. Matthews & Smith (1971) have pointed out that all similar continental fragments are in isostatic equilibrium even though they have sunk or are sinking. The high seismic velocities mentioned above suggest that a denser-than-normal continental crust underlies these blocks (nearly 3·0 g cm^{-3} compared with 2·84 g cm^{-3} (Worzel & Shurbet 1955)). If isostatic balance has always been maintained, as Matthews and Smith suppose, then the high velocities may indicate that the crust increased in density as it sank. Crustal thinning may have taken place at the same time. Alternatively, the crust may always have been denser-than-normal, in which case the fragments would have been out of equilibrium before they became detached from the main continental mass. In order to restore equilibrium they have sunk. Further, differential sinking of the blocks may have taken place by one or more of the mechanisms described by Collette (1968), Sheridan (1969) or Bott (1971).

Acknowledgments

I gratefully acknowledge the help and advice given by Dr D. H. Matthews in all stages of this work. I thank Dr J. E. Nafe and Mr J. M. Woodside for critically reading the manuscript, and Professor M. H. P. Bott and Dr R. B. Whitmarsh for advice and information. Thanks also to the Masters, Officers and crews of R.R.S. *Discovery* and M.V. *Surveyor* for their co-operation at sea. This work was supported by the Natural Environment Research Council, London under grant number 1294 to the University of Cambridge.

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References

Browne, B. C. & Cooper, R. I. B., 1950. The British Submarine Gravity Surveys of 1938 and 1946, *Phil. Trans. R. Soc. Lond.*, **242A,** 243–310.

Black, M., Hill, M. N., Laughton, A. S. & Matthews, D. H., 1964. Three non-magnetic seamounts off the Iberian coast, *Q. Jl geol. Soc. Lond.*, **120,** 477–517.

Bott, M. H. P., 1971. Evolution of young continental margins and formation of shelf basins, *Tectonophysics*, **11,** 319–328.

Bott, M. H. P. & Watts, A. B., 1971. Deep structure of the continental margin adjacent to the British Isles. Rep. No. 70/14, Inst. geol. Sci., *Geology of the East Atlantic Continental Margin*, SCOR Symposium, Cambridge, 1970, ed. F. M. Delany.

Bullard, E. C., Everett, J. E. & Smith, A. G. 1965. The fit of the continents around the Atlantic, *Phil. Trans. R. Soc. Lond.*, **258A,** 41–51.

Collette, B. J., 1968. On the subsidence of the North Sea area, *Geology of Shelf Seas* ed. D. T. Donovan, Oliver and Boyd, Edinburgh.

Davies, D. & Francis, T. J. G., 1964. The crustal structure of the Seychelles Bank, *Deep Sea Res.*, **11,** 921–927.

Drake, C. L. & Nafe, J. E., 1968. The transition from ocean to continent from seismic refraction data, *The Crust and Upper Mantle of the Pacific area,* eds L. Knopoff, C. L. Drake and P. J. Hart, (A.G.U. Geophysical Monograph 12, Washington).

DSDP Scientific Staff (1970). Deep sea drilling project: Leg 12, *Geotimes*, **15,** 9, 10–13

Ewing, J., 1969. Seismic model of the Atlantic Ocean, *The Earth's Crust and Upper Mantle*. ed. Pembroke J. Hart (A.G.U. Geophysical Monograph 13, Washington.)

Ewing, J., Antoine, J. & Ewing, M., 1960. Geophysical measurements in the Western Caribbean Sea and in the Gulf of Mexico, *J. geophys. Res.*, **65,** 4087–4126.

Ewing, M., Woollard, G. P. & Vine, A. C., 1939. Geophysical investigations in the emerged and submerged Atlantic coastal plain, Part III: Barnegat Bay, New Jersey section, *Bull. geol. Soc. Am.*, **50,** 257–296.

Ewing, M. & Ewing, J., 1959. Seismic refraction measurements in the Atlantic Ocean Basins, in the Mediterranean Sea, on the Mid-Atlantic Ridge and in the Norwegian Sea, *Bull. geol. Soc. Am.*, **70,** 291–317.

Francis, T. J. G. & Raitt, R. W., 1967. Seismic refraction measurements in the southern Indian Ocean, *J. geophys. Res.*, **72,** 3015–3042.

Gaskell, T. F., Hill, M. N. & Swallow, J. C., 1958. Seismic measurements made by H.M.S. *Challenger* in the Atlantic, Pacific and Indian Oceans and in the Mediterranean Sea, *Phil. Trans. R. Soc. Lond.*, **251A,** 23–83.

Graham, K. W. T. & Hales, A. L., 1965. Surface ship gravity measurements in the Agulhas bank area south of South Africa, *J. geophys. Res.*, **70,** 4005–4011.

Gray, F. & Owen, T. R. E., 1969. *A recording sono-radio buoy for seismic refraction work*, Oceanology International 69 Conference, Brighton, 1969.

Gray, F. & Stacey, A. P., 1970. Gravity and magnetic interpretation of Porcupine Bank and Porcupine Bight, *Deep Sea Res.*, **17**, 467–475.

Green, D. H. & Ringwood, A. E., 1963. Mineral assemblages in a model mantle composition, *J. geophys. Res.*, **68**, 937–945.

Haworth, R. T., 1968. *Cross-Coupling Errors in Gravity Measurements at Sea*, Unpublished Ph.D. Thesis, University of Cambridge.

Hill, M. N., 1952. Seismic refraction shooting in an area of the eastern Atlantic Ocean, *Phil. Trans. R. Soc. Lond.*, **244A**, 561–596.

Hill, M. N., 1963. Single-ship seismic refraction shooting, *The Sea*, Vol. 3, ed. M. N. Hill, Interscience, London.

James, D. E. & Steinhart, J. S., 1966. Structure beneath Continents: A critical review of explosion studies 1960–1965, *The Earth beneath the Continents*, eds J. S. Steinhart and J. Smith (A.G.U. Geophysical Monograph, 10, Washington).

Kaula, W. M., 1970. Earths gravity field: relation to global tectonics, *Science*, **169**, 3949, 982–985.

Laughton, A. S., 1971. South Labrador Sea and the evolution of the North Atlantic, *Nature*, **232**, 612–617.

Le Pichon, X., Hyndman, R. & Pautot, G., 1971. A geophysical study of the opening of the Labrador Sea, *J. geophys. Res.*, **76**, 4724–4743.

Ludwig, W. J., Nafe, J. E. & Drake, C. L., 1970. Seismic refraction, *The Sea*, Vol. 4, Part 1, ed. A. E. Maxwell, Wiley-Interscience, London.

Matthews, D. H. & Davies, D., 1966. Geophysical studies of the Seychelles Bank, *Phil. Trans. R. Soc. Lond.*, **259A**, 227–239.

Matthews, D. H. & Smith, S. G., 1971. The sinking of Rockall Plateau, *Geophys. J. R. astr. Soc.*, **23**, 491–498.

Menard, H. W., 1967. Crust under small ocean basins, *J. geophys. Res.*, **72**, 3061–3088.

Moorbath, S. & Welke, H., 1969. Isotopic evidence of the continental affinity of the Rockall Bank, North Atlantic, *Earth Planet. Sci. Lett.*, **5**, 211–216.

Pautot, G., Auzende, J. M. & Le Pichon, X., 1970. Continuous deep sea salt layer along North Atlantic margins related to early phase of rifting. *Nature*, **227**, 351–354.

Roberts, D. G., 1971. New geophysical evidence on the origins of the Rockall Plateau and Trough, *Deep Sea Res.*, **18**, 353–360.

Roberts, D. G., Bishop, D. G., Laughton, A. S., Ziolkowski, A. M., Scrutton, R. A. & Matthews, D. H., 1970. New sedimentary basin on Rockall Plateau, *Nature* **225**, 170–172.

Scrutton, R. A., 1970. Results of a seismic refraction experiment on Rockall Bank, *Nature*, **227**, 826–827.

Scrutton, R. A. & Roberts, D. G., 1971. The structure of Rockall Plateau and Trough, North East Atlantic. Rep. No. 70/14, Inst. geol. Sci., *Geology of the East Atlantic Continental Margin*, SCOR Symposium, Cambridge, 1970, ed. F. M. Delany.

Sheridan, R. E., 1969. Subsidence of continental margins, *Tectonophysics*, **7**, 219–229.

Shor, G. G. & Raitt, R. W., 1958. *Seismic studies in the southern California borderland*, Congreso Geologico Internacional, Seccion IX, Geofsica Applicada (Segundo Tomo), Mexico, p. 243–259.

Steinhart, J. S. & Meyer, R. P., 1961. Explosion studies of continental structure, *Carnegie Institution of Washington*, *Publication* 622.

Swallow, J. C., 1954. *Seismic investigations at sea*, unpublished Ph.D. Thesis, University of Cambridge.

Talwani, M., Le Pichon, X. & Ewing, M., 1965. Crustal structure of midocean ridges: 2, *J. geophys. Res.*, **70**, 341–352.

Vine, F. J., 1966. Spreading of the ocean-floor: new evidence, *Science*, **154,** 1405–1415.

Vogt, P. R., Avery, O. E., Schneider, E. D., Anderson, C. N. & Bracey, D. R., 1969. Discontinuities in sea-floor spreading, *Tectonophysics*, **8,** 285–317.

Whitmarsh, R. B., 1971. Seismic anisotropy of the uppermost mantle absent beneath the east flank of the Reykjanes Ridge, *Bull. seism. Soc. Am.*, **61,** 1351–1368.

Worzel, J. L. & Shurbet, G. L., 1955. Gravity interpretations from standard oceanic and continental crustal sections, *Crust of the Earth*, ed. A. Poldervaart, Geol. Soc. Am. Special Paper 62.