MARINE GEOPHYSICAL INVESTIGATIONS:

ROCKALL TROUGH TO

PORCUPINE SEABIGHT

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Ph.D.
University of Edinburgh
1983

This thesis has been composed by myself, and all the work described is my own, except where stated otherwise.

ABSTRACT

This thesis describes the collection, processing, and interpretation of marine geophysical data from southern Rockall Trough, Porcupine Abyssal Plain and Porcupine Bank. 4,200 km of gravity, magnetic, bathymetric and single channel seismic data were collected on two cruises. Time to basement and free air gravity anomaly maps are produced. A new feature, the Clare Lineament, is mapped. The Lineament lies at the southern end of Rockall Trough, trends WNW-ESE, and links the Charlie Gibbs Fracture Zone with Porcupine Bank. A colinear magnetic feature, the Clare Trend, crosses Porcupine Bank. The Clare Lineament and Clare Trend are interpreted as the probable offshore continuation of the Iapetus Suture. A 500 gamma magnetic anomaly to the south of the Clare Trend is correlated with anomalies of the Avalon Platform of Newfoundland. Modelling of the crustal thickness beneath southern Porcupine Seabight indicates that the crust may be as thin as 2 km. The mapped continuity of structure across the mouth of the Seabight, from Porcupine Bank to Goban Spur, shows that the Seabight must be floored by continental crust. It is thought that the southern Porcupine Seabight was initiated as a Devonian wrench fault basin, with further Permo-Triassic and Mesozoic stretching episodes. The continent-ocean boundary is mapped in Porcupine Abyssal Plain, approximately 55 km west of Porcupine Bank. The continent-ocean boundary lies about 10 km to the east of anomaly 32(34). Initial spreading is therefore of Campanian or Santonian age. Porcupine Bank to the north of the Clare Trend is characterised by NNE-SSW Caledonide trends. A graben within the Bank and probable igneous centres are described. Within southern Rockall Trough, magnetic modelling and interpretation of the seismic profiles show that magnetic lineations in the centre of the Trough are caused by shallow igneous intrusions within continental crust, and not by sea-floor spreading reversals. In southern Rockall Trough massive fracturing and intrusion occurred in the Upper Cretaceous when the Charlie Gibbs Fracture Zone was initiated as a leaky transform fault. Crustal thicknesses in southern Rockall Trough are approximately 7 km, giving a stretching factor of 4. The sedimentary section, which is up to 4 seconds (TWT) thick, is described in a separate paper by Dingle, Megson, and Scrutton (1982). Probable Late Cretaceous/Palaeocene sills are present within the section. Southern Rockall Trough is probably the site of Permo-Triassic, Jurassic, and Cretaceous rifting episodes. A maturation study suggests that deeply buried sediments within the Rockall Trough are within the oil window of hydrocarbon generation.

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CHAPTER 1

REGIONAL GEOLOGY AND CRUSTAL STRUCTURE

The first section of this chapter consists of a brief literature review of the crustal structure and geology of the continental margin west of the UK and Ireland, concentrating on those areas for which this thesis presents new data; Porcupine Seabight, Porcupine Bank, Porcupine Abyssal Plain, and Rockall Trough.

The latest major event to affect this margin was the Cretaceous and Tertiary sea floor spreading which produced the North Atlantic and separated the European margin from its conjugate margin: the east coast of Canada. One objective of this study is to clarify the early sea floor spreading history of the Irish sector of the European margin; there are short sections on North Atlantic spreading, the Canadian margin, and published Atlantic refits in this chapter. Published theories on the evolution of the Northwest European margin are also to be presented, which at this stage should be regarded as hypotheses which will be tested with new data, and which are discussed further in the following chapters. The final section of this chapter outlines the problems with which this thesis is concerned.

1.1 THE CONTINENTAL MARGIN WEST OF THE UK AND IRELAND

Figure 1.1 shows the physiography and sedimentary basins of the continental margin west of the UK and Ireland. The first sub-section discusses the shelf basins very briefly, and following sub-sections consider in more detail the physiographic units adjacent to and in the study area.

1.1.1 THE SHELF BASINS

Diagrammatic cross-sections of most of the shelf basins are given in the following figures: Fig. 1.2, West Shetland, North Minch and South Minch Basins; Fig. 1.3, the Outer Hebrides Basin; Fig. 1.4, Fastnet Basin; and Fig. 1.5, the Celtic Sea and Western Approaches Basins. In all cross-sections the Tertiary and Quaternary sediments are shown stippled.

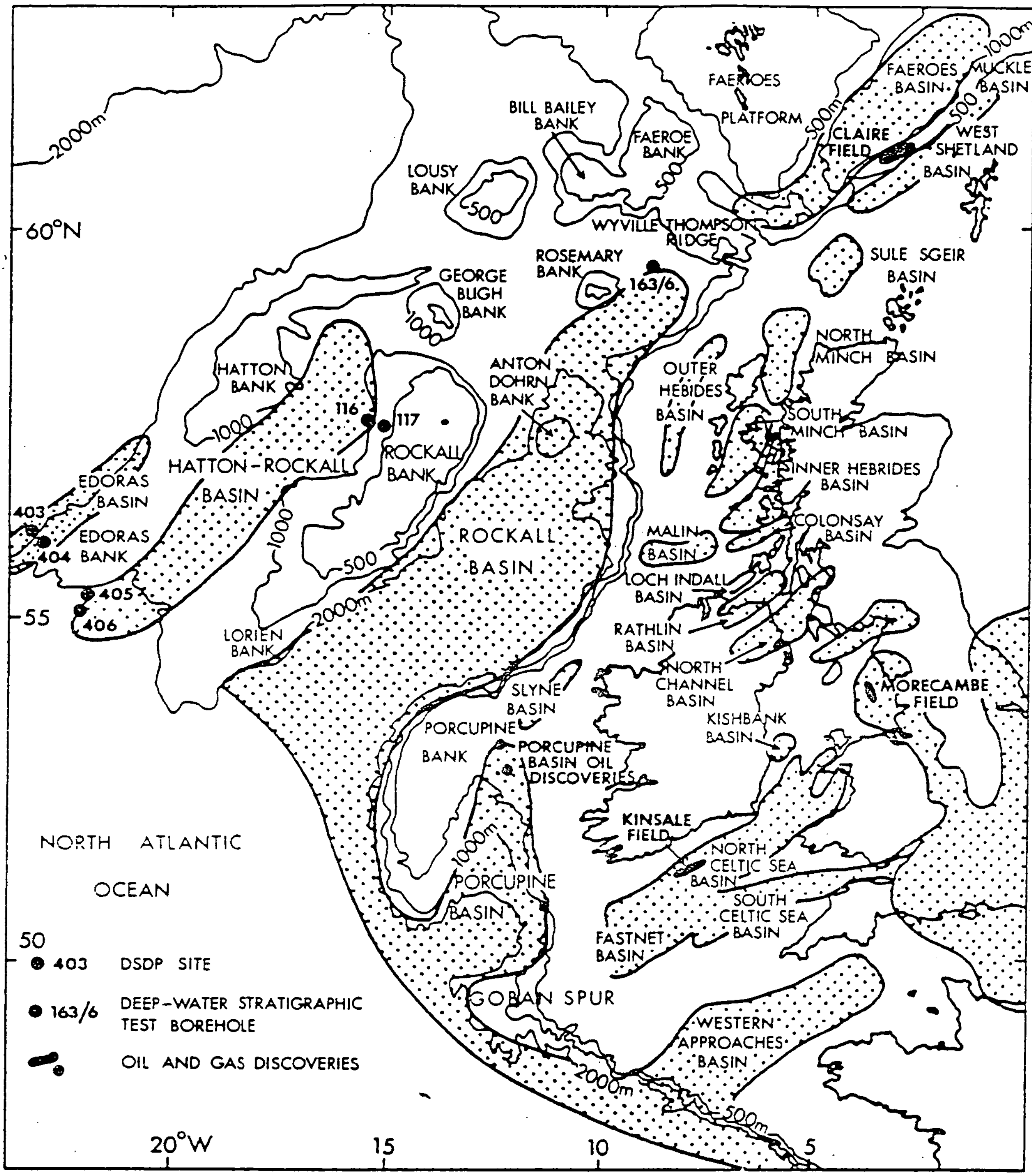

Fig. 1.1 Sedimentary basins west of the UK and Ireland (Stow, 1981)

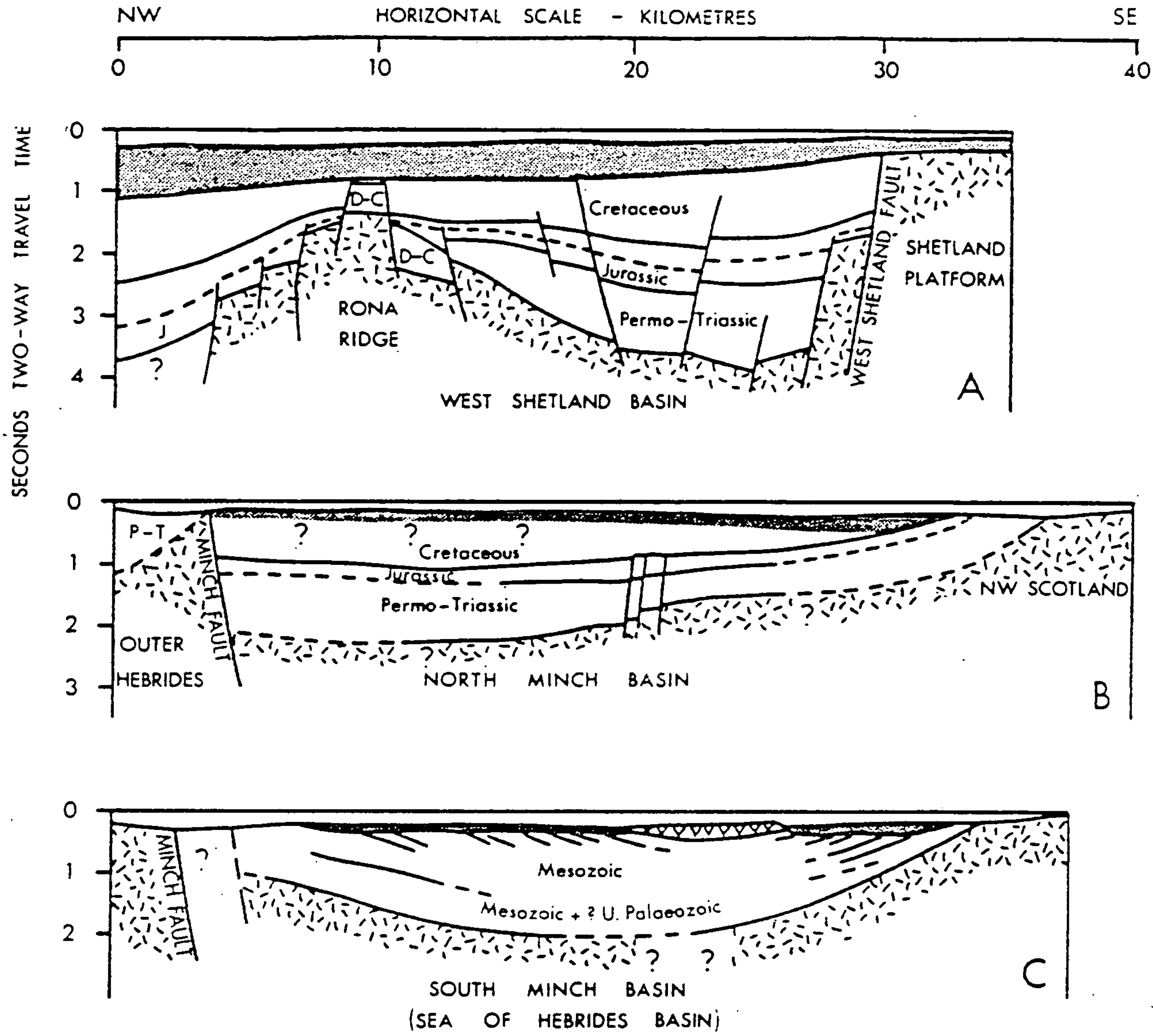

Fig. 1.2 Interpretative cross-sections of the West Shetland, North Minch, and South Minch Basins, by Stow, (1981). The Tertiary and Quaternary are shaded.

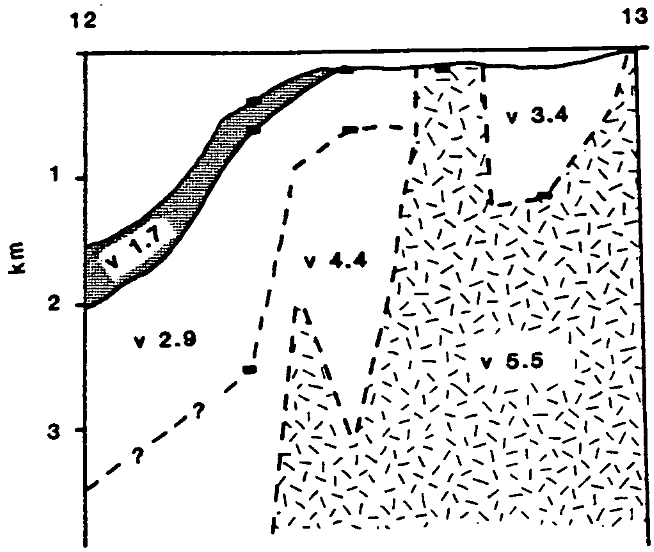

Fig. 1.3 Inferred section across the Outer Hebrides Basin, by Jones (1978). Refer to Fig. 1.9 for profile position.

Very heavy lines	: seismic refraction control
Solid lines	: seismic reflection control
Dashed lines	: interpreted boundary positions, based on seismic refraction data and gravity modelling.
V_p = 1.7 km/s	: Tertiary and Quaternary sediments
V_p = 2.9 km/s	: ?Jurassic - Cretaceous sediments
V_p = 4.4 km/s	: Pre-Jurassic sediments. ?Permo-Triassic
V_p = 3.4 km/s	: Mesozoic sediments
V_p = 5.5 km/s	: Lewisian

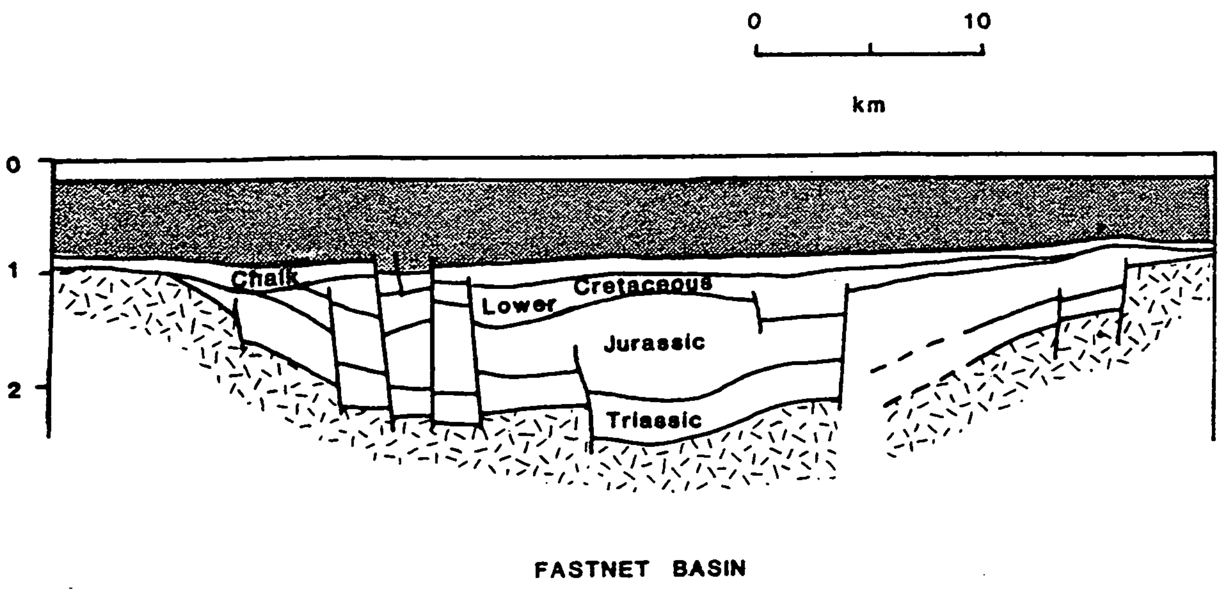

Fig. 1.4 Interpretative section across the Fastnet Basin, after Robinson et al, 1981

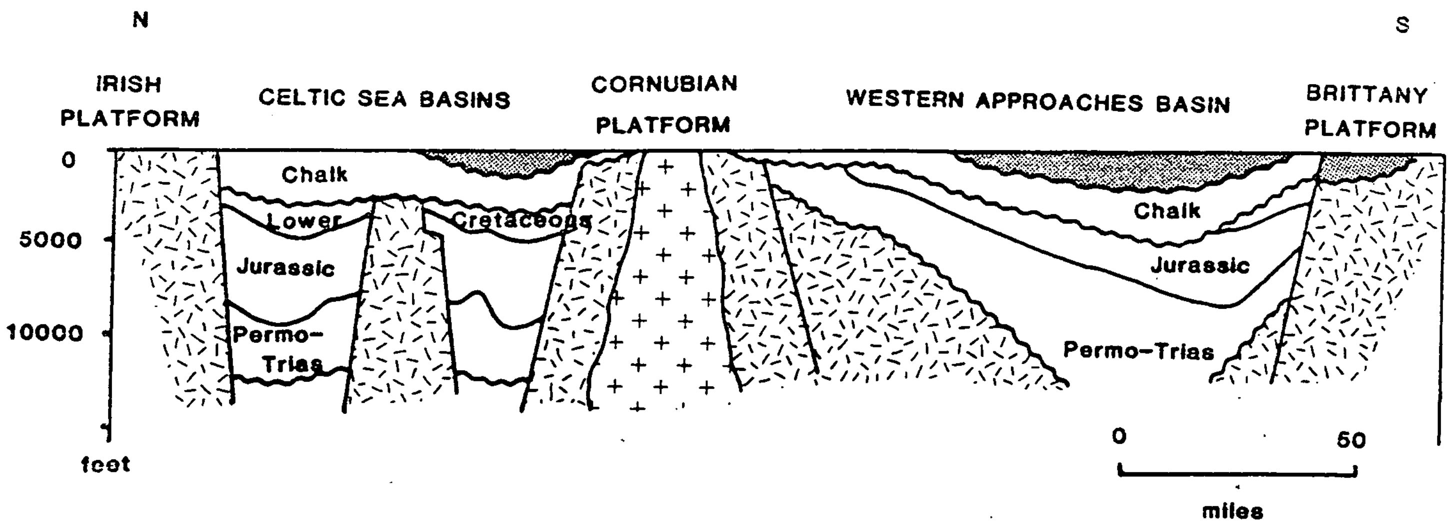

Fig. 1.5 Interpretative sections across the Celtic Sea and Western Approaches Basin, after Naylor and Mounteney

a) Northern Basins

In general, the shelf basins west of Shetland and Scotland are markedly asymmetric. Included are basins not shown here such as the Rathlin, Sule Sgeirr, Fair Isle and Inner Hebrides Basins. Bounding fault trends are NE-SW or NNE-SSW due to Caledonide control (Fig. 1.6). For commercially explored basins, crustal stretching is known to have started in the Devonian or Permo-Triassic with the deposition of continental sediments in grabens (Naylor and Mounteney 1975). Throughout this province there was a major break in sedimentary deposition during the Late Jurassic and Early Cretaceous, and other than to the west of Shetland, Tertiary sediments are thin or absent. This area was extensively affected by the Thulean igneous episode; volcanics are found on the shelf e.g. the Upper Cretaceous centre of Blackstones, and the Tertiary centres of the Inner Hebrides, and are even more widespread west of the shelf in the Faeroe-Shetland Basin, on the Faeroes continental block, and in the Rockall Trough (refer to later sections).

b) Southern Basins

The shelf basins to the southwest of Ireland and the UK are in general more symmetric, have a thicker and more extensive Tertiary sedimentary succession, and different structural trends compared with the northern basins. The Celtic Sea Troughs, Channel Basin, and Western Approaches Basin are controlled by ENE-WSW trending faults, as are the troughs of the Grand Banks area (Naylor and Mounteney, op. cit.) reflecting the underlying Hercynian trend. On the other hand, the trends of the Fastnet Basin, which is southwest of the Celtic Sea Troughs and the closest shelf basin to the study area, are NE-SW, with a secondary set of NW-SE strike-slip faults. Robinson et al (1981) suggest that this was due to the development in this region of a dextral simple shear regime in Jurassic to Cretaceous time. They also report the near-absence of Middle to Late Jurassic sediments in the Fastnet Basin, which may be due either to non-deposition or to extensive erosion at the late Kimmerian unconformity of lowermost Cretaceous age. The earliest igneous activity in the Fastnet Basin, which is controlled by the NW-SE faulting, corresponds in time with

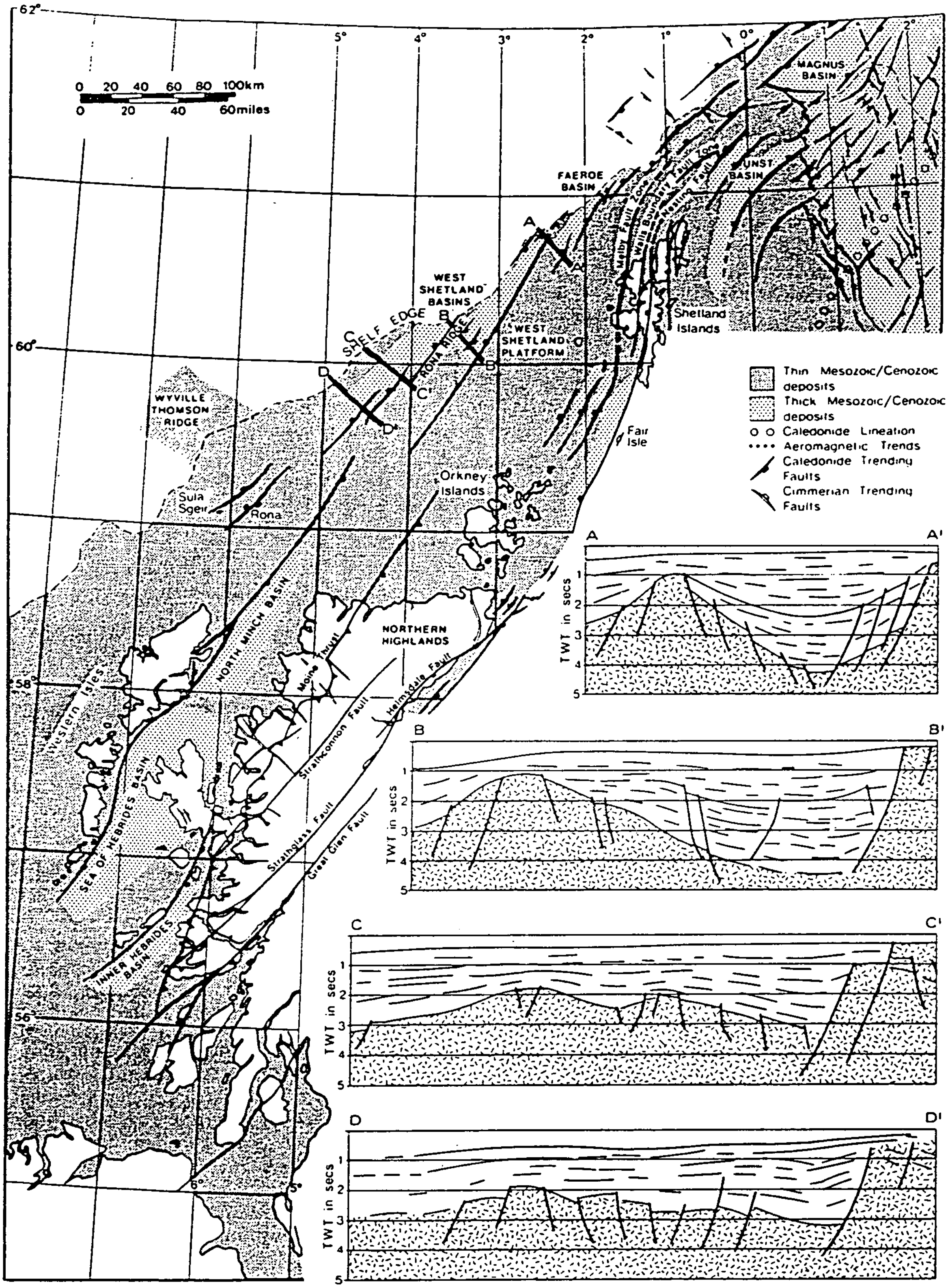

Fig. 1.6 Caledonide lineations and half-graben development in the West of Shetland area, from Johnson and Dingwall, 1981

the hiatus in sedimentation. Later igneous activity is of (post) Upper Cretaceous age, in the form of sills and plugs (Caston et al 1981).

1.1.2 THE FAEROE BASIN

The Faeroe Basin and northern Rockall Trough are colinear and the development of the Faeroe Basin and the Norwegian margin is probably closely linked to that of Rockall Trough. The Faeroe-Shetland Basin is separated from Rockall Trough by the NW-SE-trending Wyville-Thomson and Ymir Ridges, which may be youthful features (Ridd, in press). Seismic sections show these ridges to be anticlinal structures at the level of the Lower Tertiary (Late Paleocene and Early Eocene) volcanics which occur over practically all of the northern Rockall Trough (Smythe et al, 1978), the Faeroe Basin (Ridd, op. cit.), and the Faeroes Platform. Fig. 1.7. shows one section across the Ymir and Wyville-Thomson Ridges.

Little is known directly about the Faeroe Basin; the Basin has only been drilled in the east adjacent to Rona Ridge where it is known that Mesozoic sediments lie beneath the Tertiary volcanics The Tertiary volcanics here are tuffs (Ridd, op. cit.). It is likely that a reflector which defines the top of a sedimentary succession above the volcanics is equivalent to Roberts (1975) R4 horizon, and there is a thick Tertiary succession in the Faeroe Basin (Fig. 1.8): evidence of Tertiary subsidence. Approximately along the axis of the Faeroe Basin there is a zone where the reflection from the top of the volcanics becomes arched in places, and Ridd interprets this area as a possible chain of Early Tertiary igneous centres (Ridd, op. cit.). Each is up to 50 km wide, and they are seen singly or in pairs in any one seismic section. Fig. 1.8 summarises ideas on the <u>possible</u> geology of the Faeroe Basin.

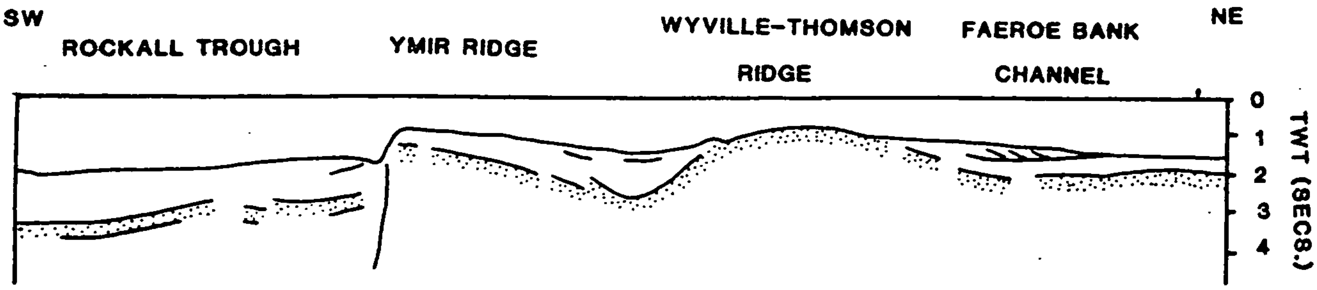

Fig. 1.7 Section across the Ymir Ridge and Wyville Thomson Ridge interpreted from seismic profiles, from Ridd, 1981. The shading shows rocks beneath the top of the Lower Tertiary volcanic reflector

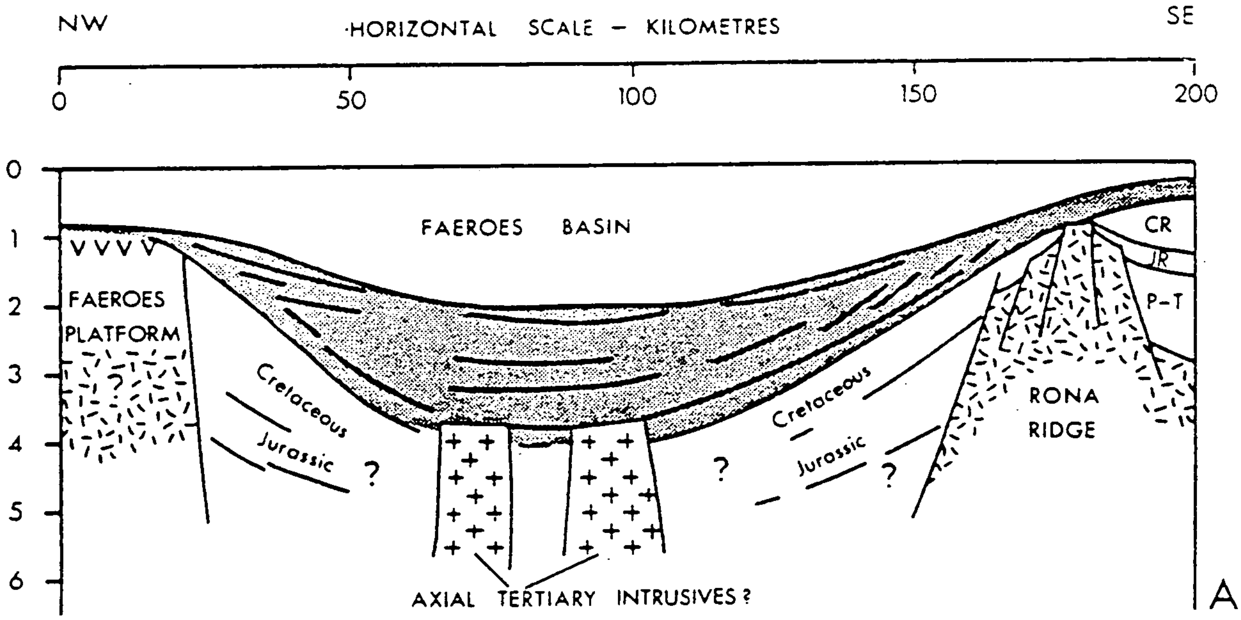

Fig. 1.8 Largely hypothetical cross-section of the Faeroes Basin, From Stow, 1981

1.1.3 ROCKALL BANK TO GOBAN SPUR

The areas from Rockall Bank to Goban Spur are discussed in more detail, as they are more relevant. Fig. 1.9 shows the positions of all available seismic refraction experiments and crustal models in this area and the seismic refraction results are summarised in Fig. 1.10, 1.11, and 1.12. Where depths to the Moho are known, these give values appropriate for continental crust on Rockall Bank, Hatton-Rockall Basin, Porcupine Bank, and the Celtic and Hebridean Shelves. The refraction data from Rockall Trough and Porcupine Seabight however are open to various interpretations. The ambiguity of using crustal thickness alone to distinguish continental from oceanic crust is discussed in Appendix A.

a) PORCUPINE BANK AND PORCUPINE SEABIGHT

Porcupine Seabight separates all but the northernmost end of Porcupine Bank from the remainder of the Irish continental shelf. Whitmarsh et al's (1974) long refraction experiment on Porcupine Bank established its continental nature, although it was not possible to differentiate between a layered and unlayered crust (Fig. 1.13).

Sediment cover on the Bank is thin, and absent on the shallowest area. There is little direct information on the basement geology of the Bank, although it is generally accepted that the north of the Bank is a continuation of the Caledonian Province of Northern Ireland and Scotland. Fig. 1.14 shows two published versions of the inferred geology.

Porcupine Seabight is in its northern region an almost N-S trending basin with a thick sedimentary succession. Its northern termination at approximately 53°N is an ENE-WSW trending discontinuity against which the Seabight's Mesozoic-Recent succession thins considerably (Bailey, 1979). To the northeast of this discontinuity is the Slyne Trough, a complex of three NE-SW trending grabens disposed en echelon against Caledonoid faults (Bailey op. cit). As in the other shelf basins, the thin Cenozoic cover oversteps these grabens' boundary faults without deformation. Fig. 1.15 shows three

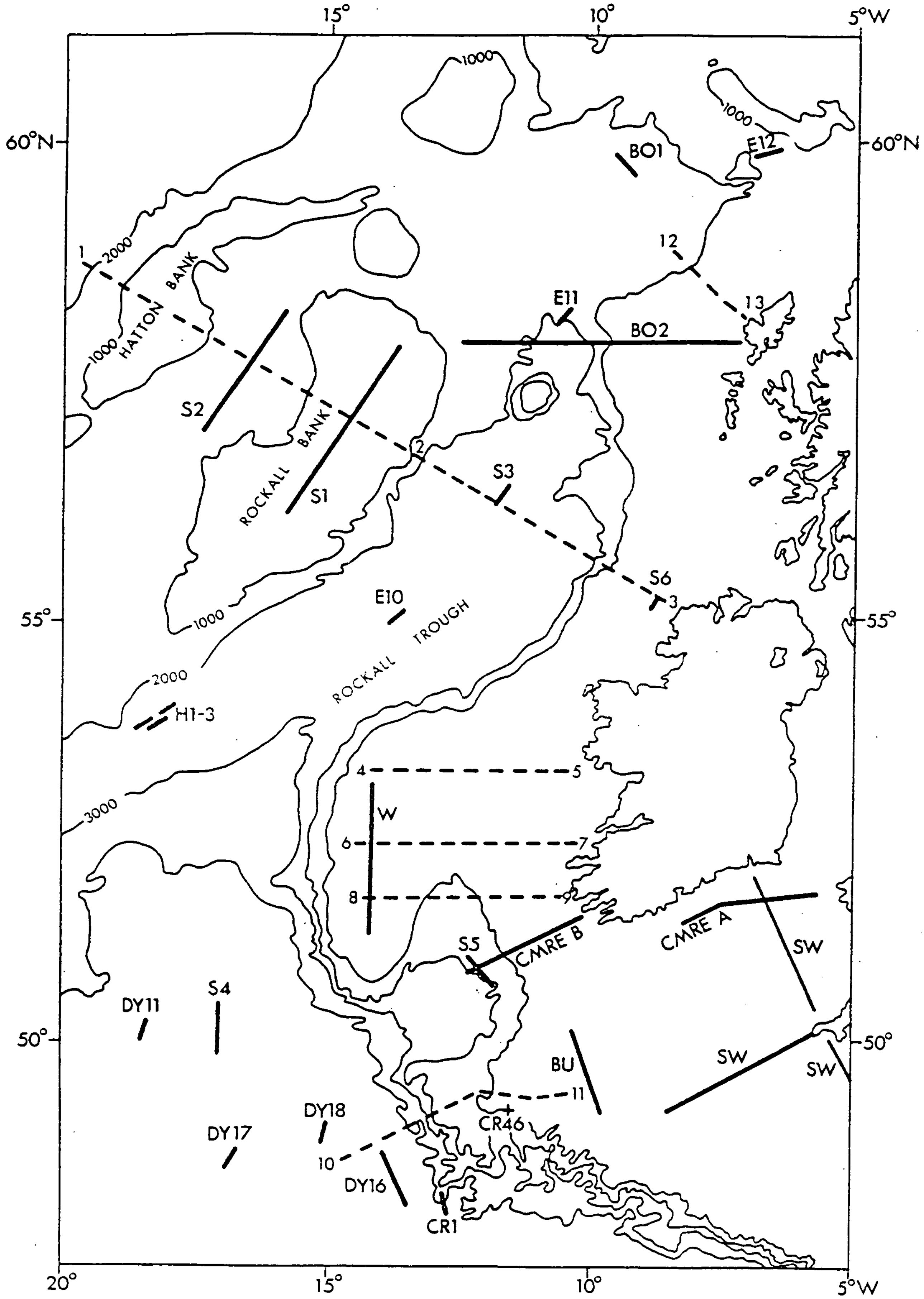

Fig. 1.9 Positions of seismic refraction experiments and crustal models west of the UK and Ireland

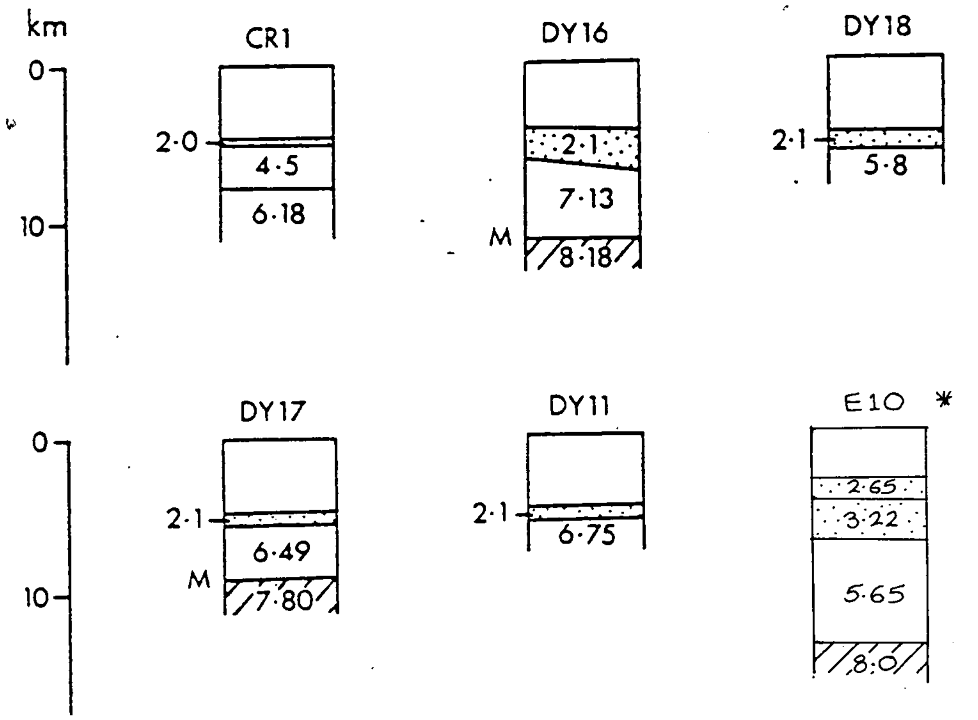

Fig. 1.10 Figs. 1.10, 1.11 and 1.12 tabulate the results of seismic refraction experiments whose positions are given in Fig. 1.9. Seismic velocities are in km/sec. Many velocities are assumed velocities, and reference should be made to the original publications which are as follows:

E 12, E 11, E 10	: Ewing and Ewing 1959 *
BO 1, BO 2	: Bott et al, 1979
S 1, S 2, S 3, S 6	: Scrutton, 1972
H 1-3	: Hill, 1952
W	: Whitmarsh et al, 1974
S 4, S5	: Scrutton, Stacey and Gray, 1971
CMRE A	: Bamford, 1972
CMRE B	: Handley, 1971
SW	: Bott et al, 1970
BU	: Bunce et al, 1964
CR 46, CR 1	: Gaskell et al, 1958
DY 11, DY 17, DY 18, DY 16	: Hill and Laughton, 1954

* E10, E11 reinterpreted by Jones et al., 1970

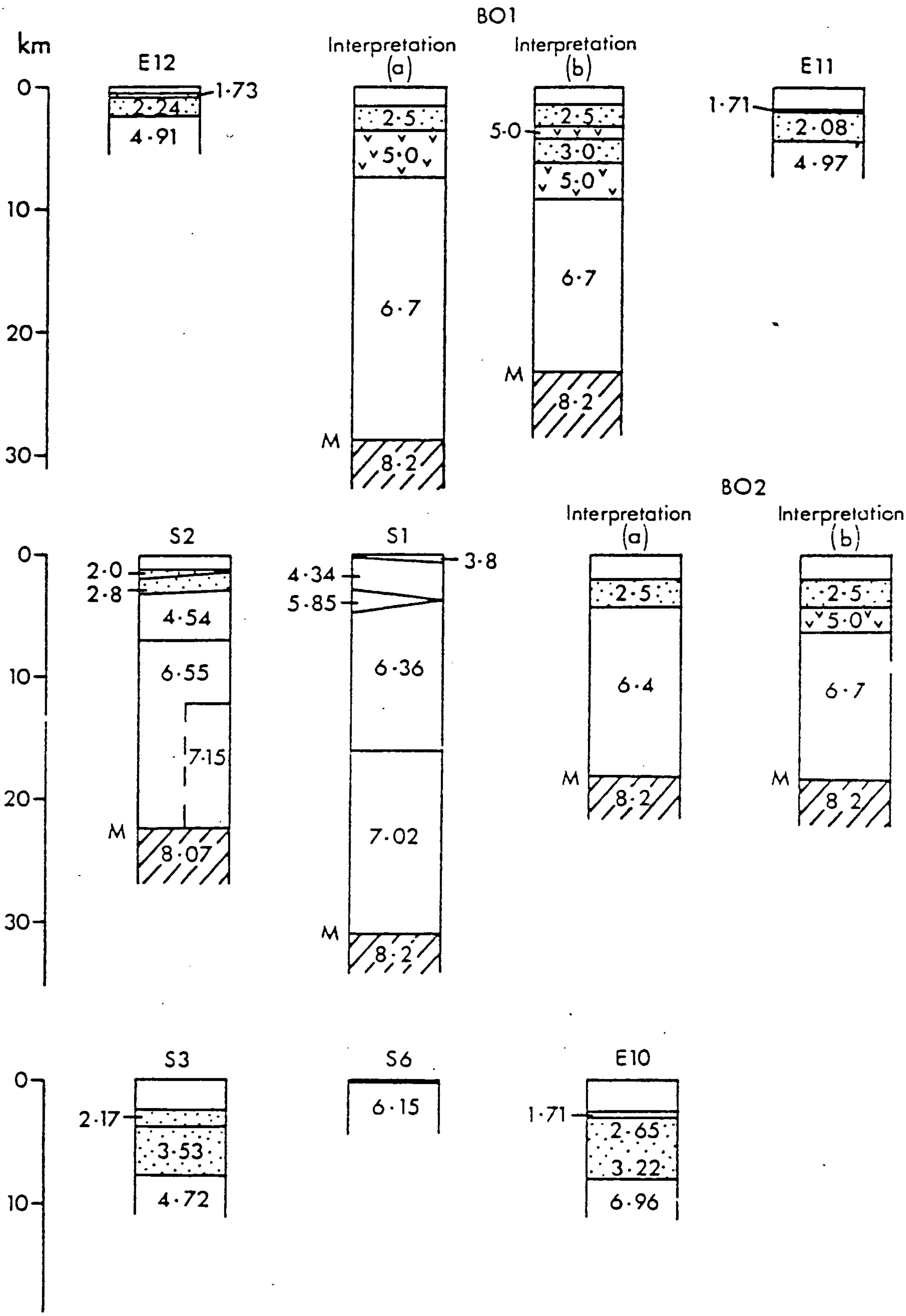
km
0
10
20
30
E12
1·73
2·24
4·91
BO1
Interpretation (a)
2·5
5·0
6·7
M
8·2
Interpretation (b)
2·5
5·0
3·0
5·0
6·7
M
8·2
E11
1·71
2·08
4·97
S2
2·0
2·8
4·54
6·55
7·15
M
8·07
S1
3·8
4·34
5·85
6·36
7·02
M
8·2
BO2
Interpretation (a)
2·5
6·4
M
8·2
Interpretation (b)
2·5
5·0
6·7
M
8·2
S3
2·17
3·53
4·72
S6
6·15
E10
1·71
2·65
3·22
6·96

Fig. 1.11

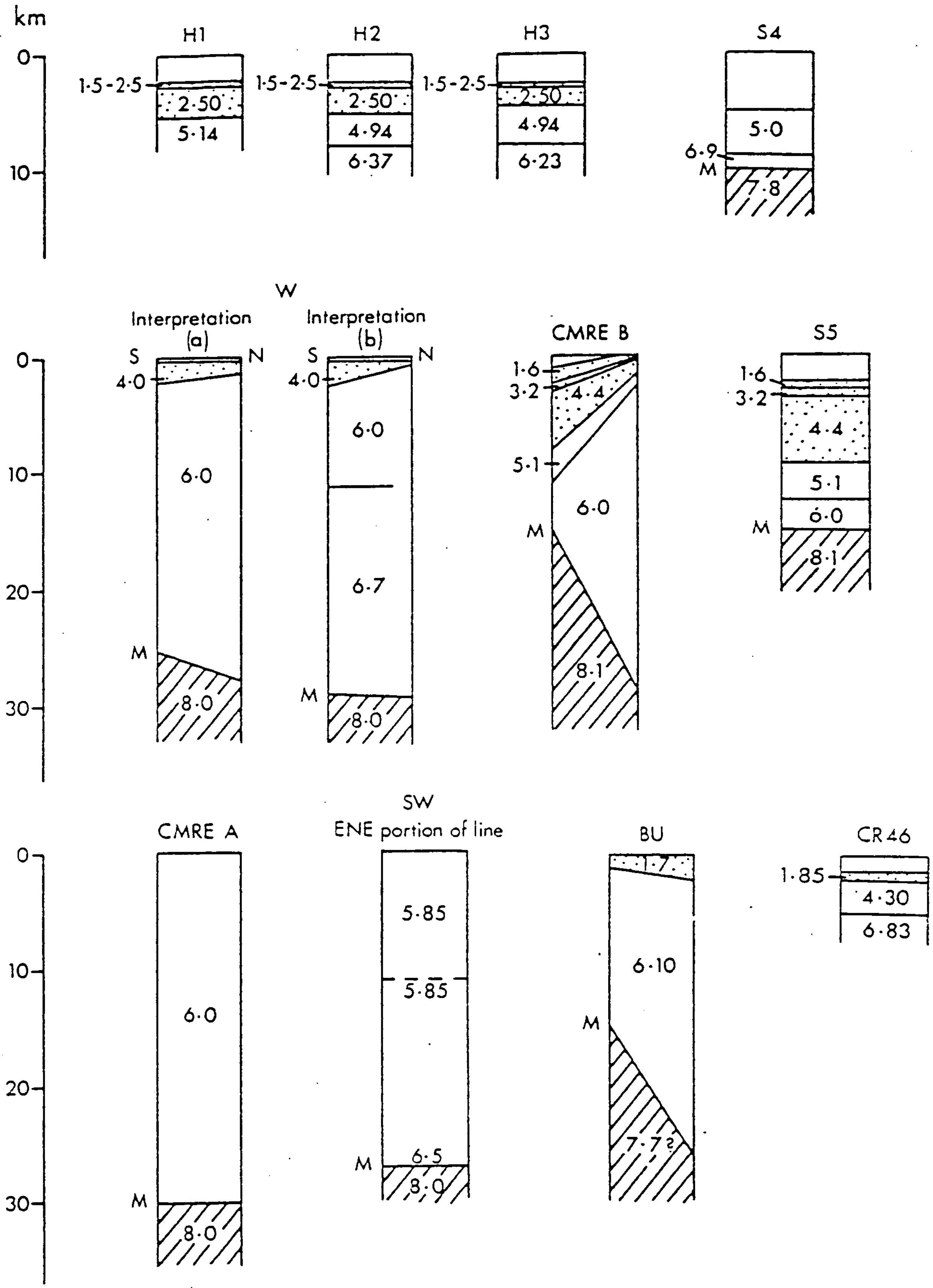
km
H1
H2
H3
S4
1·5-2·5
2·50
5·14
4·94
6·37
6·23
5·0
6·9
M
7·8
W
Interpretation (a)
Interpretation (b)
CMRE B
S5
S
N
4·0
6·0
6·7
M
8·0
1·6
3·2
4·4
5·1
8·1
SW
ENE portion of line
CMRE A
BU
CR46
5·85
6·5
1·7
6·10
7·7?
1·85
4·30
6·83

Fig. 1.12

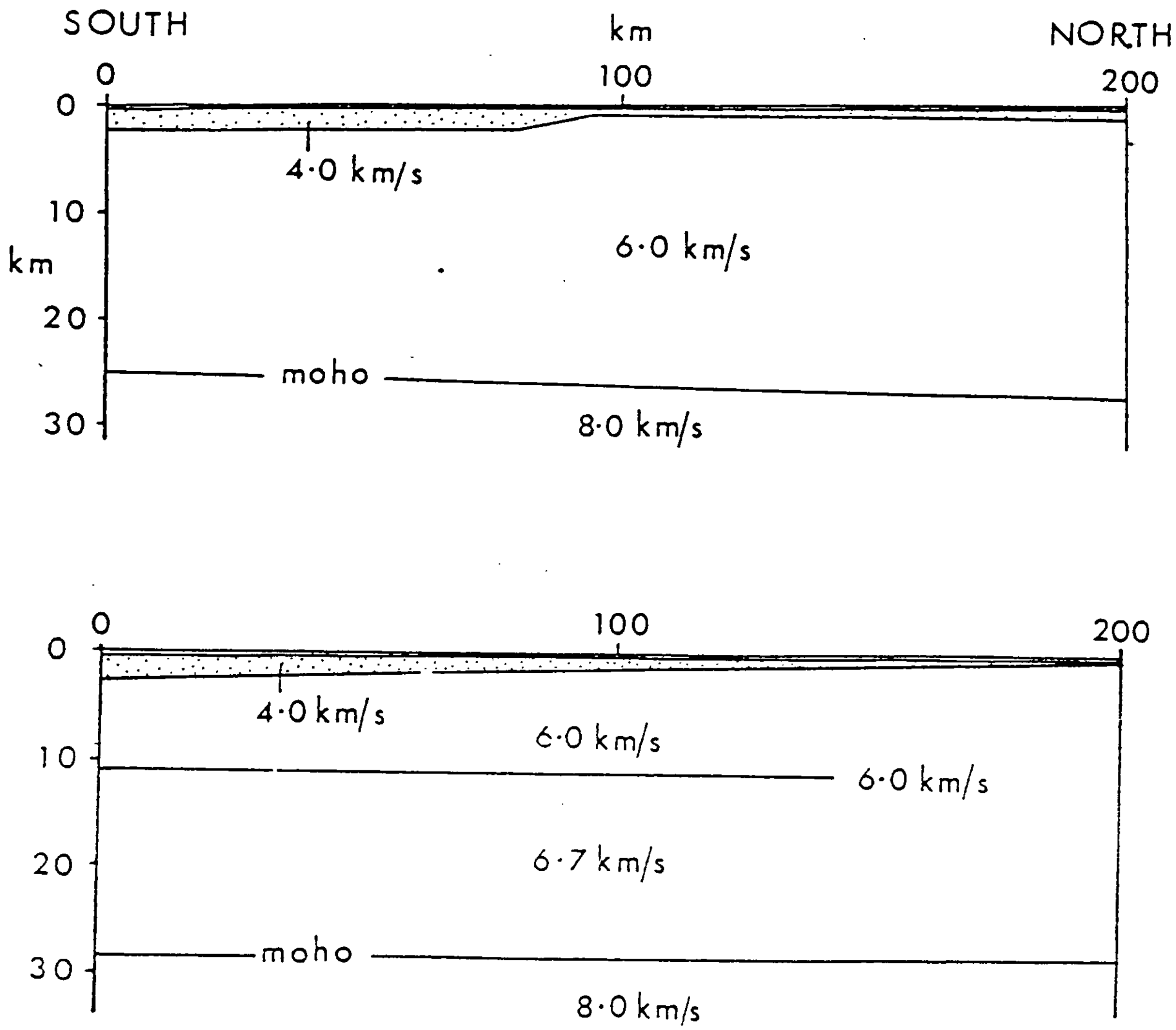

Fig. 1.13 Two models of the crustal structure beneath Porcupine Bank, from Whitmarsh et al, 1974. The velocity of 4.0 km/s assigned to the sediments is an assumed value.

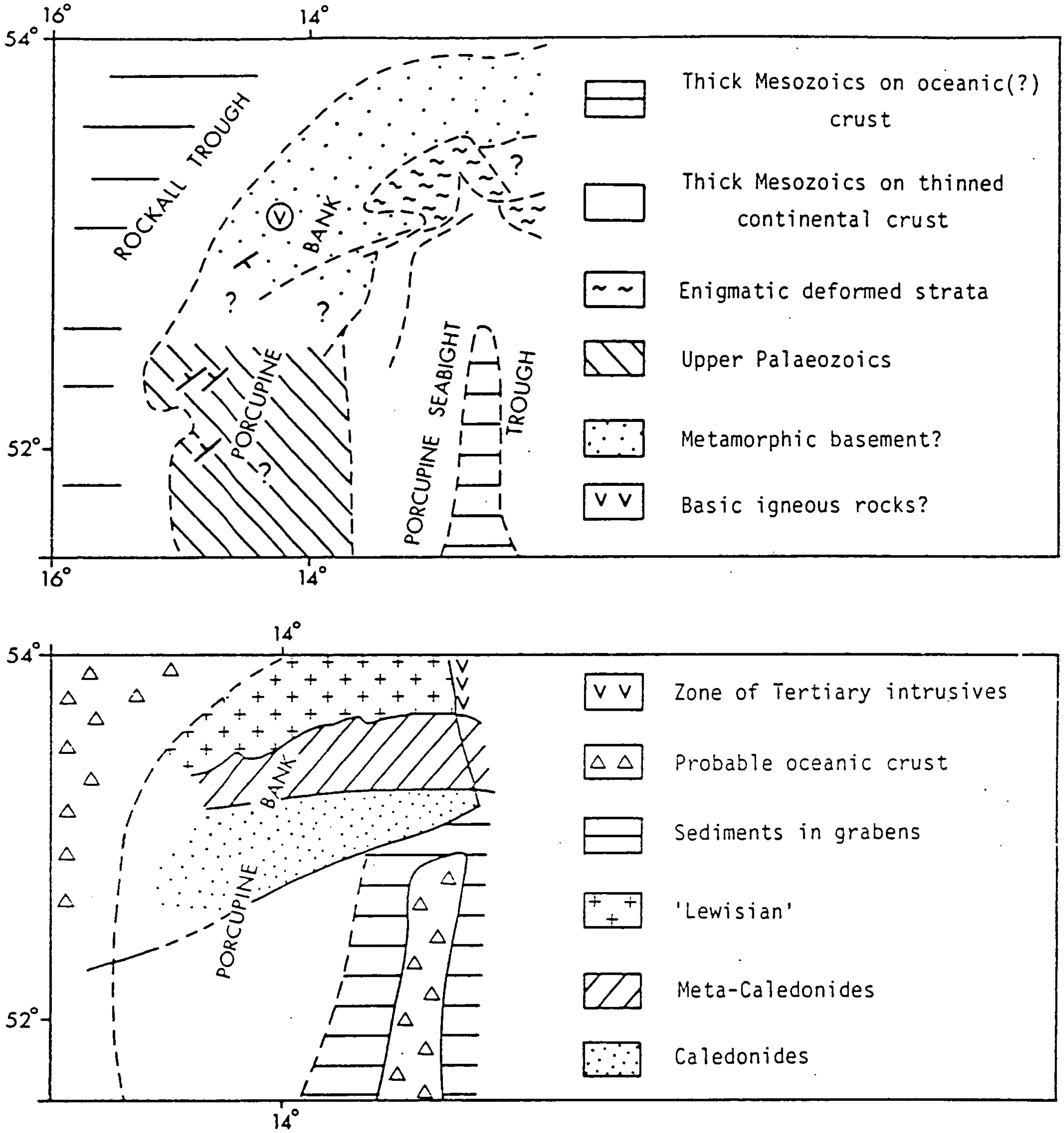

Fig. 1.14 Two versions of the inferred geology of Porcupine Bank and adjacent areas.

Above : Bailey, 1981

Below : Riddihough and Max, 1976

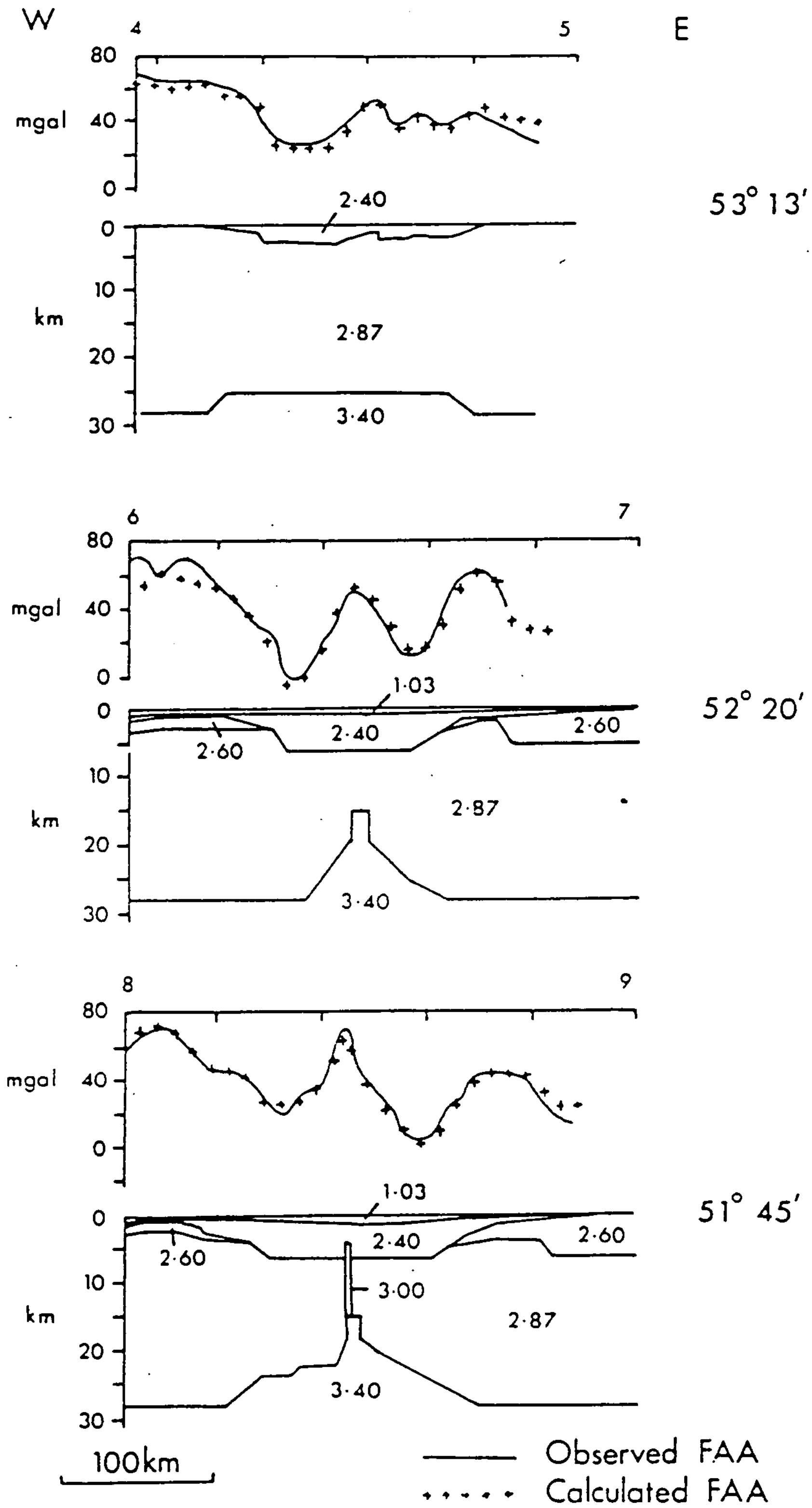

Fig. 1.15 Three east-west gravity models across Porcupine Seabight Trough, from Buckley and Bailey, 1974. Positions of the models are shown in Fig. 1.9. Density values are in g/cm³.

east-west crustal models across northern Porcupine Seabight; Buckley and Bailey (1975) assume progressive thinning of the crust from north to south.[1] Most academic workers (e.g. Bailey, 1979; Riddihough and Max, 1976) have assumed that Porcupine Seabight is underlain by oceanic crust, but the published section across the northern Seabight in Naylor and Mounteney (op. cit.), which is based on commercial data, shows continuous continental crust overlain by three seconds (TWT) of Permian and post-Permian sediments (Fig. 1.16). In contrast to the Fastnet Basin, sedimentation in Porcupine Seabight has been almost continuous from the Bathonian/Bajocian until the late Kimmerian erosional event (Caston et al, 1981).

Within southern Porcupine Seabight there are two refraction lines, the results of which are shown in Fig. 1.17. It can be seen that in southern Porcupine Seabight, water depths are greater than in the north and the crust thinner, but the sedimentary succession probably of the same thickness. The crust is layered, and of a lower average velocity than Porcupine Bank. The velocity of the 4.4 km/s sediment is exactly that found by Jones (1978) at >1 km depth on the margin of Rockall Trough, for what he assumed were Permo-Triassic sediments.

A chart of the structure of the mouth of Porcupine Seabight has been produced by Roberts et al (1981); this will be discussed in a later chapter in conjunction with new data. Published hypotheses on the evolution of Porcupine Seabight and Rockall Trough are presented towards the end of this chapter.

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[1] There is no refraction data for this area, and the gravity profiles could alternatively be interpreted as the result of high density intrusive bodies within the crust (cf. Baker and Wohlenberg, 1971).

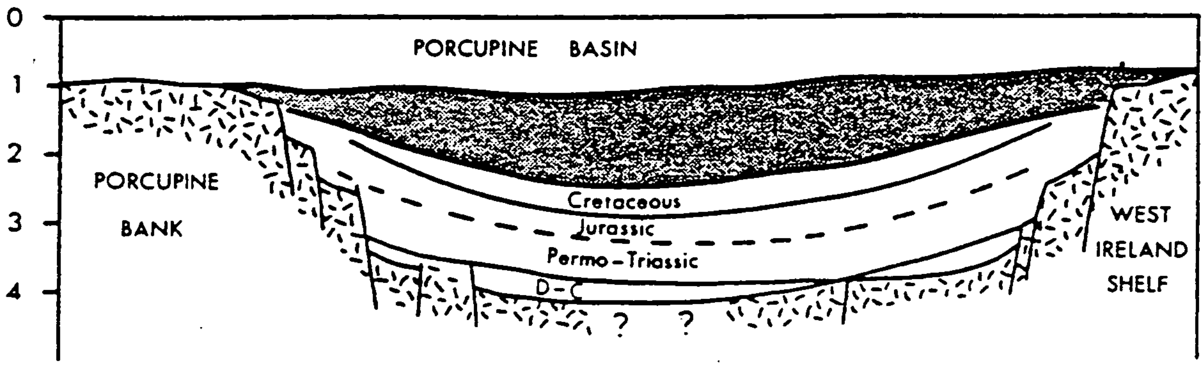

Fig. 1.16 Cross-section across the Porcupine Basin from Stow, 1981

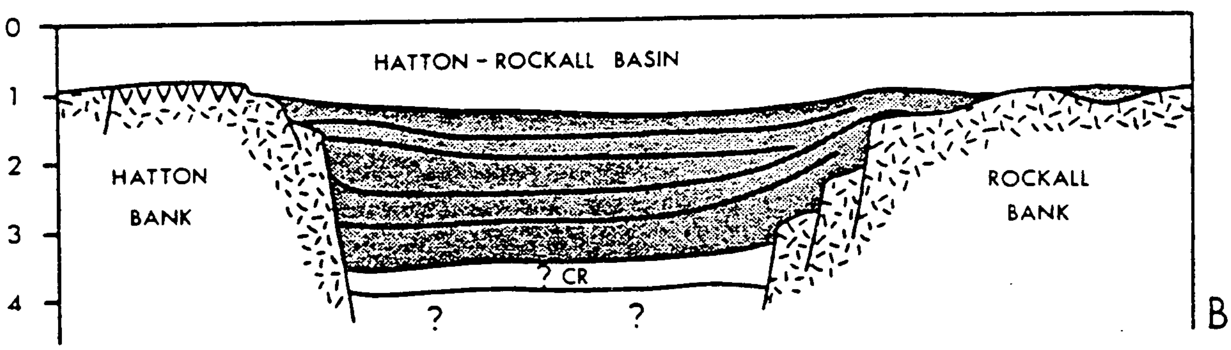

Fig. 1.20 Interpretative cross-section through Hatton-Rockall Basin; Stow, 1981

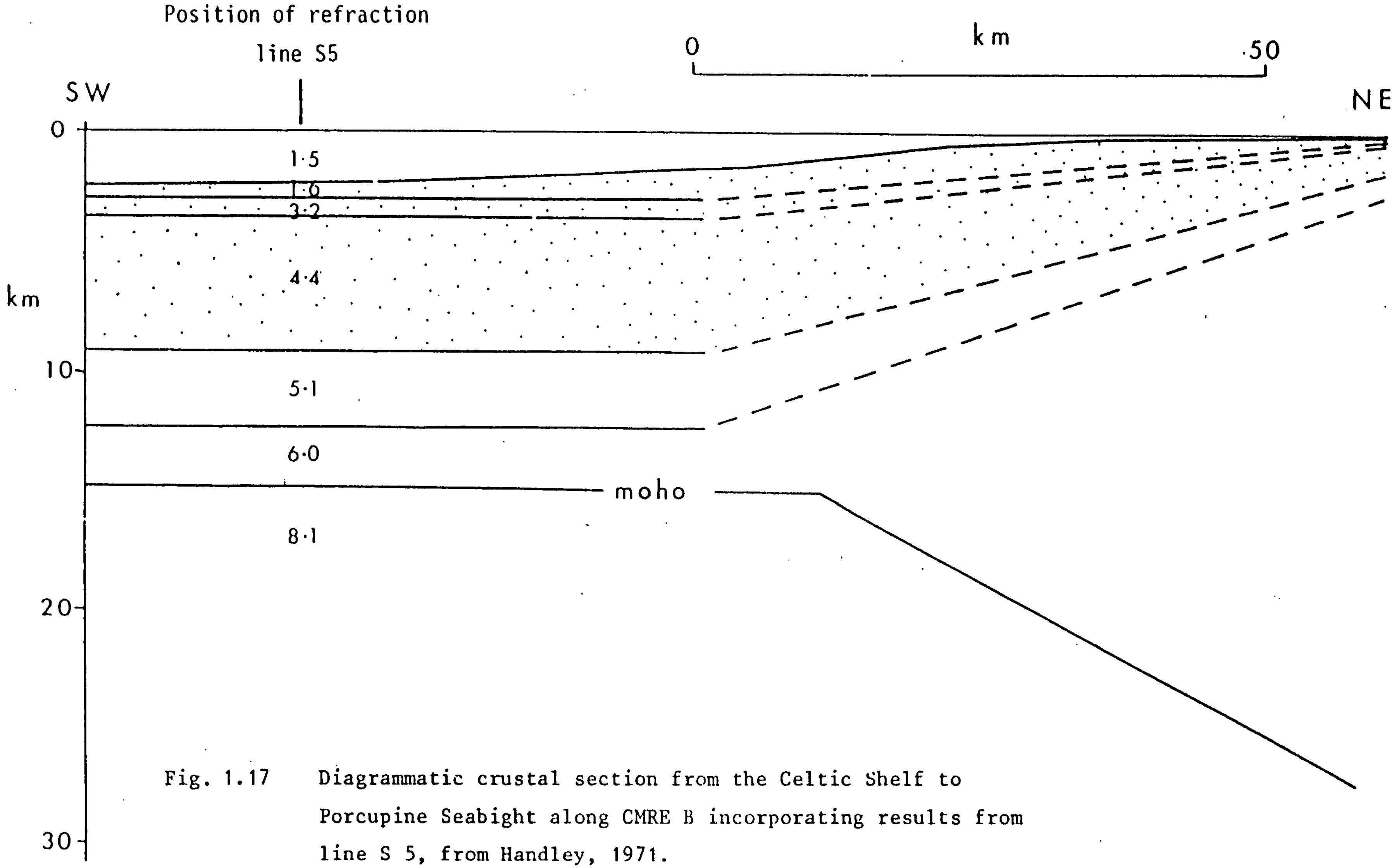

Fig. 1.17 Diagrammatic crustal section from the Celtic Shelf to Porcupine Seabight along CMRE B incorporating results from line S 5, from Handley, 1971.
Velocities are in km/s.

b) PORCUPINE ABYSSAL PLAIN

Porcupine Abyssal Plain is largely underlain by Cretaceous oceanic crust. The Charlie Gibbs fracture zone, the largest offset North Atlantic fracture zone, defines the southern end of Rockall Trough and provides a useful constraint for pre-drift reconstructions. Cherkis (1973) has mapped the eastern end of the Charlie Gibbs fracture zone as far as 17°W. Fig. 1.18 shows the most salient features of the ocean crust west of Porcupine Bank and Rockall Plateau. The Clare Lineament is a feature first described in this thesis.

Anomaly 32 (34 in Kristofferson's (1978) terminology) is the oldest universally recognised anomaly off Porcupine Bank (Scrutton et al, 1971), which is of Campanian age (72 m.y.) (Hailwood, 1979). Srivastava (1978) and Roberts et al (1981) recognise two older anomalies; 33 and 34, which Kristofferson (1978) believes are "noise" due to their linear discontinuity.

The continent-ocean boundary has been defined west of Goban Spur by Scrutton (1979); this thesis discusses the position of the continent-ocean boundary off Porcupine Bank and Porcupine Seabight.

c) GOBAN SPUR

The Spur is an interesting area of continental crust at "intermediate" water depths, situated at the pronounced change in trend between the Biscay spreading and the spreading in the Porcupine Bank - Orphan Knoll sector; and is presumably affected by both tensional regimes.

This thesis is only periphally concerned with Goban Spur. References to the geophysics of Goban Spur can be found in Scrutton (1979) and to the geology in Dingle and Scrutton (1979), and Roberts et al (1981). Fig. 1.19 shows two gravity models from Goban Spur to Porcupine Abyssal Plain.

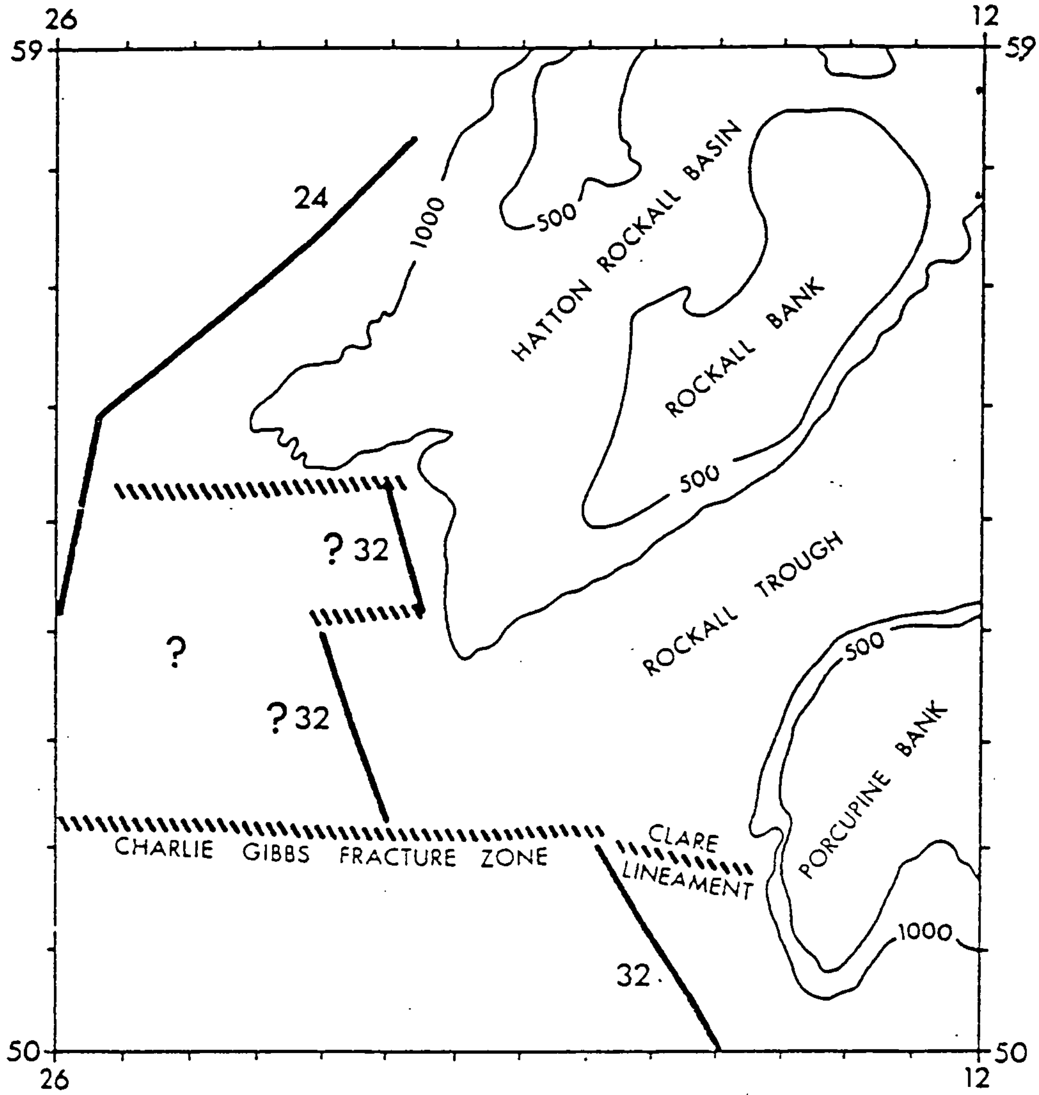

Fig. 1.18 Oldest identified sea-floor spreading anomalies in the area adjacent to Rockall Plateau and Porcupine Bank (heavy lines) and associated fracture zones and related features (hachured lines). Anomaly 32 has been reidentified as anomaly 34 by some authors (Kristoffersen, 1978). Bathymetry is in fathoms.

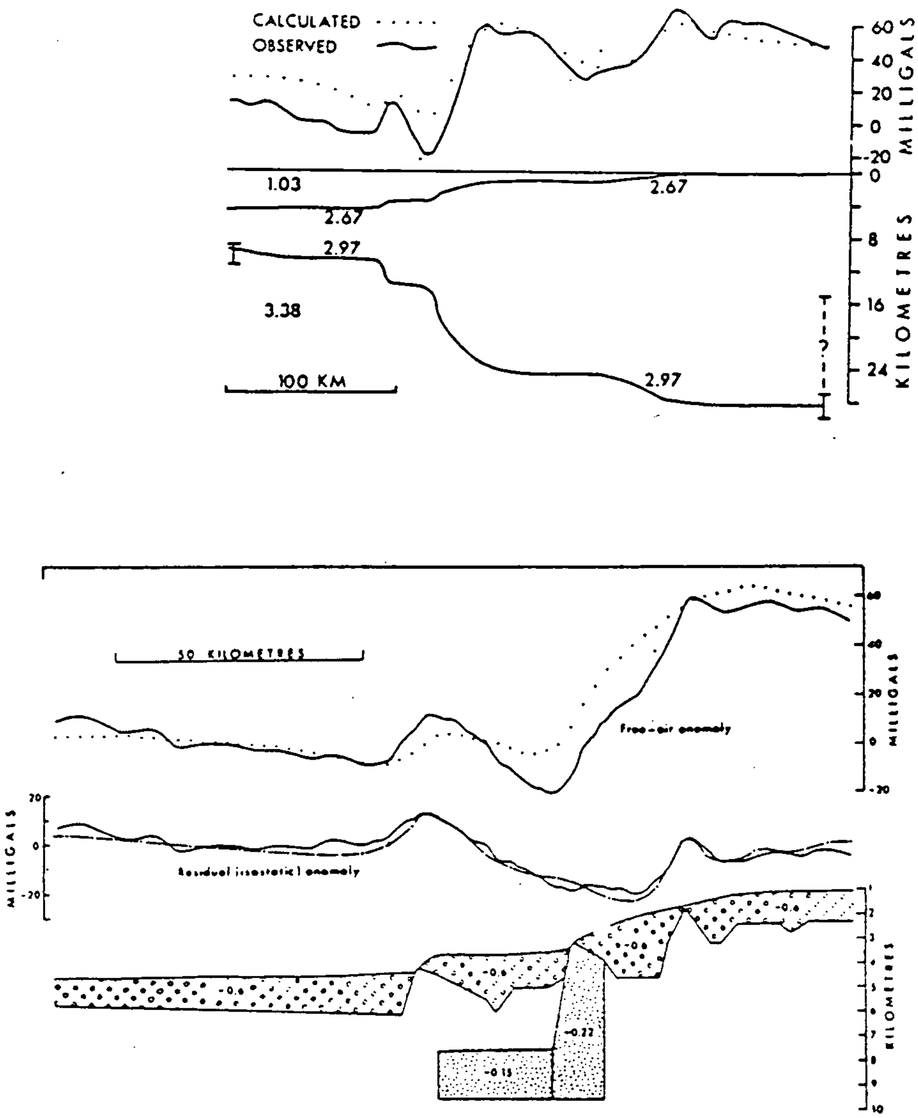

Fig. 1.19 Gravity models across Goban Spur, Scrutton (1979)

Above : An Airy-type model of the crust with densities in g/cm³. The vertical bars show ranges of depths to the Moho determined from refraction experiments.

Below : A detailed gravity model. Density contrasts in g/cm³.

Upper Section; solid line - observed free air anomaly
dotted line - calculated anomaly due to modified Airy model.

Middle Section, the chained line that matches the isostatic anomaly (solid line) is calculated from the density model below.

Profile position 10---11 on fig. 1.9

Dingle and Scrutton (1979) surmise that the grabens on Goban Spur may contain sediments of Jurassic age. They suggest that faulting was over by Albian times. Most faults are parallel to the continent-ocean boundary trend of NNW-SSE or NW-SE.

d) ROCKALL PLATEAU

Rockall Plateau can be divided into three main units; the central Hatton-Rockall Basin (Fig. 1.20), Rockall Bank to the east, and Hatton Bank to the west, with smaller banks and basins to the south and north (Roberts, 1975). Although this thesis is not directly concerned with new data from Rockall Plateau (as with Goban Spur), six DSDP holes have been drilled on the plateau (see Fig. 1.1) which provides a useful regional framework. A summary of the geology and geophysics of Rockall Plateau is found in Roberts (1975), and in the Initial Reports of the Deep Sea Drilling Project, volume 48.

A structural framework for Rockall Plateau is given by Scrutton (1972) in a crustal model across Rockall Plateau using seismic refraction, seismic reflection and gravity data (refer to Fig. 1.21)[1] which suggests that the whole plateau is continental. Further evidence of its continental nature is isotopic analysis of the Tertiary granite of Rockall Island (Moorbath and Welke, 1969), and sampling of Laxfordian and Grenvillian granulites on Rockall Bank (Roberts et al, 1972; Roberts et al, 1973a). Upper Cretaceous and Lower Tertiary igneous rocks are probably widespread on the plateau (Laughton et al, 1972; Naylor and Mounteney, 1975).

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[1]The refraction line S1 (Fig. 1.9) on Rockall Bank has been reinterpreted by Bunch (1979) using synthetic seismogram analysis; the differences in velocity structure between the results of Scrutton (1972) and Bunch (1979) are small however. The velocity structure of lines S1 and S2 show higher velocities than for 'normal' continental crust (Scrutton op. cit.). The differences in velocity between lines S1 and S2 (Rockall Bank and Hatton-Rockall Basin) could be due to a change in geology, or could conceivably be a result of the processes which caused the subsidence of Hatton-Rockall Basin.

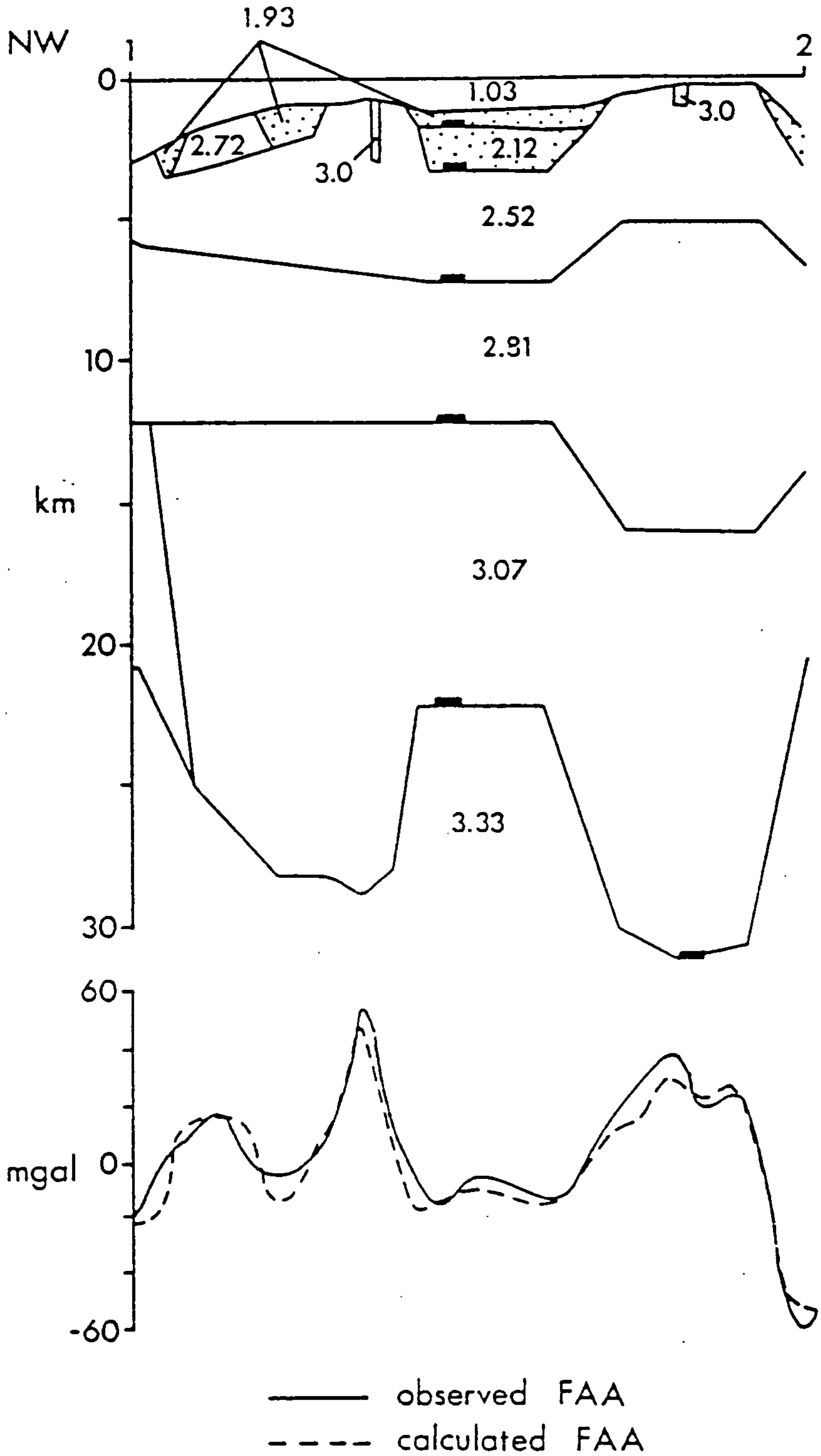

Fig. 1.21 A crustal model of Rockall Plateau, and comparison of observed and calculated gravity (Scrutton, 1972). The position of the profile is between numerals 1 and 2 in Fig. 1.9. Very thick lines represent seismic refraction control.

There is a rectangular deep-water area to the southwest of Rockall Plateau where the presence of oceanic crust is disputed (see Fig. 1.18). The age of spreading here, if any, is important in the reconstruction of early ocean formation west of Britain. Anomalies 32-34 have been recognised here by Roberts (1975), but Howarth (1980) is not sure that the evidence is convincing and correlates these anomalies with an anomaly sequence over continental crust off Newfoundland.

e) ROCKALL TROUGH

There is little data on the geology and geophysics of Rockall Trough that is not open to various interpretations. No deep wells have been drilled, and all estimates of the ages of the fairly thick sedimentary sequence within Rockall Trough are based on extrapolation from DSDP holes (Fig. 1.1) drilled in nearby shelf areas (such as Rockall Plateau) or tectonically and topographically similar areas (the northern Bay of Biscay). The seismic refraction data too is limited. Experiment BO2 (Bott et al, 1979) in northern Rockall Trough (Fig. 1.9) has given invalid depths to the basement and Moho as it received first arrivals from the top of the Tertiary lavas ubiquitous in this area (as already mentioned) and not from true basement. The two interpretations of line BO1 (Fig. 1.9) also make (unavoidable) assumptions on thickness of lava sequences and sediments, which result in Moho depths being poorly determined. (Reinterpretations of these two lines with data on basalt and sediment thickness from commercial seismic lines would give valid Moho depths). The only other seismic refraction experiment to give information on Moho depths was line E10 in central Rockall Trough as reinterpreted by Jones et al (1970) (Fig. 1.9 and 1.12). A crustal thickness of about 7 km was found and a Moho depth of 14 km, which agrees reasonably well with Scrutton's (1972) crustal model across the Trough which was based mainly on gravity data (Fig. 1.22). Geological data on Rockall Trough obtained before 1975 are given in Roberts (1975), in which a stratigraphy of Rockall Trough is given, based on single-channel seismic reflection profiles. The age of the deeper sediments is not known, neither can the basement geology be ascertained from the reflection profiles, although Roberts (1975)

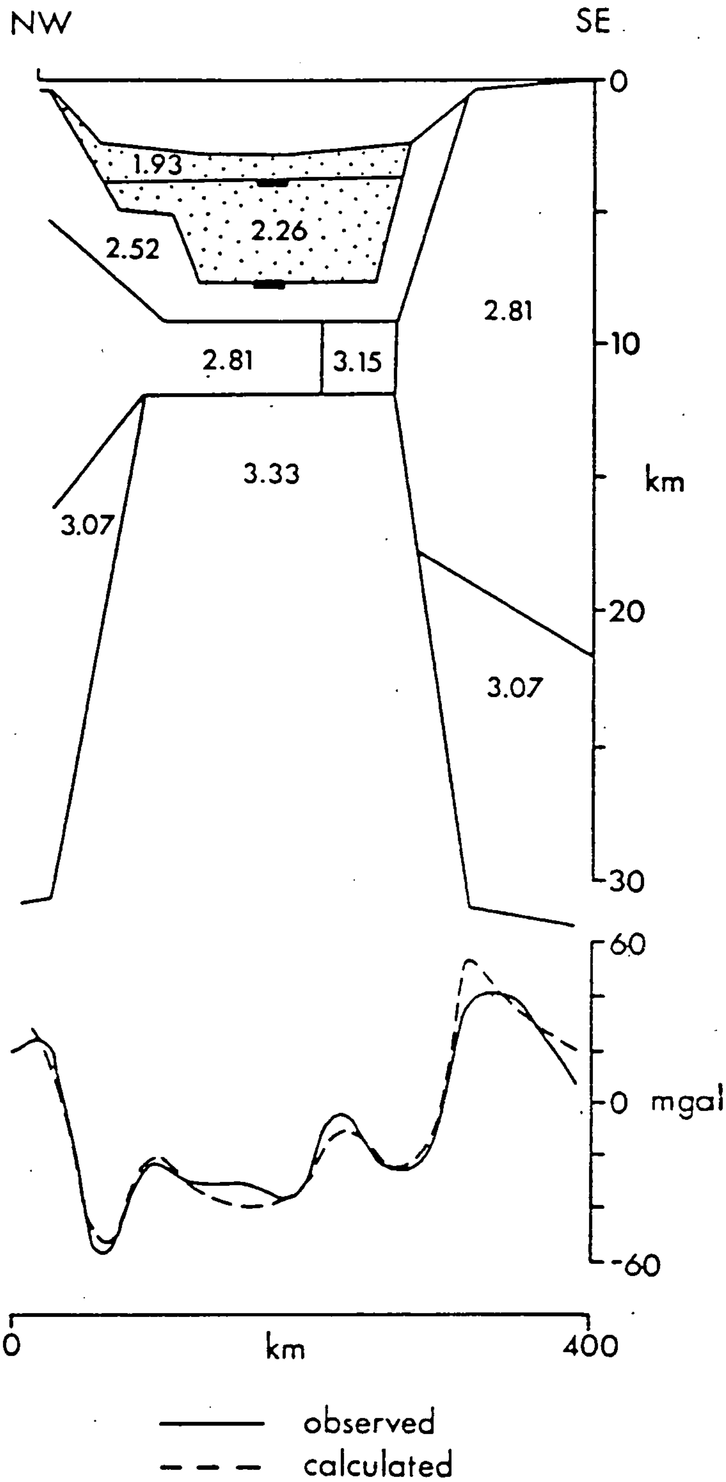

Fig. 1.22 Crustal model across Rockall Trough, Scrutton (1972). (Profile located between numerals 2 and 3 in Fig. 1.9). Figures are densities, in g/cm³, and very heavy lines represent seismic refraction control.

assumes that most of the basement within the Trough is oceanic.

An important sediment feature within the Trough is the Feni Ridge, a sediment ridge on the west of the Trough which is covered with sediment waves (Roberts and Kidd, 1979; Flood et al, 1979). Jones et al (1970) attributed this to sediment distribution controlled by contour currents from the Norwegian Sea after the early Tertiary. Roberts et al's (1981) revised stratigraphy and Dingle, Megson and Scrutton's stratigraphy (in press) for Rockall Trough is discussed in the next chapter.

In addition to extensive volcanic horizons, northern Rockall Trough is known to contain large igneous centres which appear as seamounts; Anton-Dohrn Seamount, Rosemary Bank and the Hebrides Terrace Seamount. Alkali basalts have been dredged from Anton-Dohrn Knoll (Jones et al. , 1974) which unfortunately were too altered to date. Jones and Ramsay suggest that the basalts are of Upper Cretaceous age as the (unbaked) sediments intimately associated with the volcanics are Maastrichtian chalk. Russell and Smythe (1978) on the contrary interpret Anton-Dohrn Knoll, Rosemary Bank, and the igneous centres in the Faeroe Basin as being of Permian age, on the basis of magnetic vector modelling. A recent paper by Miles and Roberts (1981) reinterprets the Rosemary Bank magnetic data to give an Upper Cretaceous-Tertiary age.

1.2 THE CANADIAN MARGIN AND TRANS-ATLANTIC CORRELATIONS

The Canadian margin which is placed by all Atlantic refits adjacent to the margin to the south of Rockall is that which is NE of Newfoundland, and includes the offshore highs of Orphan Knoll and Flemish Cap. This margin is broad and complex (Fig. 1.23) and includes deep basins such as the East Newfoundland, or Orphan Basin. Geophysical studies (Haworth and Keen, 1979; Keen and Barrett, 1981) indicate that the crust between Orphan Knoll and Flemish Cap and the shelf is thinned and subsided continental crust, and that this crust occupies a 450 km wide belt between the shelf and the continent-ocean boundary to the east of Orphan Knoll. Fig. 1.24 is a crustal cross-section through the Orphan Basin area. The Orphan Basin contains a very thick Tertiary sequence (3-4 km), Cretaceous sediments and a pre-Cretaceous section; these earlier sediments may be late Palaeozoic, Triassic or Jurassic, (Cutt and Laving, 1977), and are found in fault-bounded basins.[1] Tertiary subsidence in this area was very rapid. Various authors have published papers on geological correlations between this Canadian margin and the Porcupine Bank-Goban Spur margin, which provide useful parameters for Atlantic refits (Le Pichon et al 1977, Bailey 1974; Rast et al 1976; Bailey, 1975; Lefort and Haworth, 1977; Cherkis et al, 1973; Haworth and Lefort, 1979; Haworth, 1980). Geological markers that might be correlated if traceable across the continental shelves, which are unusually wide, are the Grenville front, the Caledonian front and Iapetus suture, the Hercynian front and the Avalon zone.

The Grenville front has been identified on Rockall Bank by Roberts (1974), but as this seems to have been interpolated between two localities only on the bank, the trend of the front is very poorly constrained.

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[1] There are probably two important unconformities within the sedimentary succession; one in the early Cretaceous, and another in the late Cretaceous. Lower Cretaceous and Jurassic marine sediments suggest that an epicontinental seaway existed before Cretaceous sea floor spreading (Cutt and Laving, 1977).

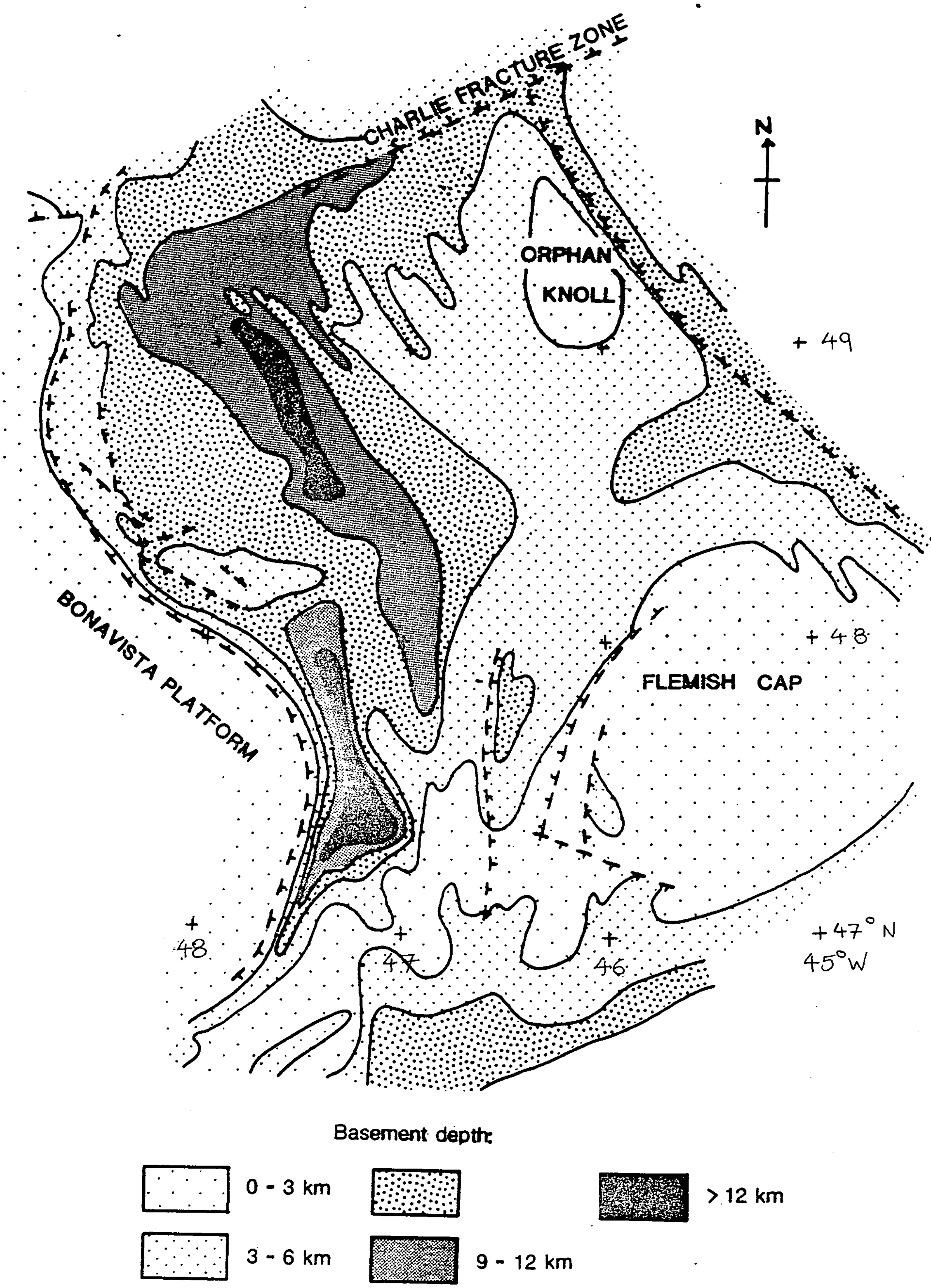

Fig. 1.23 Structural map of the Canadian margin south of the Charlie Gibbs Fracture Zone.

From: Geological Survey of Canada, 1975. Offshore Geology of Canada

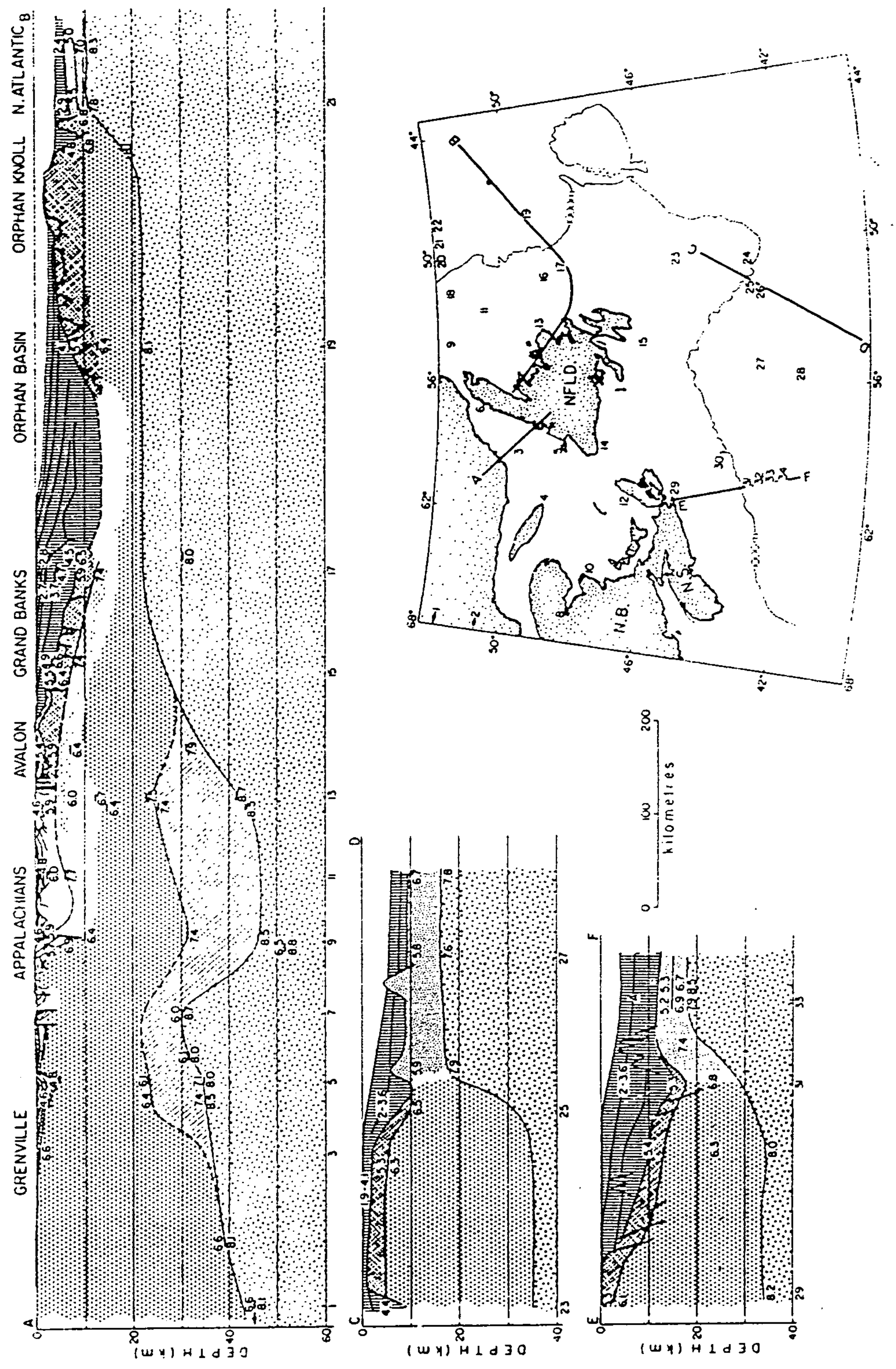

Fig. 1.24 Crustal cross-section through the Orphan Basin and Grand Banks areas (Haworth and Keen, 1979).

The Avalon Zone has been traced geophysically off the Newfoundland shelf (Lefort and Haworth, 1977) as a series of magnetic zones; but this is as yet unrecognised on the European shelf. (The zone is identified later in this thesis).

While Caledonian trends determined the trends of the northern shelf basins of western Britain (the Hebrides Basins etc) Hercynian trends influenced the formation of the Celtic Sea and Western Approaches Basins.

Cherkis et al (1973) postulated that the Charlie Gibbs Fracture Zone might be the link between the Hercynian fronts of Canada and Europe; however (J. Waldron, pers. comm.) there is no exact equivalent of the Hercynian front in Newfoundland or mainland Canada.

Fault trends can be used in constraining relative pre-drift plate orientations, but not relative plate positions. Fig. 1.25 is a trans-Atlantic correlation of Late Hercynian east-west faults.

When more is known about the geology of both margins, trans-Atlantic correlations will become obvious e.g. the continuation of the Cornubian granite system, and direct matching of basement geology. In most recent refits (see next section) the symmetry of banks and basins of the two opposing margins is striking; southern Porcupine Bank corresponding to Orphan Knoll; northern Porcupine Seabight to Orphan Basin, southern Porcupine Seabight to the basin between Orphan Knoll and Flemish Cap, and Goban Spur to Flemish Cap (see Fig. 1.26). Also of interest are the differences between the margins; i.e. the great Tertiary subsidence of the Canadian margin and the relative buoyancy of Porcupine Bank and the Celtic Shelf.

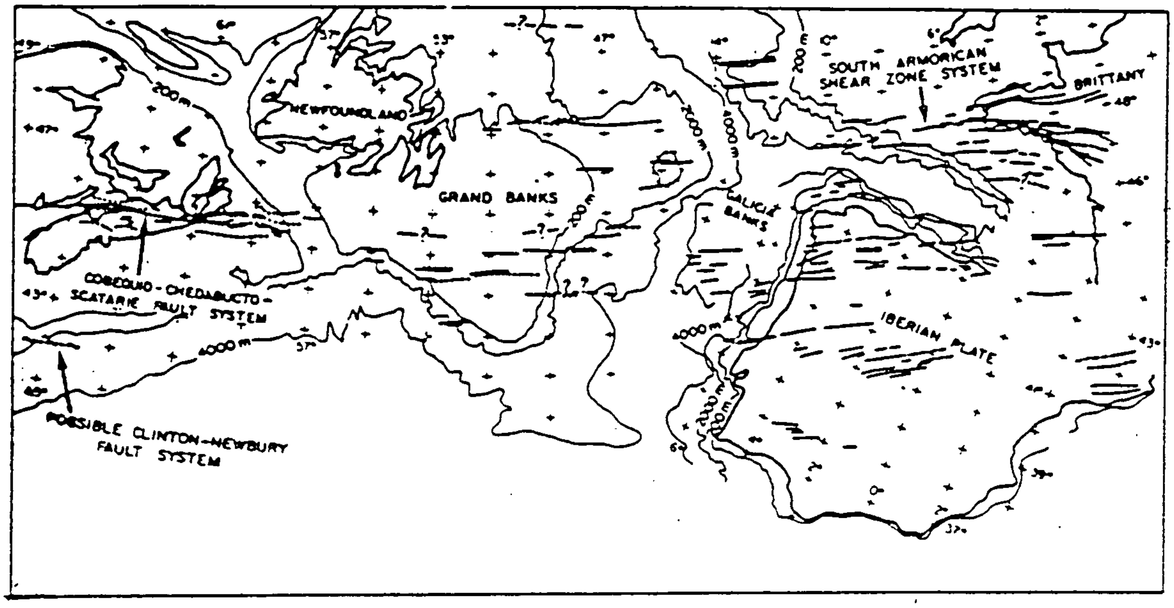

Fig. 1.25 Trans-Atlantic correlation of late Hercynian east-west faults (Haworth and Lefort, 1978).

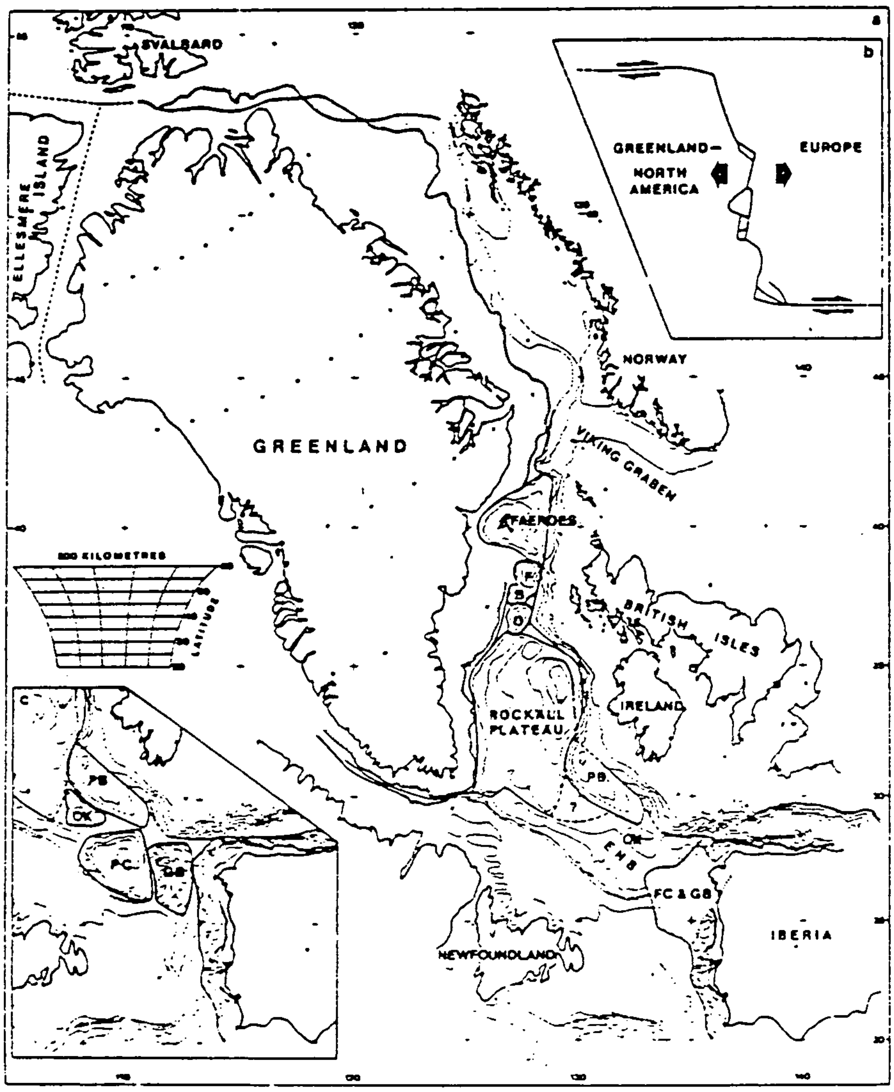

Fig. 1.26 Pre-anomaly 32 North Atlantic reconstruction, Russell and Smythe (1978).

1.3 ATLANTIC RE-ASSEMBLY AND THEORIES OF FORMATION OF ROCKALL TROUGH AND PORCUPINE SEABIGHT

This section should be treated as a framework for the further discussion and elaboration of Atlantic refits and theories of formation of these offshore basins in later chapters. Most authors (Roberts, 1975; Roberts et al 1981; Bott, 1978; Russell and Smythe, 1978) have considered most or all of the crust underlying Rockall Trough to be oceanic, and intimately connected with the formation of the North Atlantic ocean. Therefore a review and criticism of published fits serves to assess the plausibility of each theory.[1]

1.3.1 PUBLISHED ATLANTIC RE-ASSEMBLIES

a) Laughton's (1972) Atlantic refit is a very tight fit, and the shoal areas e.g. Orphan Knoll, Flemish Cap, are rearranged into a small area ignoring the deeper continental crust between these knolls and the continental shelf.

b) Le Pichon et al's (1977) refit (Fig. 1.27) uses the 2,000 m isobath and constraints from marginal fracture zones, which must result in a reasonable fit. The main unknown here is the position of the Iberian peninsular; on the basis of mapping of tectonic units across the Ibero-Armorican arc, most authors (Williams, 1975; Reis, 1978; Haworth and Lefort, 1979) would place the Iberian Peninsula further to the east. The Rockall Trough is partially closed in this

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[1]Notes on continental reassemblies. Le Pichon (1977) discusses the relative accuracy of methods of finding a continental refit, and concludes that the closing of an ocean along the originally offset (transform) margin is the best constraint, the only unknown being the tightness of fit. There are two main classes of continental reassembly; pre-stretching refits and post-stretching refits. Obviously the pre-stretching continental refit can only be found unequivocally if the position of the continent ocean boundary is known, the extent of thinned crust between this and unthinned continental crust, and the amount of crustal thinning. This information is only found after considerable investigation at a margin. Appendix A discusses crustal stretching at well-studied margins. For the Armorican margin this 'pre-stretching' limit of continental crust is at the base of the continental slope, and for the east American margin within the continental shelf.

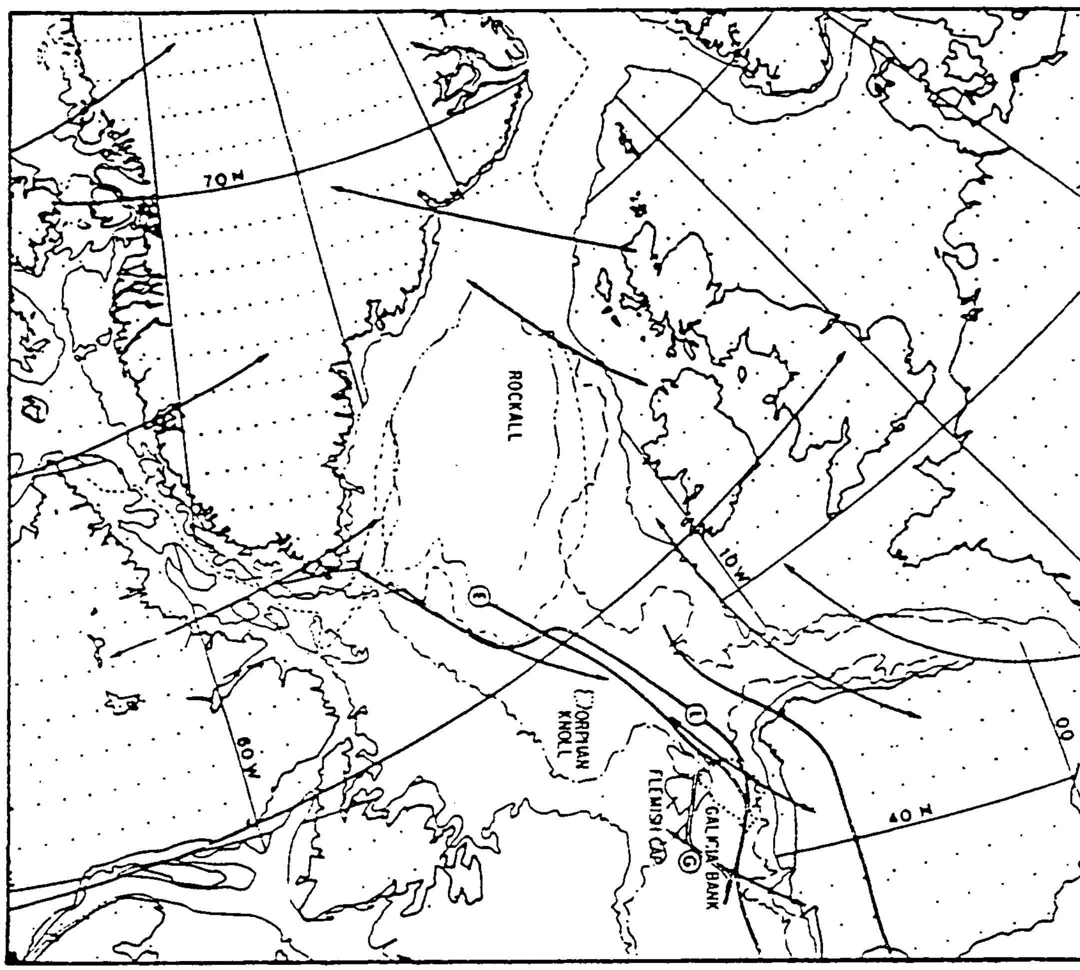

Fig. 1.27 Atlantic refit of Le Pichon et al, 1977

reconstruction, but the amount of closure of the Trough is dependent on the tightness of the fit at other margins, and it is not adequately known what this should be.

c) Srivastava's (1978) refit also utilised fracture zones, and used magnetic anomaly identifications; the fit (Fig. 1.28) is rather similar to Le Pichon's. Again Rockall Trough is partially closed. There is a large overlap of the 2,000 m isobath off Greenland which suggests that perhaps the whole fit should be rather looser.

d) Russell and Smythe's (1978) reconstruction incorporates an early Permian 'proto-North Atlantic' in Rockall Trough, the Faeroe-Shetland Channel, and off Norway (Fig. 1.26). Their fit as it is rearranges continental fragments in an unlikely manner (cf. Laughton), and places Iberia too far to the west (cf. Le Pichon),

e) Kristofferson's (1978) reconstruction is again not dissimilar to Le Pichon's and Srivastava's (Fig. 1.29). Kristofferson assumes that the crust in southern Rockall Trough is oceanic and formed before anom. 32 (34) time (there is a noisy magnetic zone here whose width is that of the pre-32 (34) noisy zone oceanic crust south of 52°N). This however is the only reason for the closure of the Trough - the refit would seem to be as good if Rockall Plateau were moved to the NW and Rockall Trough were already open at the time of initial oceanic formation.

1.3.2 THEORIES OF FORMATION OF ROCKALL TROUGH: CONTINENTAL OR OCEANIC?

Using constraints imposed solely by published North Atlantic reconstructions, the question of whether Rockall Trough is underlain by ocean crust, and if so of what age, is impossible to answer. The refits of Srivastava and Le Pichon partially close Rockall Trough, which is consistent with either limited ocean formation within the Trough, or pervasive syn-Atlantic opening continental crustal

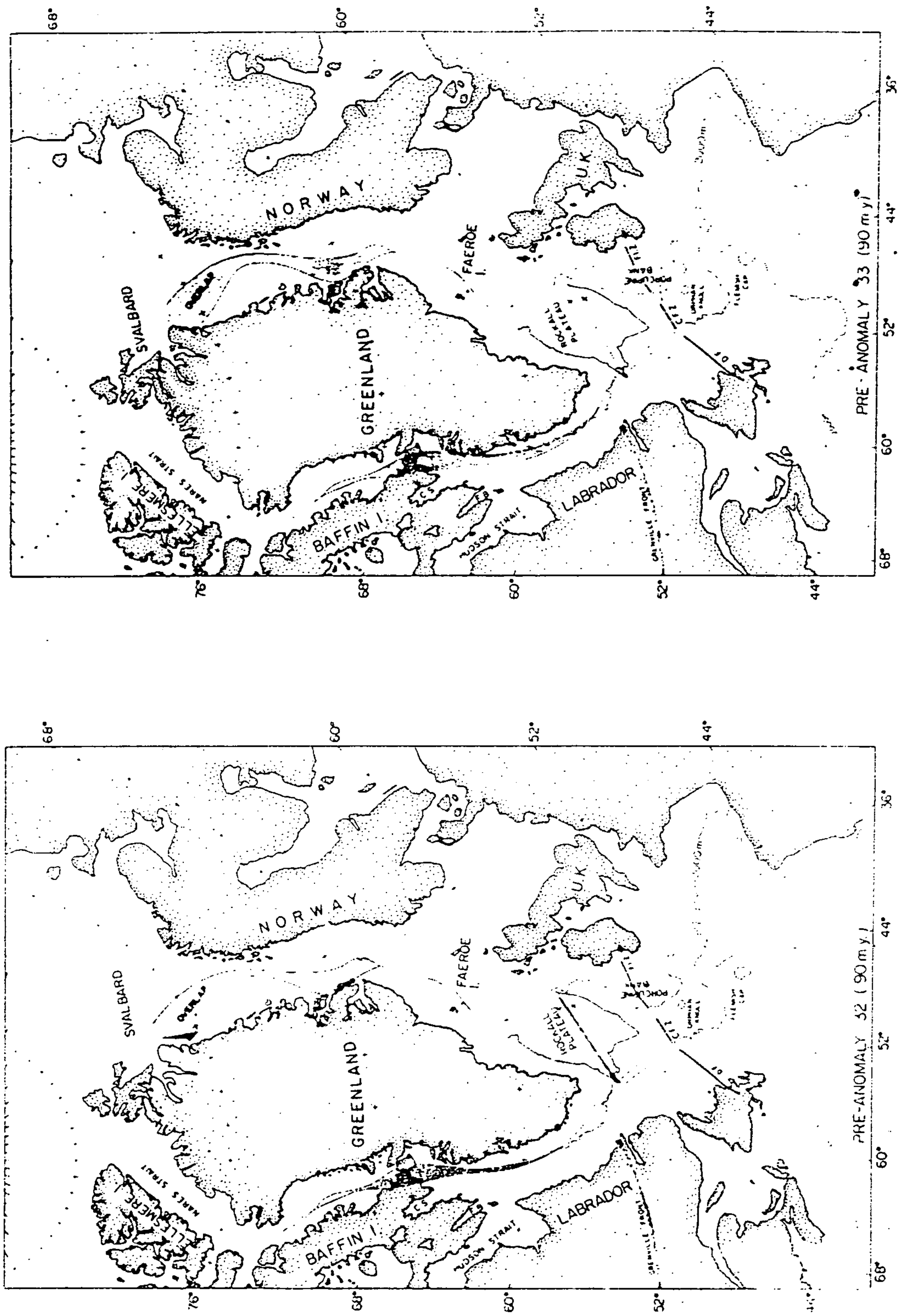

Fig. 1.28 Atlantic refits of Srivastava, 1978

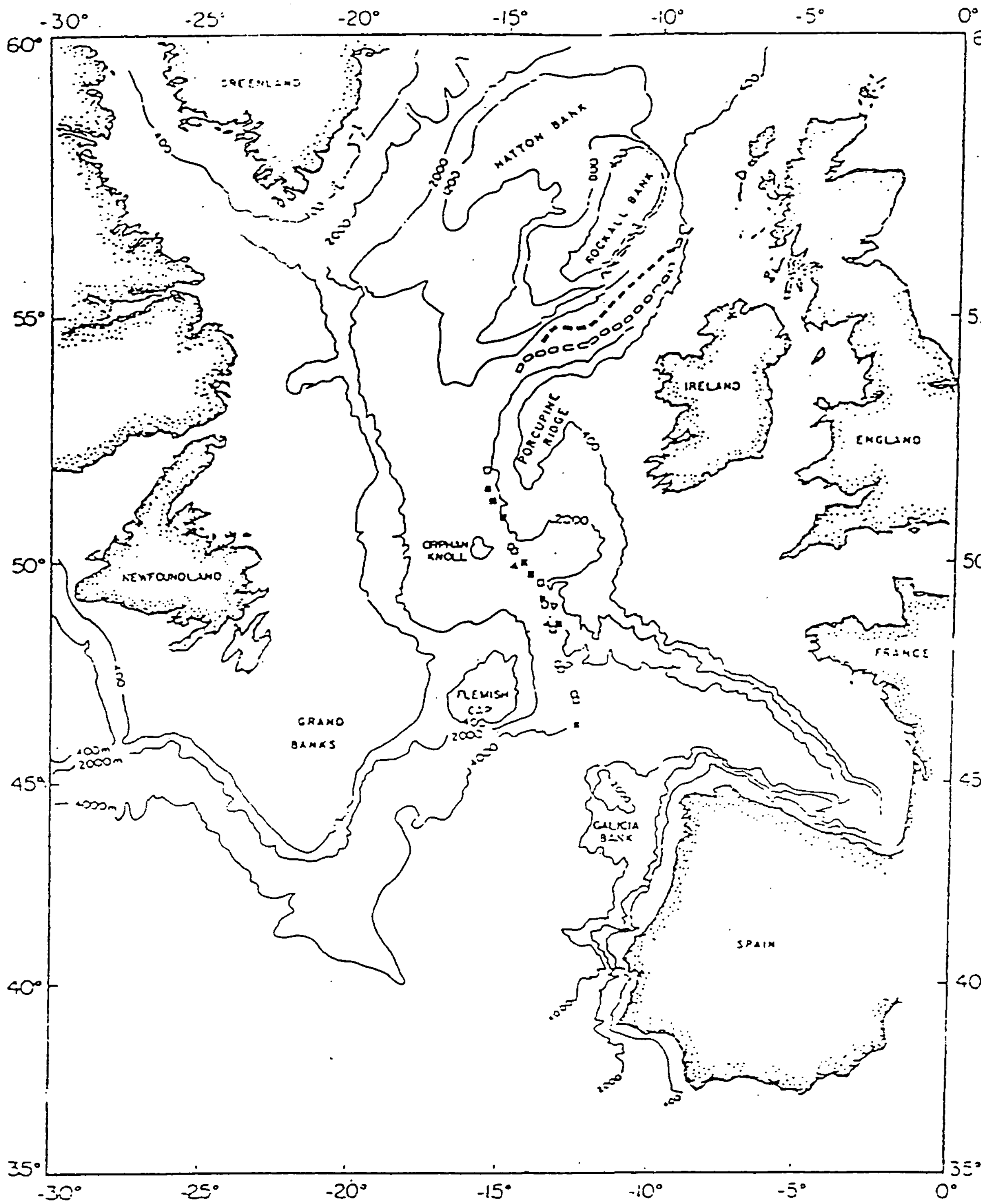

Fig. 1.29 Atlantic refit of Kristofferson, 1978

extension. On the other hand, Kristofferson's refit could be modified so that Rockall Trough was open at the time of initial spreading in the Atlantic, which would imply either earlier ocean formation (cf. Russell and Smythe), or earlier crustal stretching. The problem is that the amount of closure in Rockall Trough seems to be entirely dependent on the tightness of the fit elsewhere, especially in the Porcupine Bank - NE Newfoundland sector, and there is little direct information on the position of the continent-ocean boundary here and the amount of stretching.

a) Talwani and Eldholm (1972 and 1977), working on the Norwegian margin, postulate that a wide swathe of sea floor from the continental shelf edge to the Voring escarpments is underlain by very deep continental crust, and that from geometrical considerations, the Faeroe-Shetland Channel and Rockall Trough are also underlain by continental crust. Russell and Smythe postulate that this is Permian oceanic crust.[1]

b) Roberts (1975) assumed that the whole of Rockall Trough was underlain by ocean crust which formed in the Late Cretaceous quiet zone, as did Bott (1978), but Roberts et al (1981) now agrees more with Kristofferson in proposing that the centre of southern Rockall Trough is underlain by pre-anomaly 32 (34) oceanic crust, flanked by thinned continental crust. The paper by Roberts et al (op. cit.) is discussed extensively in Chapter 4, and new data from southern Rockall Trough are presented.

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[1]The author is prejudiced against Permian oceanic crust partly due to comparisons with other margins (Appendix A) where, except at transform margins, the usual transition to ocean crust is gradual, and not abrupt; but also because there is no Permian ocean crust observed in the world, other than in ophiolites. However, if it is possible to construct a refit from geometrical considerations in which Rockall Trough is underlain by continental crust which stretched before Atlantic spreading, then the same refit can be used for Permian oceanic crust.

1.3.3 THE FORMATION OF PORCUPINE SEABIGHT

As stated before, most authors have assumed that Porcupine Seabight is at least partially underlain by oceanic crust. Riddihough and Max (1976) assume that this was allowed to form in the northern Seabight by the lateral translation of Porcupine Bank along the E-W discontinuity at 53°N, after ocean had been formed in the southern Seabight along a NE-SW trend. It is generally assumed that the formation of the Trough was associated with Cretaceous spreading in Rockall Trough (e.g. Roberts, 1974, 1975); while Russell and Smythe (1978) propose that it is part of the postulated North Atlantic Permian ocean.

To the author, it seems that the question of the type of crust underlying Rockall Trough and Porcupine Seabight cannot be solved solely by published Atlantic reconstructions, or by previously available geophysical data. A large number of different theories have arisen because of the ambiguities inherent in distinguishing continental from oceanic crust, and in locating the continent-ocean boundary. This is discussed further in Appendix A; which is an important part of the background to this thesis.

1.4 OUTSTANDING PROBLEMS

In this area where there has been little academic exploration (see next chapter), a study will invariably produce information on previously unobserved basement and sediment features, and regional gravity and magnetic anomaly fields. Obvious problems are the position of the continent-ocean boundary off Porcupine Bank and the extent of continental crustal thinning, the amount of pre-anomaly 32 (34) ocean crust, and the type of boundary between Rockall Trough and the ocean crust of Porcupine Abyssal Plain. These data can be used to construct an improved Atlantic refit and a history of earliest plate separation in this area; if it is not possible to discover the type of crust underlying Rockall Trough and Porcupine Seabight directly (by observation and geophysical modelling) then it may be possible to infer this from the more accurate refit.

Other unknowns are the timing of tectonic events, geology of continental basement, and sedimentology of the margin; some of which are outside the scope of this thesis.

1.5 EXPLORATION ACTIVITY

Hydrocarbon exploration west of Ireland has been concentrated on northern Porcupine Seabight, with lesser activity on Goban Spur.

In the late 1970's discovery wells were drilled in Porcupine Seabight, and it seemed that the area might become a centre of production. Oil was found in blocks 26/28 and 35/8, with the Upper Jurassic being the main reservoir. However these finds are probably of the order of only 50-100 million barrels in size, and given the water depths of 1,200 to 1,500 feet are at present uneconomic. With the fall in oil prices, exploration activity has dropped in the Porcupine area. Two wells may be drilled in 1983.

Exxon is drilling a well on Goban Spur (block 62/12). With water depths of 2,000-5,000 feet any discoveries will be of commercial significance only in the very long term, and exploration interest in this area will be dormant for a considerable time.

CHAPTER 2

DATA COLLECTION AND REDUCTION, AND MODELLING TECHNIQUES

This thesis is based mainly on work based on new seismic reflection, gravity, and magnetic data obtained from two NERC funded cruises; Shackleton 3/79, and Challenger 6/80. All other available reflection seismic and gravity data from the study data were used in compiling the depth to basement and gravity charts. Chart 1 is a track chart of the Shackleton 4/79 and Challenger 6/80 cruises.

2.1 THE SHACKLETON 3/79 CRUISE

2.1.1 Data Collection

This cruise combined continuous seismic gravity, magnetic and bathymetric profiling with a very short seabed sampling programme. 1,600 km of seismic reflection profiles* were obtained, 2,000 km of magnetic data and 2,300 km of bathymetric and gravity data. The weather was bad, with gale force winds for over 50% of the time. However, although some proposed courses had to be modified, no time was lost and the instruments were towed in conditions as severe as gale force eight winds. On some lines the gravity data is noisy, although still useful. The reflection seismic records are also noisier on some lines, but nevertheless of reasonable quality.

The seismic reflection system was a 300 cubic inch Bolt airgun, firing every 16 seconds at approximately 1,800 psi, and a Geomechanique array with three active sections (Fig. 2.1). The data from the two channels were recorded, and recorded in analogue form only, as is usual with academic cruises. Each of the two channels were recorded on an EPC 4100 graphic recorder, and also recorded unfiltered on a RACAL store 4D tape recorder for later playback. Refer to Fig. 2.2. for details of the tape recording and air gun triggering system. The data recorded directly on the EPC recorders were first filtered through a 20-100Hz or 10-100Hz filter. One channel was recorded on a 4 second sweep, and the other on an 8

* All seismic reflection data are single channel

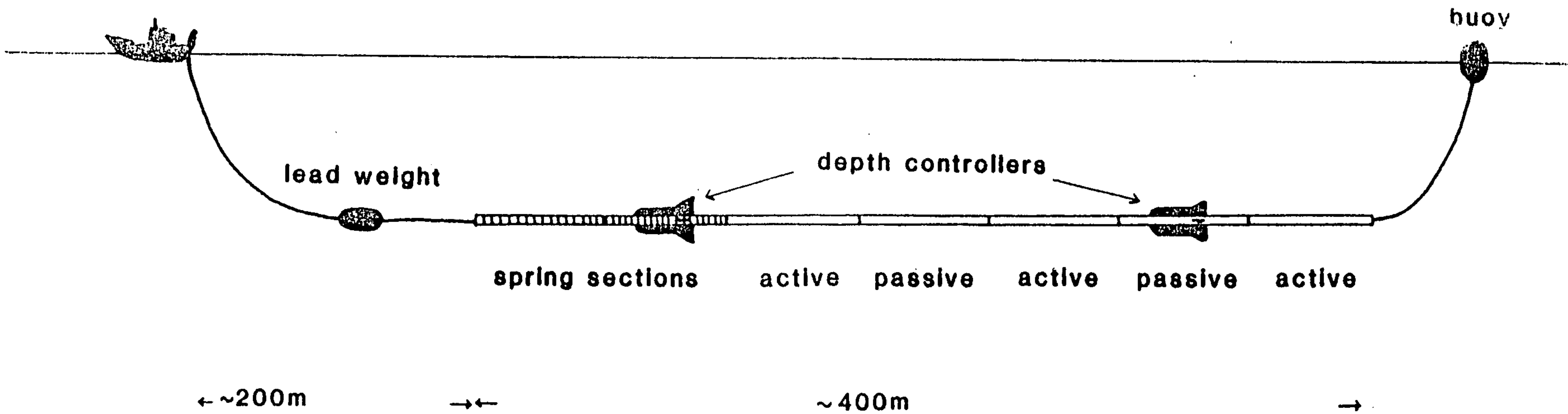


Fig. 2.1 Schematic diagram of the Geomechanique seismic array.

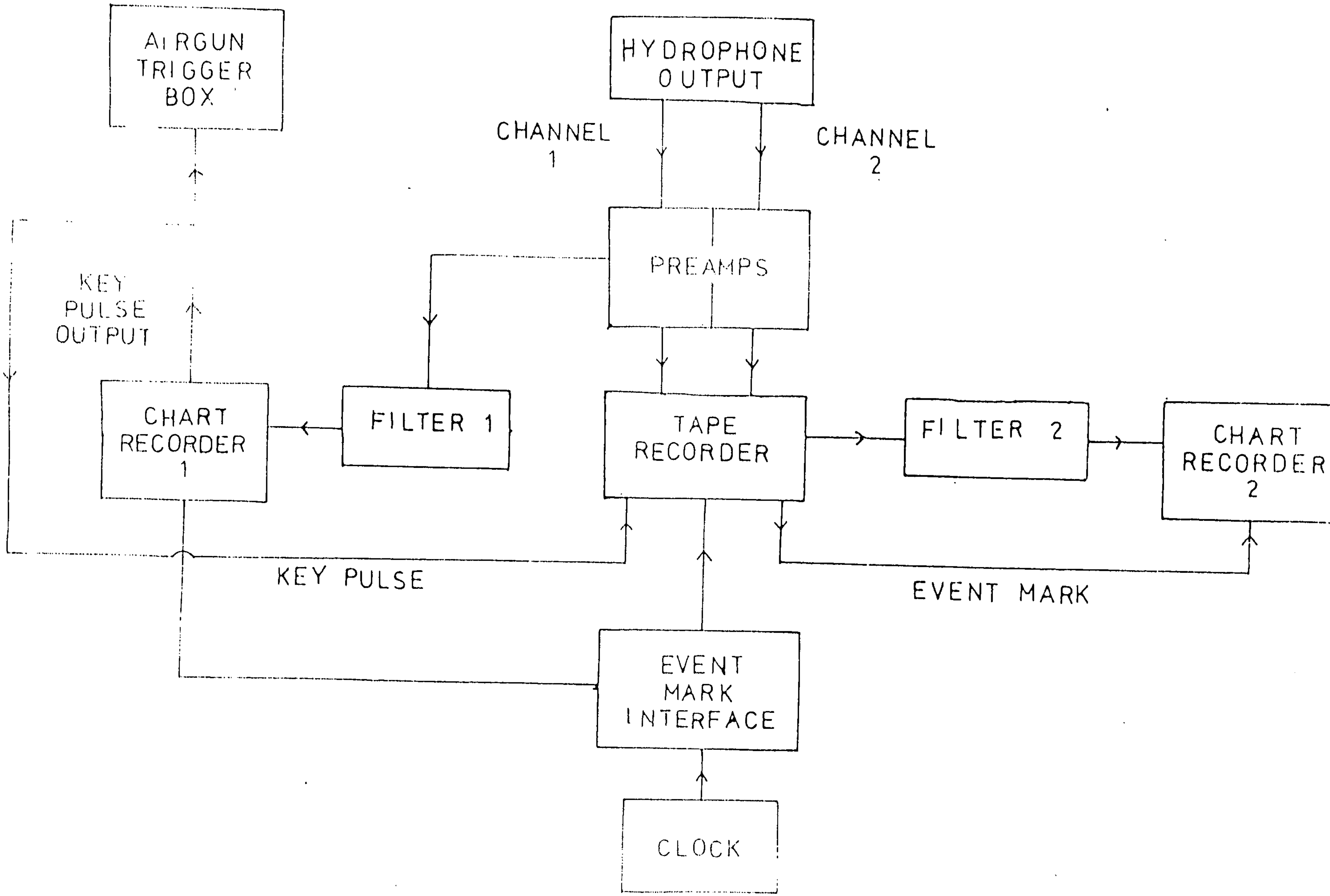


Fig. 2.2 Seismic recording system used on the Shackleton 3/79 cruise.

second sweep: the 8 second sweep record was found generally to be more useful.

Magnetic data were collected from a towed Varian proton precession magnetometer, gravity data from the La Coste-Romberg S40 Gravimeter, and bathymetric data from a towed IOS precision echo sounder. Digital data from all these sources were logged every second on the ship-borne data logger. Analogue hard copy records were also produced. The raw data were also logged by hand every ten minutes in the case of magnetic and bathymetric data, and every thirty minutes for gravity data. The hard copy records were annotated for calibration purposes at the same intervals.

Navigation was principally by Magnovox satellite navigation system. On the Shackleton 3/79 cruise the receiver failed occasionally and LORAN C fixes of reasonable quality were used at these times. When the satellite system was working satellite fixes were spaced at two hour intervals at most. The input from the dead reckoning sensors was recorded on the data logger at one second intervals.

2.1.2 Data Processing

The magnetic, navigation, gravity and bathymetric data were processed on the ship-borne IBM 1130 computer by staff from the NERC Barry base. Fig. 2.3 is a flow chart summarising the data processing software system (C. Poulson, pers. comm.). The reduced data are available at 2 minute intervals.

Further processing of the reflection seismic data took place on shore. This involved replaying the magnetic tape recordings onto an EPC graphic recorder at different frequency band passes. The process was speeded up by replaying most tapes at 4 times original speed. Various filters were tested on a few tapes e.g. 2-50Hz, 10-100Hz, 20-100Hz, 15-75Hz etc. All the tapes were replayed through a 20-100Hz filter with an 8 second sweep, and the better control of gain, stylus renewal etc., resulted in a noticeably better set of records than the originals. Because on the original records data was lost off the bottom of the records in some cases, most

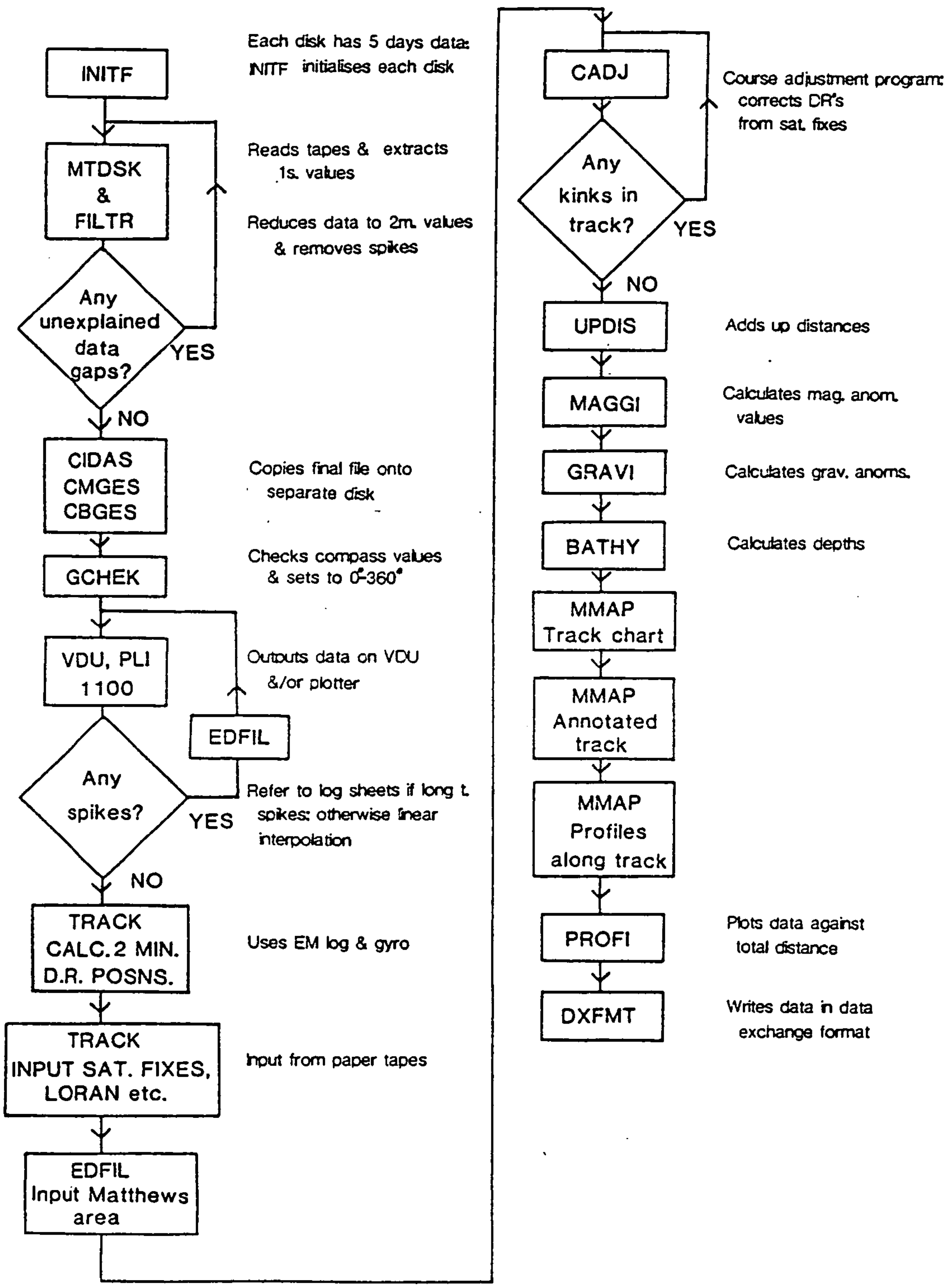


Fig. 2.3 Data processing flowchart

tapes were replayed with a 12 second sweep and low bandpass filter (2-50Hz). This enabled us to see occasional reflectors at 9 seconds and deeper.

2.2 THE CHALLENGER 6/80 CRUISE

We originally intended that the greater part of this cruise would be spent obtaining two or three long reversed seismic refraction lines in the mouth of Rockall Trough to investigate the deep crustal structure. However, the arrival of the Pop-Up Bottom Seismometers which were to be used was unavoidably delayed, and the whole cruise was devoted to underway geophysical surveying instead. The Shackleton 3/79 data were supplemented with further gravity, magnetic, bathymetric and seismic reflection information in southern Rockall Trough. The weather on this cruise was so good that excellent progress was made and we were able to run a longer survey than had been expected.

2.2.1 Data Collection

The equipment used was almost the same as that on the Shackleton 3/79 cruise; a Magnavox 1107 satellite navigator, 105 Precision echo sounder, Barringer proton precession magnetometer, La Coste and Romberg 586 gravimeter, and Decca Maglog digital data logger. The reflection seismic sound source was again a Bolt PAR airgun operating at 2,000 psi. Most of the profiles were collected using a 300 cubic inch chamber, but three short lines used a 1,000 cubic inch chamber in an attempt (largely unsuccessful) to gain deeper penetration. A problem with airgun mistriggering towards the end of the cruise resulted in the loss of two hours of reflection profiles, and a slight deterioration in the quality of the later records. No tape recorder was used in the collection of seismic data on this cruise; the two channels of output from the Geomechanique array were recorded directly on two EPC graphic recorders, one channel passing through a 15-100Hz filter, and the other through a 5-80Hz filter. Approximately 2,600 km of data were obtained, and the bathymetric, gravity and magnetic data logged.

2.2.2 Data Processing

No ship-borne computer was carried on this cruise, and it proved impossible to process the logged data on returning to Barry.

Pressure of time made it necessary to process the navigation and gravity data by hand, in Edinburgh, using the hard copy gravity record and the navigation data recorded manually on the data sheets. Sufficient navigational fixes had to be produced to enable the seismic and magnetic data to be used, and to process the gravity data.

a) Gravity and Navigation

The programme used to produce navigation information and to reduce the gravity data was LACROM, a programme developed by Scrutton (1971) to reduce gravity records to free-air gravity anomalies. Initially the programme requires navigational fixes, from which it interpolates position, heading, and speed at ten minute intervals, assuming constant speed and heading.

Given the gravity readings at 10 minute intervals, the FAA is then calculated. It proved possible to input satellite fixes (spacing range 5 minutes to 2 hours; most commonly half hour to 1 hour) and by adjusting the programme, obtain outputs of position and the gravity anomaly every 10 minutes. Neither the navigation nor the gravity data were reduced for the curved ships tracks at the ends of lines; the tracks were treated as a series of straight lines. Where a satellite fix did not coincide with the end of a line or a within-line course change, a position was extrapolated using heading and speed data from adjacent satellite fixes.

Gravity data were read from the hard copy analogue output at ten minute intervals. As there is a delay of 4 minutes within the gravimeter, readings were taken at 4, 14, 24 minutes for data at 0, 10,20 minutes.

Fig. 2.4 is the output for one line, showing the input satellite positions at the top with the calculated speed and course between fixes, and below this the interpolated positions, Eotvos corrections and free-air anomaly. The output was checked by comparison with manual calculations by examining output for consistency of speed and course, and for consistency of the Eotvos correction within lines. Errors were picked up in this way. This method of calculating

LINE 5 GRAVITY

ZONE	DAY	HR	MIN	LAT	LONG	SPEED	COURSE
24.	98.	16.	5.	53. 15.0	18. 35.7	0.0	0.
24.	98.	16.	55.	53. 13.5	18. 28.1	5.7	108.
24.	98.	18.	10.	53. 11.1	18. 17.9	5.3	111.
24.	98.	18.	36.	53. 10.5	18. 14.6	4.8	107.
24.	98.	18.	53.	53. 10.0	18. 12.3	5.2	110.
24.	98.	19.	40.	53. 9.0	18. 6.0	5.0	105.
24.	98.	19.	53.	53. 8.7	18. 4.2	5.2	106.
24.	98.	20.	30.	53. 7.9	17. 59.0	5.2	104.
24.	98.	20.	38.	53. 7.7	17. 57.9	5.2	107.
24.	98.	21.	44.	53. 6.0	17. 49.8	4.7	109.
24.	98.	22.	20.	53. 4.9	17. 45.4	4.8	113.

DAY	HR	MIN	LAT	LONG	FAA	EOTVOS
98.	16.	10.	53 14.9	-18 -34.9	32.6	24.6
98.	16.	20.	53 14.6	-18 -33.4	31.5	24.6
98.	16.	30.	53 14.3	-18 -31.9	29.5	24.6
98.	16.	40.	53 14.0	-18 -30.4	30.4	24.6
98.	16.	50.	53 13.7	-18 -28.9	29.4	24.6
98.	17.	0.	53 13.3	-18 -27.4	28.9	22.0
98.	17.	10.	53 13.0	-18 -26.1	30.7	22.1
98.	17.	20.	53 12.7	-18 -24.7	29.7	22.1
98.	17.	30.	53 12.4	-18 -23.3	27.7	22.1
98.	17.	40.	53 12.1	-18 -22.0	27.7	22.1
98.	17.	50.	53 11.7	-18 -20.6	28.6	22.1
98.	18.	0.	53 11.4	-18 -19.3	27.6	22.1
98.	18.	10.	53 11.1	-18 -17.9	25.1	20.6
98.	18.	20.	53 10.9	-18 -16.6	23.1	20.6
98.	18.	30.	53 10.6	-18 -15.4	22.0	20.6
98.	18.	40.	53 10.4	-18 -14.1	21.4	22.0
98.	18.	50.	53 10.1	-18 -12.7	20.4	22.0
98.	19.	0.	53 9.9	-18 -11.4	19.2	21.8
98.	19.	10.	53 9.6	-18 -10.0	18.2	21.8
98.	19.	20.	53 9.4	-18 -8.7	18.2	21.8
98.	19.	30.	53 9.2	-18 -7.3	16.2	21.8
98.	19.	40.	53 9.0	-18 -6.0	14.9	22.5
98.	19.	50.	53 8.8	-18 -4.6	14.8	22.5
98.	20.	0.	53 8.5	-18 -3.2	13.1	22.8
98.	20.	10.	53 8.3	-18 -1.8	14.1	22.8
98.	20.	20.	53 8.1	-18 -0.4	15.1	22.8
98.	20.	30.	53 7.9	-17 -59.0	14.6	22.4
98.	20.	40.	53 7.6	-17 -57.7	13.2	20.0
98.	20.	50.	53 7.4	-17 -56.4	15.2	20.0
98.	21.	0.	53 7.1	-17 -55.2	16.2	20.0
98.	21.	10.	53 6.9	-17 -54.0	17.2	20.0
98.	21.	20.	53 6.6	-17 -52.8	20.1	20.0
98.	21.	30.	53 6.4	-17 -51.5	21.1	20.0
98.	21.	40.	53 6.1	-17 -50.3	23.0	20.0
98.	21.	50.	53 5.8	-17 -49.1	23.0	19.9
98.	22.	0.	53 5.5	-17 -47.9	23.9	19.9
98.	22.	10.	53 5.2	-17 -46.6	24.9	19.9
98.	22.	20.	53 4.9	-17 -45.4	25.8	19.9

Fig. 2.4 Gravity information for a complete line calculated using the programme LACROM.

positions by simply interpolating between satellite fixes results in a track made up of short straight line segments, with abrupt but small changes in the Eotvos correction and hence in the FAA. These 'jumps' in the Eotvos correction are commonly of about 2 mgal magnitude, and of 5 mgal magnitude at most. While this is clearly undesirable, and would be smoothed out by the more sophisticated processing usually used, it is felt that the results are sufficiently accurate for marine gravity data: 2 mgal is a relatively small amount of noise. The final test is that of crossover errors. Crossover errors within the Shackleton 3/79 cruise (excluding crossovers occurring where FAA gradients are very steep) have an average value of approximately 3 mgal. Within the Challenger 6/80 cruise the average value is approximately 4 mgal.

Although it would be useful to eventually process the data in the usual manner for purposes of comparison, this method is evidently satifactory for the production of a provisional gravity map.

b) Magnetic Data

It was decided not to reduce all the magnetic data to obtain anomalies, as a good total field contour map (Riddihough, 1975) and anomaly contour map (Roberts and Jones, 1976) already exist for this area. When an anomaly profile along a Challenger 6/80 track was required for modelling, the IGRF (IGS pub. 71/1) was removed by hand.

Contour charts of geophysical data (magnetic and gravity fields) and geological data (depth to basement, bathymetry) are necessary in order to order data and provide a framework for study, and also to discover whether a two-dimensional crustal model is valid or not. A substantial part of the original work of this thesis was the preparation of contour charts of free air anomalies, depth to basement, sediment thickness etc. All charts were originally plotted on mercator projection and at 1:500,000 scale. It was felt that this choice of scale (as opposed to 1:1,000,000) gave the greatest accuracy possible to our data spacing. All contouring was done by hand.

2.3 DATA BASES AND DATA HANDLING

2.3.1 Gravity Data

A free air anomaly contour chart was available for the eastern part of the study area, i.e. Porcupine Bank (Buckley and Bailey, 1975) These data were recontoured with the new Edinburgh data and all other available data. Fig. 2.5 shows ships tracks for which there is free air anomaly information. As discussed in the last section, the Edinburgh data are good. The other data were on the whole collected a decade or more ago and show much greater inconsistencies from line to line, presumably both because of less accurate navigation, more frequent tares, and greater drift within the gravimeter. The change of reference systems further increases the data incompatibility. Occasionally crossover errors are 15 mgal or more. The Edinburgh data were treated as a reference set and the other data adjusted by up to 10 mgal line by line to minimise crossover errors with the Edinburgh data. The contour map (Chart 2) is of good quality where the Edinburgh tracks are closely spaced, in particular over the SW margin of Porcupine Bank where the continental slope is very steep and the bathymetry can be used as a guide to the free air anomaly contouring.

In other areas, inconsistencies remain between tracks and some data had to be discarded. Where the gravity field is fairly flat and there are no Edinburgh data, such as on northern Procupine Bank, the contours may be inaccurate by up to 10 mgal. The contour chart shows two different categories of contour; a solid contour for confident interpolations between closely-spaced good data, and a dashed contour for interpolations between widely spaced and/or poor data.

2.3.2 Magnetic Data

As mentioned in the previous section, there was no need to contour the Edinburgh magnetic data as fair quality magnetic contour maps already existed for this area. For most of the period in which the work for this thesis was done, a total field chart was

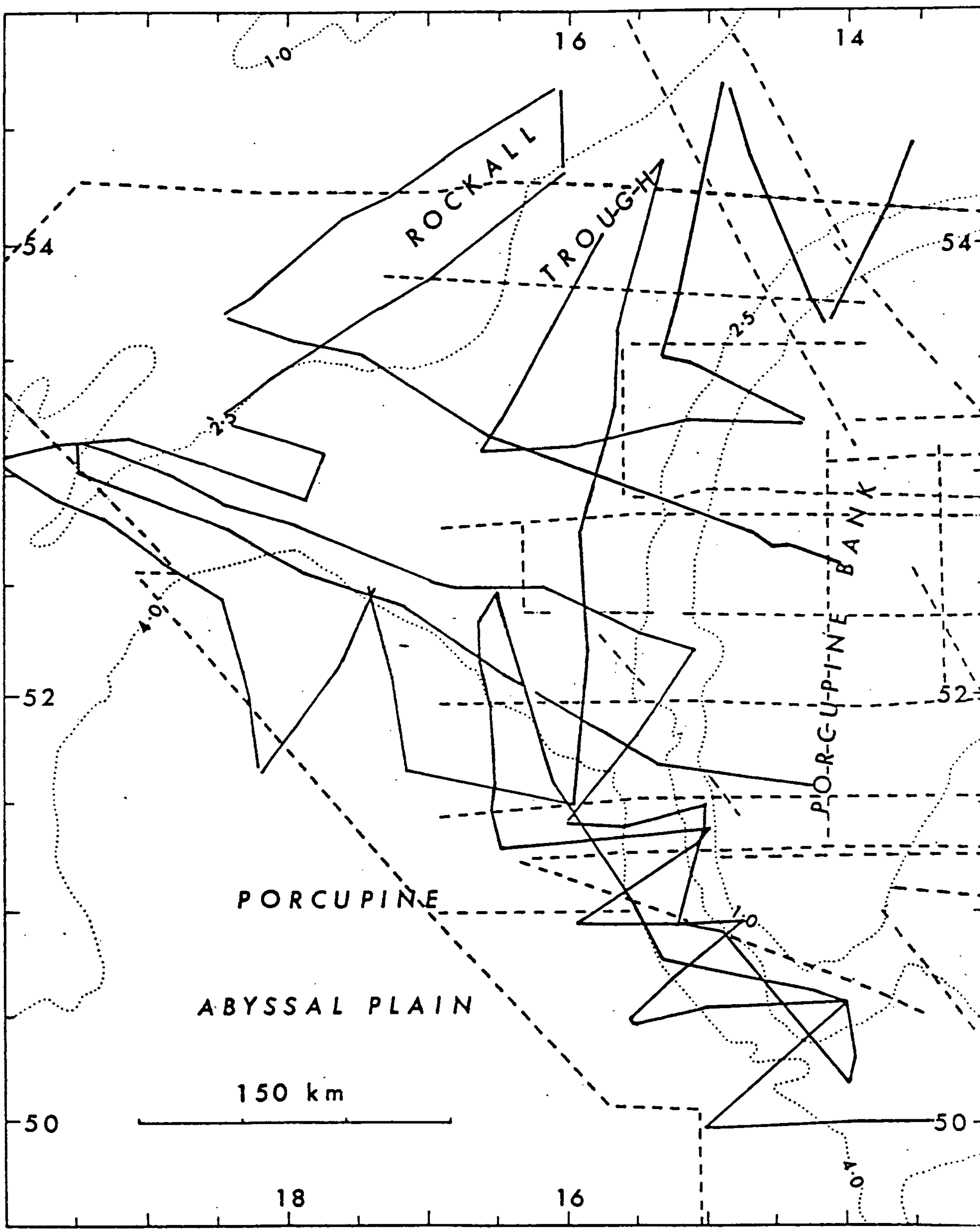


Fig. 2.5 Location of gravity data in the study area.

Solid lines : Edinburgh data

Dashed lines : University College of North Wales, Conrad, Vema, and Charcot data.

used (Riddihough, 1975. This chart is based on US Navy data). Towards the end of this time an anomaly chart became available (Roberts and Jones, 1976).

2.3.3 Bathymetric Data

A bathymetric chart was compiled for the area by combining two different contour charts; a Canadian chart for the area to the west of 18°W, and the Institute of Oceanographic Sciences bathymetric chart to the east of 18°W. The match between the two charts was good to the north of 52°: to the south of 52°, in Porcupine Abyssal Plain, the fit was bad and a match between the two at 18°W was not attempted. Fig. 2.6 shows this compilation chart. It was found that some fairly large discrepancies exist between this chart and the new Edinburgh bathymetric data. However, it was not thought worth re-contouring, and Edinburgh bathymetric data is used for modelling purposes.

2.3.4 Seismic Data

a) Depth to Basement Chart

A depth to acoustic basement chart of the area was prepared (Chart 3) in contours of seconds two-way travel time. Too little is known about the velocity structure of the sediments to justify the production of a depth to basement chart with depth contours. Fig. 2.6 shows the Edinburgh ships tracks along which the seismic reflection system was run, and positions of other seismic reflection records used in the compilation. These other records are from cruises of Charcot (GSCOB, 1971) and Cherkis, 1973. The data quality is not as good as the Edinburgh records.

Roberts et al (1981) published a contour chart of depth to basement within Rockall Trough. Values of depth to basement at positions where the ships tracks intersect were picked off, and used to modify the contour chart presented in this thesis.

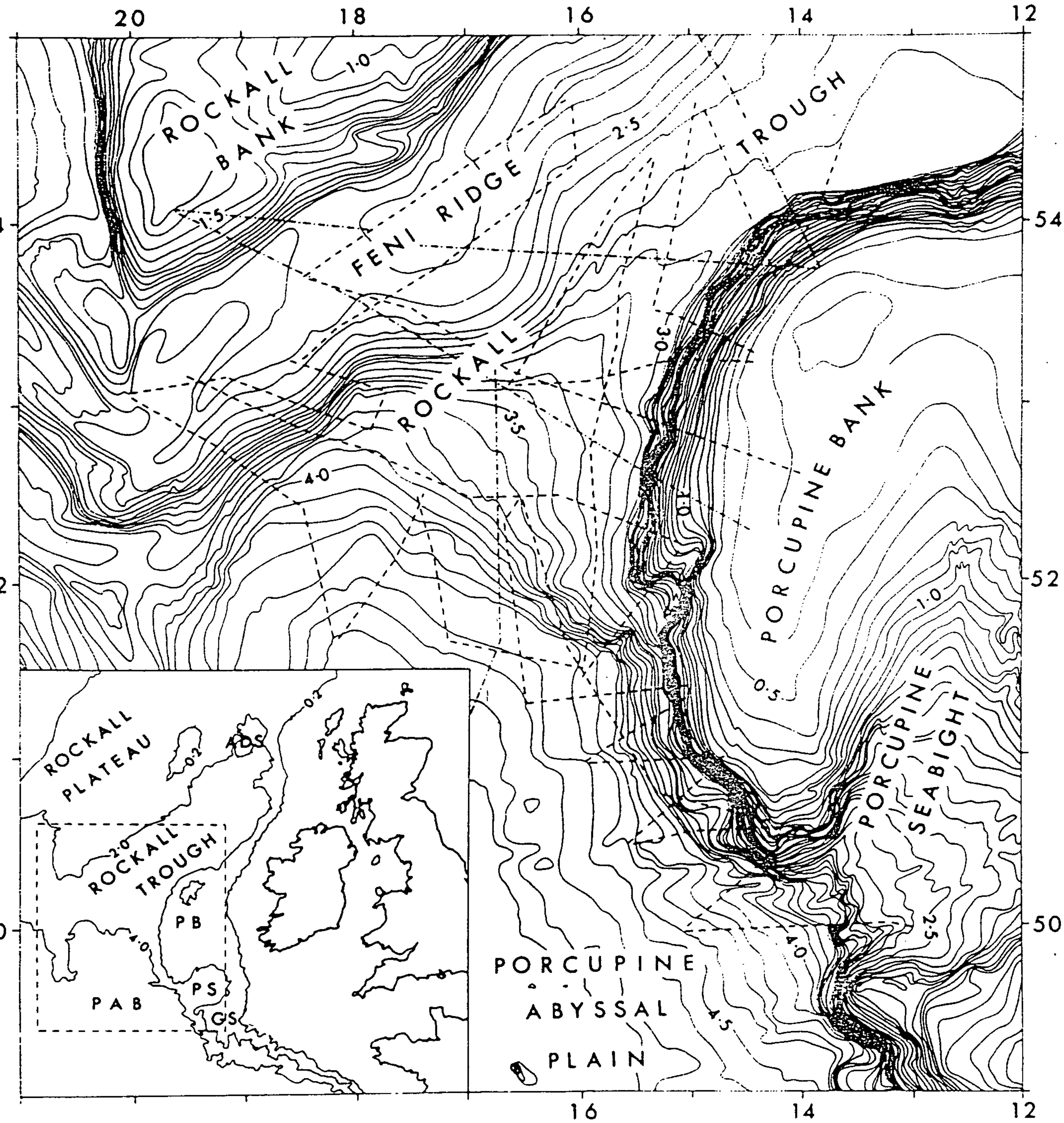


Fig. 2.6 Bathymetry in kilometres in the study area. Tracks for which seismic data are available are also shown:

- - - - Edinburgh cruises

-·-·-·- Charcot cruise, and Cherkis (1973)

b) Problems With Interpretation

Accuracy of interpretation and timing of events is limited by the data quality. Noise is introduced by

i) water-bottom multiples. In relatively shallow areas such as Porcupine Bank, strong water-bottom multiples may completely obscure primary information. Over the deepest part of Rockall Trough and Porcupine Abyssal Plain water depths are sufficiently great for the first multiple to appear well below basement depths.

Other multiples, such as reflections off the water-sediment interface or the R4 (Challenger) reflector are unremovable and also degrade the data.

ii) the bubble pulse. In these data the bubble pulse is almost 0.5 seconds (TWT) long and greatly reduces resolution. This is due both to the method of shooting with a single airgun instead of a tuned array, and to the lack of later processing such as deconvolution.

iii) diffractions. Migration of data collapses diffractions and makes interpretation of faulted areas much more simple. Pervasively block faulted basement appears as a mass of diffractions on unmigrated data, and resolution of basement type may be difficult.

Problems in interpretation may also be introduced by geology. In southern Rockall Trough volcanic horizons locally obscure the depth of true basement.

c) Seismic Stratigraphy

The seismic stratigraphy of the area is treated in detail in Dingle, Megson, and Scrutton 1982 (Appendix B).

In brief, Roberts (1975) set up a provisional seismic stratigraphy in Rockall Trough, which recognised four major horizons: R4 (Upper Eocene-Oligocene), X (?60 m.y.), Y (?76 m.y.), and Z (?100 m.y.). In Roberts, 1979, the age of the Y reflector was redefined as 52-55 m.y.

Dingle, Megson, and Scrutton set up a separate stratigraphy, using Edinburgh seismic data (Fig. 2.7). The Challenger reflector is roughly equivalent to Roberts' R4 (cherty?) reflector, and the Shackleton reflector probably correlates with Roberts' Y horizon.

Two important differences between the interpretations of Dingle, Megson and Scrutton, and Roberts, are that

i) Dingle, Megson and Scrutton recognise a Charcot - Challenger sediment package associated with the formation of the Feni Ridge
and

ii) While Roberts (1975) shows faulting which cuts the R4 (Challenger) horizon, Dingle, Megson, and Scrutton recognise no faulting younger than the time of Shackleton deposition.

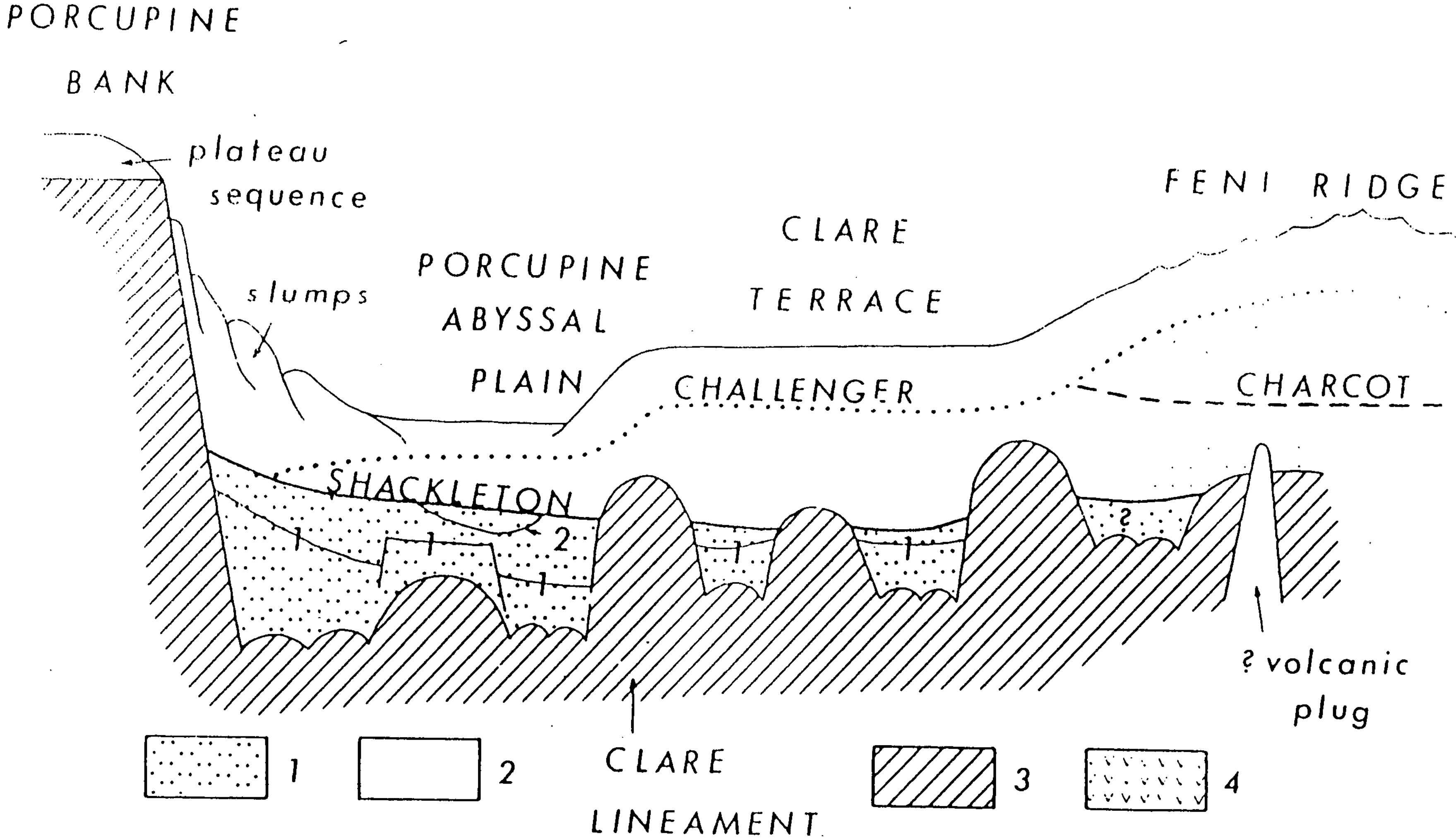


Fig. 2.7 . Schematic correlation chart showing the relationships between the seismic stratigraphic subdivisions used in Dingle, Megson, and Scrutton. The section extends westwards from Porcupine Bank, and turns northwards over Porcupine Abyssal Plain towards the Feni Ridge, keeping east of 17°W.

1. Pre-Shackleton sequence containing three packages defined by reflectors 1 and 2
2. Post-Shackleton sequence
3. Acoustic basement
4. Lavas or sills.

(Not to scale)

2.4. GRAVITY AND MAGNETIC MODELLING TECHNIQUES

2.4.1 Limitations of Modelling Potential Field Data

In this thesis, gravity and magnetic crustal models are constructed in order to transform a set of geophysical data into valid geological information. Thus the problem is one of forward conversion of potential field data, which is inherently non-unique. If enough is known about parameters such as density or magnetisation, and the shape of the anomaly-producing body, the other parameters of the anomaly-producing body can be found from the data by using an inverse method, and the solution is a unique one (Smith, 1961; Bott 1961). The solution however is only as valid as the original assumptions allow. Even if sufficient reasonable initial assumptions can be made, the solving equation may be a non-linear function (Bott, 1961), or the shape of the anomaly-producing body may be too complex to express as a set of solvable equations. The alternative technique of finding a model to fit the data is termed the forward method, where models are constructed, and the potential field due to the model calculated and tested against the observed field. If no initial assumptions are made, an infinite number of models can satisfy a given anomaly (this does not mean that the parameters of the models are infinitely variable, but that there are an infinite number of models that can exist within certain model parameter bounds). Again, the validity of the model depends upon the assumptions made in constructing the model. If these constraints are based on geological knowledge of the area, or other geophysical information, the models are more likely to reflect reality. The construction of models is a very subjective matter, as different workers use different model values, if these are not well bounded, and place different weights on initial information such as seismic refraction data, geological maps, etc.

Techniques Used In This Thesis

In this thesis, the forward method only is used. Complex shapes are assumed, and the observed potential field profile (in the case of gravity data particularly) is a result of anomalies produced by several different layers. All crustal models are two-dimensional, and modelled along seismic reflection profiles. The two-way travel time to reflecting horizons (usually sediment water interface and sediment acoustic basement interface) are converted to depths in kilometres using travel times of Vp = 1.5 km/s for water, and Vp = 2.5 km/s for the whole sediment column. This provides the parameters for a unique, and hopefully valid solution. Bodies of different densities or magnetisations are treated as 2D prisms in the programmes used to calculate the anomaly due to the model. The anomaly due to the prisms is calculated every 5 km or 2.5 km. At either end of the model, the polygons are continued uniformly for 150 km to eliminate edge effects.

The initial assumptions made in this modelling are that:

a) acoustic boundaries are coincident with density/magnetisation boundaries

b) geological structures can be represented by layers of constant density/magnetisation, which in turn can be represented by polygons of constant density/magnetisation.

c) the velocities used in the conversion of travel times to depths are reasonable. A Vp of 2.5 km/s was chosen for the sediment column as this is the velocity that Hill (1957) found for most of the sediment column in southwestern Rockall Trough.

d) the line can be modelled two-dimensionally. Close consideration is given to the contour chart to see whether the anomalies along a profile are affected by any nearby structure which cannot be incorporated in a 2D model. Any line which is not nearly orthogonal to structural trends is rotated. The survey was designed however so that most lines are within 10° or 20° of being normal to basement structure.

2.4.2 Gravity Modelling

The starting point for gravity modelling was a density model constructed for the Porcupine Bank - Goban Spur area by Scrutton (unpub.). The continental crust is divided into upper crust (ρ = 2.76 g/cm³) and lower crust (ρ = 2.94 g/cm³) by a boundary at 11.5 km (refer to Fig. 2.8).

The mass this gives of 8.61 x 10^6 g for a 1 cm³ column from 0-30 km below sea level, is used as the basis for isostatic calculations.

The other variable used for the initial models is a density of 2.2 g/cm³ for the whole sediment column. This is consistent with a Vp of 2.5 km/s for the whole sediment column. This can be compared with North Sea average sediment densities of 2.06 g/cm³ for the Eocene, 2.21 g/cm³ for the Palaeocene, 2.44 g/cm³ for the Cretaceous (2.35 if deeply-buried chalks are ignored), and 2.43 g/cm³ for the Jurassic-Triassic (Donato and Tulley, pers. comm). The density of 2.2 g/cm³ may therefore be a reasonable average for a Cretaceous and younger sedimentary column.

No use is made of the isostatic gravity anomaly in this thesis. The isostatic gravity anomaly is the observed anomaly minus the anomaly produced by any model in equilibrium for which the whole sediment and crustal column is assigned the same density, and the isostatic crustal thickness calculated. The isostatic gravity anomaly is usually calculated when there is no information on sediment thickness, and can be used as a guide to the presence of sedimentary basins. All models in this thesis are constructed where there is reflection seismic information.

The 'Residual Isostatic Anomaly'

The model is constructed by plotting the sediment and acoustic basement surfaces on the boundaries between the sediment and water prisms, and the sediment and upper crustal prisms, and the appropriate

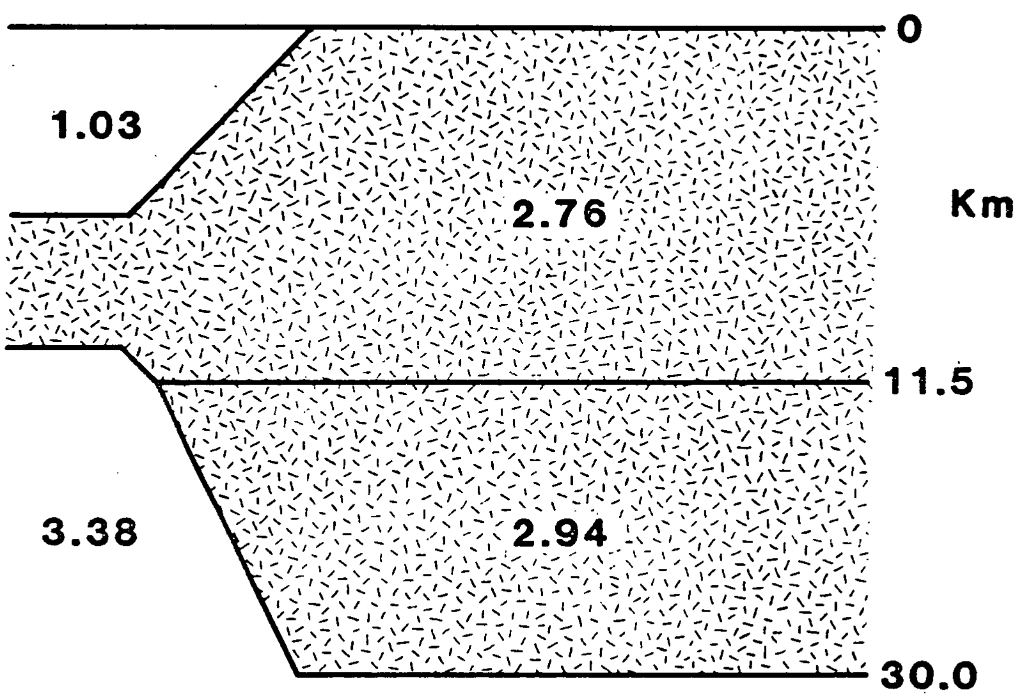


Fig. 2.8 Model of crustal densities in the Goban Spur - Rockall Trough area, by Scrutton (unpub.). Densities are obtained by converting velocities from seismic refraction experiments run in the area.

velocities are assigned. Crustal thickness are calculated (manually) by assuming local Airy isostasy. Any crust deeper than 11.5 km is given a density value of 2.94 g/cm^3. The residual isostatic gravity anomaly is the observed gravity minus the gravity calculated using the isostatic model. The model can then be altered by inserting lateral density changes or assuming that the crust is not all in isostatic equilibrium to give a better fit between calculated and observed anomalies.

The crustal thicknesses calculated should be taken as a rough guide only, as calculated thickness is dependent on the assumed crustal density, which is very badly constrained (refer also to Appendix A). It is further assumed in the models that there are no lateral density variations in the mantle.

2.4.3 Magnetic Modelling

In general, magnetic modelling involves more unknown variables than gravity modelling, and the results should be treated with even more caution. If it is assumed that the magnetisation of the anomaly-producing body is not all induced, then the variables that have to be considered are:

a) The inclination and declination of the remanent magnetic vector

b) The magnitude of remanent magnetisation. Kent et al., (1978) report that NRM intensities of oceanic gabbros range over three orders of magnitude.

c) The inclination and declination of the induced magnetic vector

d) The magnitude of induced magnetisation.

The vector addition of (a), (b), (c), and (d), gives the total magnetisation vector. In practice, simplifying assumptions are usually made. It is assumed in this thesis that the magnetisation of oceanic rocks (apart from these at fracture zones) is predominantly remanent, and the induced magnetisation neglible, and that the magnetisation of continental rocks is predominantly induced. The vector inclinations and declinations are also simplified. An inclination of +68° and a declination of -14° was chosen for the induced vector, which are the parameters of the present-day magnetic field at the latitudes and longitudes of the study area (IGS Report 71/1). Furthermore, it was assumed for most models initially that the direction of remanent vector of oceanic crust, if positively magnetised, is the same as the induced vector. In one model for which the basement topography was well constrained, a Cretaceous vector was substituted for a present day vector, and the anomaly was found to be very little different.

e) Depth to Base of Magnetic Layer

The question of which magnitude of magnetisation to use for various rocktypes is complex. Measured magnitudes vary enormously. In practice, when modelling, the choice of a reasonable magnetisation is also dependent upon the assumed thickness of the magnetic layer, which is also very poorly constrained. For instance, when making the original models of oceanic magnetic reversals, Vine and Matthews (1963) assumed that the magnetic portion of the crust was defined by the depth to the Curie point isothem, which is about 20 km below sea level. Vine and Wilson (1965) on the other hand assumed that the magnetic layer was one to two kilometres thick and consisted of basalts, overlying a nearly non-magnetic serpentinite layer. More recent studies have shown that if a magnetic intensity of 5 to 15×10^{-3} gauss is assumed for oceanic basalts, then the magnitude of observed anomalies can be accounted for by a magnetic layer 0.5 km thick (Kent et al, 1978). However, there is evidence that the gabbros beneath the basalts in ocean crust also contribute to the observed anomalies (this is fully discussed in Kent (op. cit.)). In effect, modellers often choose, almost arbitrarily, combinations of magnetisation and layer thickness for oceanic and continental crust that give results that agree with the observed anomalies. The problems of magnetic modelling, in particular of continent-ocean boundary modelling, are discussed further in Appendix A.

In this thesis, a value of magnetisation of 350 gamma for a 4 km thick layer was frequently used for oceanic crustal models.

The 'amount' of magnetisation this produces is a bit larger than the 500–1,500 gamma (5- 15×10^{-3} gauss) in an 0.5 km thick layer mentioned above. On the other hand, this value seems to give reasonable fits with anomalies due to topography in the study area. Also, in this thesis continental crust is assigned a magnetisation of about 150 gamma and when the continent-ocean boundary is modelled, it is the magnetisation <u>contrast</u> between oceanic and continental crusts which produces the observed anomaly.

In general, magnetic modelling is very poorly constrained. Models can be quite spurious if geological and topographic information from seismic profiles is not incorporated. In practice, gravity and magnetic modelling often go hand in hand to produce an 'optimum' solution.

CHAPTER 3

SOUTHERN PORCUPINE SEABIGHT, SOUTHERN PORCUPINE BANK AND PORCUPINE ABYSSAL PLAIN

The objectives of this chapter are to

a) determine the crustal thickness beneath southern Porcupine Seabight

b) describe the possible geology and structural history of southern Porcupine Seabight

c) locate the position of the continent-ocean boundary to the west of Porcupine Bank and Porcupine Seabight

d) give a date for earliest oceanic spreading in Porcupine Abyssal Plain

e) describe the geology and crustal structure of southern Porcupine Bank and the thinned continental crust between southern Porcupine Bank and the continent-ocean boundary

Roberts and Jones (1975) magnetic anomaly map of the area is shown in Fig. 3.1.

3.1 SOUTHERN PORCUPINE SEABIGHT

Southern Porcupine Seabight is here defined as that part of the Porcupine Seabight which lies to the south of 51° 15'N. Marginal trends are predominately NE-SW or ENE-WSW, as opposed to the N-S trends of the Porcupine Seabight trough to the north.

Water depths over the basin are 2,000-3,000 metres. The magnetic field over the southern Porcupine Seabight is fairly quiet and mostly negative (0- -150 gamma). The free air gravity anomaly field over the basin is at present poorly defined (Chart 2). It also appears quiet, and close to zero mgal.

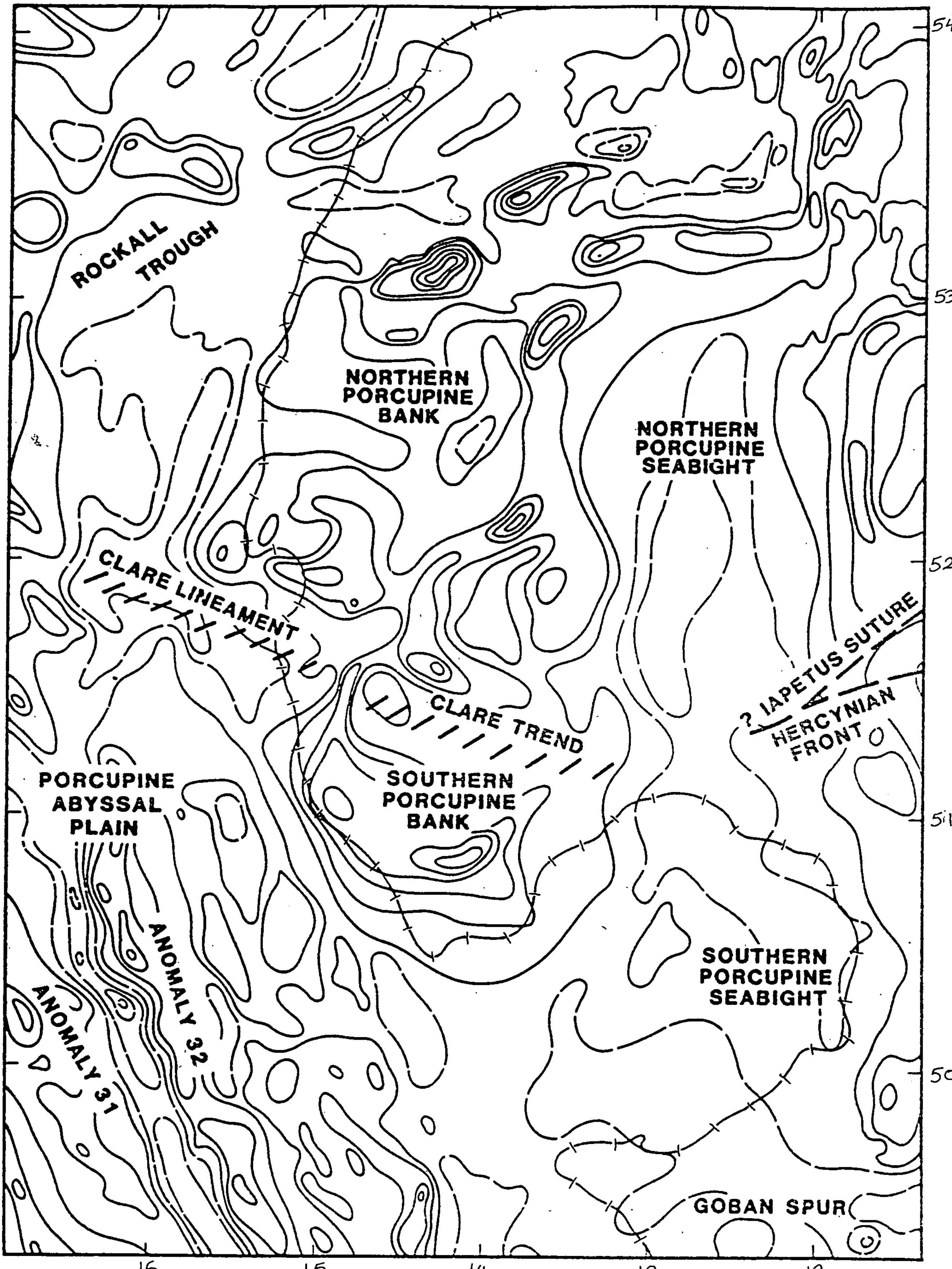


Fig. 3.1 Magnetic anomaly map of Porcupine Bank, Porcupine Seabight and eastern Porcupine Abyssal Plain (Roberts and Jones, 1975) Iapetus Suture, Clare Lineament, and Clare Trend : this thesis Hercynian Front : Bailey, 1975
Contour interval : 100 nT. Solid contours are positive anomalies, dashed negative. —I—I— 2000 m. isobath

3.1.1 Structural Position

a) The Iapetus Suture

The Iapetus Suture has been mapped across Ireland and Britain (e.g. Phillips et al., 1976). Refer to Fig. 3.2. This suture marks the position of the welding of two separate plates in the Caledonian orogeny, and therefore the site of a distinct change in pre-Devonian geology.

Offshore Ireland there exists a NE-SW magnetic lineation on strike with the Iapetus Suture. This has been tentatively identified as the Iapetus Suture and is shown in Fig. 3.1.

b) The Hercynian Front

The position of the Hercynian Front from Bailey, 1975 is shown on Fig. 3.1. This Front represents the northern limit of thrust faulting associated with the Hercynian Orogeny. As such it is a transitional zone, and it is expected that there is no major change is crustal properties across it; the crust to the north was also deformed by the Hercynian orogeny.

c) The Munster Basin

In southern Ireland, to the south of the Iapetus Suture up to 5 km of Devonian of Old Red Sandstone facies is developed in the Munster Basin (Fig. 3.3). The basin is on strike with the southern Porcupine Seabight, and shows the same NE-SW trends. Old Red Sandstone basins in the British Isles are commonly associated with strike-slip faulting along Caledonian faults e.g. the Orcadian Basin. It has been suggested that southern Porcupine Basin and the Munster Basin were both formed by Late Caledonian wrench faulting (P. Bennett, pers. comm.).

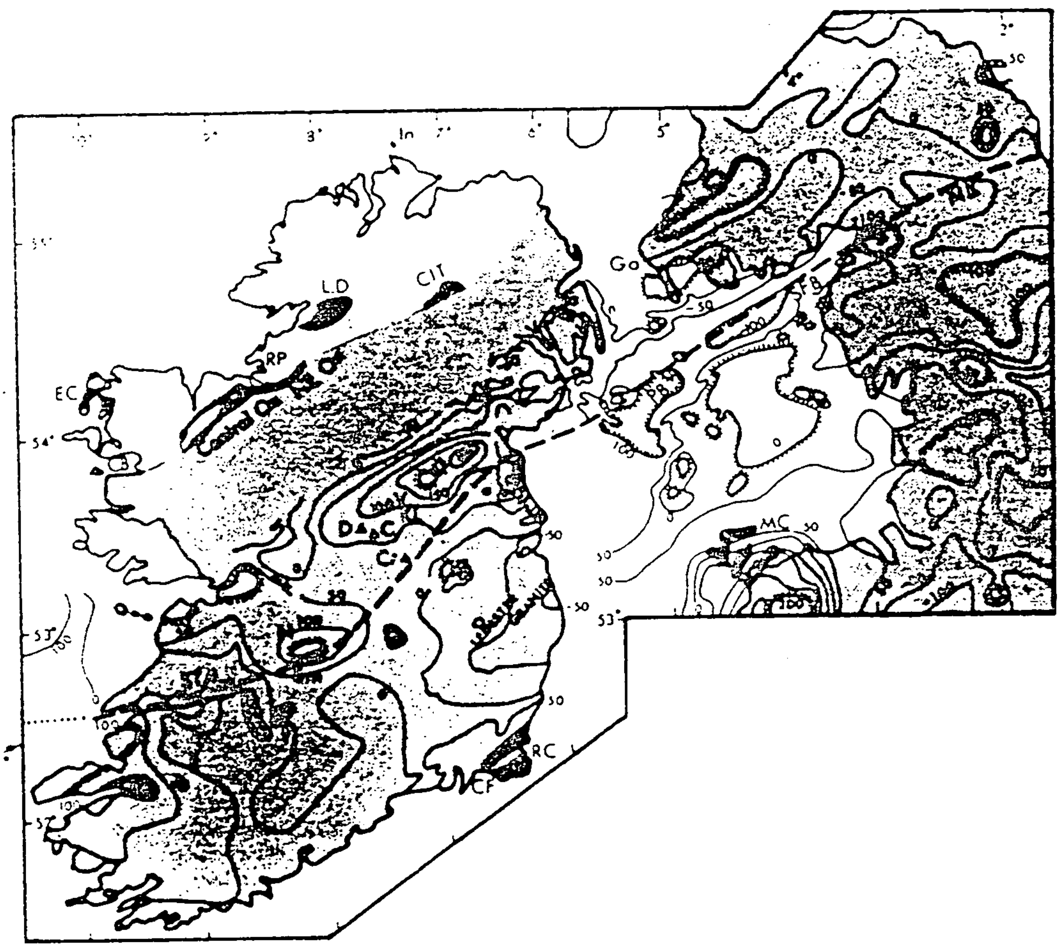


Fig. 3.2 The position of the lapetus suture in the UK and Ireland (Kennan et al, 1979). The lapetus suture is represented by the thick broken line. Contours are regional magnetic anomalies.

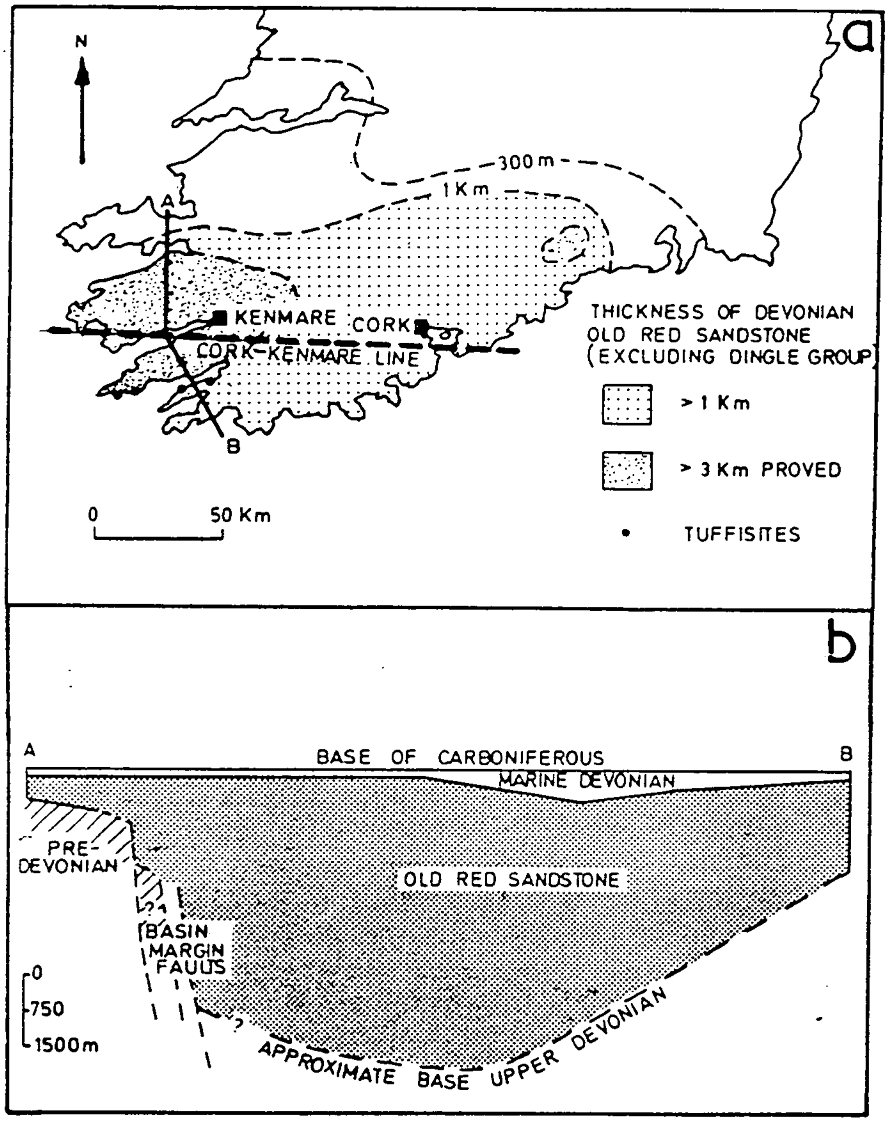


Fig. 3.3 The Munster Basin, SW Ireland (Naylor et al, 1981)

d) Permo-Triassic Rifting

There are no wells drilled in the southern Porcupine Seabight. Authors with access to the results of wells from northern Porcupine Seabight (e.g. Naylor and Mounteney; Ziegler, 1981) describe Permo-Triassic rifting.

With Permo-Triassic basins wide spread to the south (Fastnet, South-West Approaches) and north, and in the North Sea, it would be surprising if there was no Permo-Triassic rifting in southern Porcupine Seabight (Fig. 3.4).

3.1.2 Available Data

a) Reflection Seismic

Fig. 3.5 is a time to basement map. Shackleton 3/79 line 1 was modelled. Roberts et al.,(1981 (a)) published a time to basement map of Goban Spur (Fig. 3.6) from which the 7.0 seconds time was taken for Fig. 3.5.

b) Refraction Seismic

Two refraction seismic lines have been shot on the eastern margin of the southern Porcupine Basin (Scrutton, 1971; Handley, 1971), Figs. 1.9 and 1.10.

With these limited data, approximate crustal thicknesses for the basin centre have been calculated. The basement structure and modelling of the western margin is presented in a later section.

3.1.3 Crustal Thickness Calculations

It was assumed that the crust and sediments of southern Porcupine Seabight are in isostatic equilibrium, as the gravity field is close to zero mgal. Sediment velocities from the refraction experiments were converted to densities using Ludwig, Nafe, and Drake's relationship between velocity and density (Fig. 3.7).

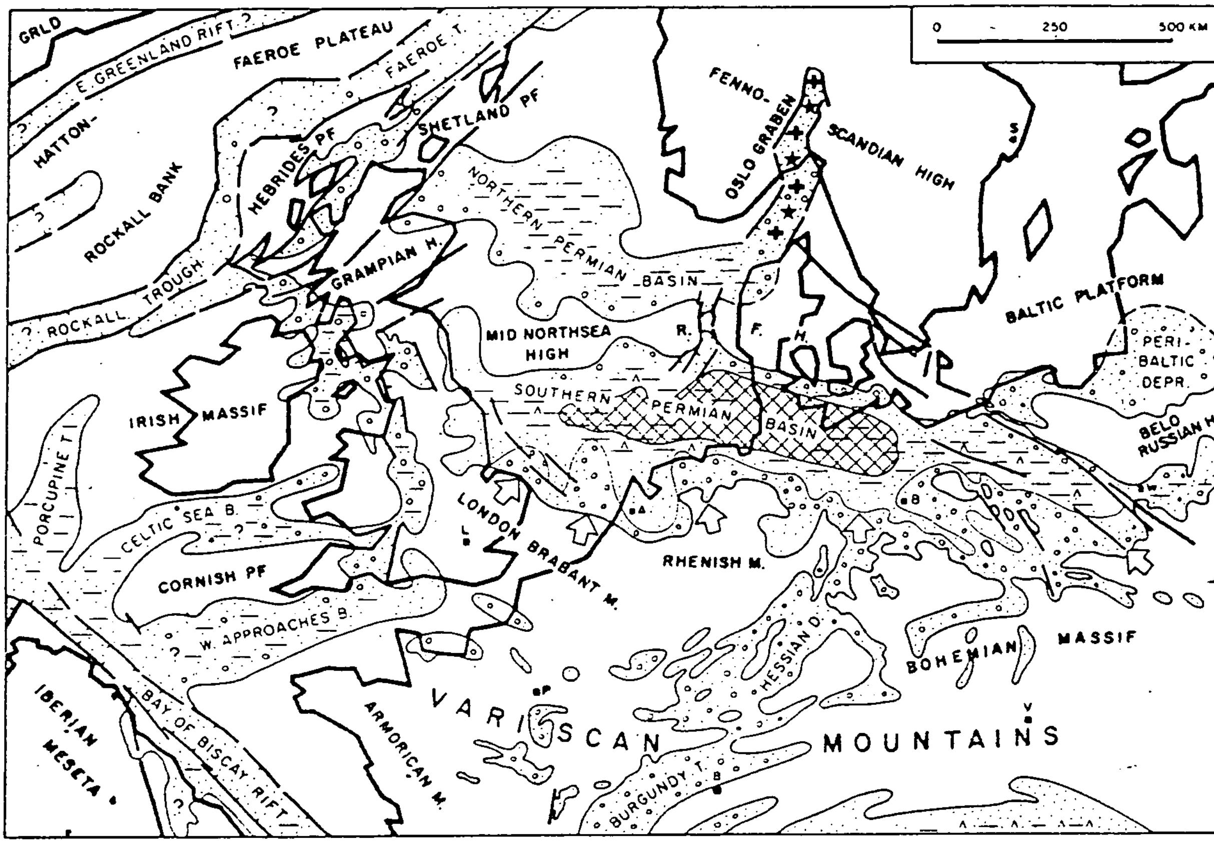


Fig. 3.4 Early Permian Rotliegend palaeogeography, from Ziegler, 1981

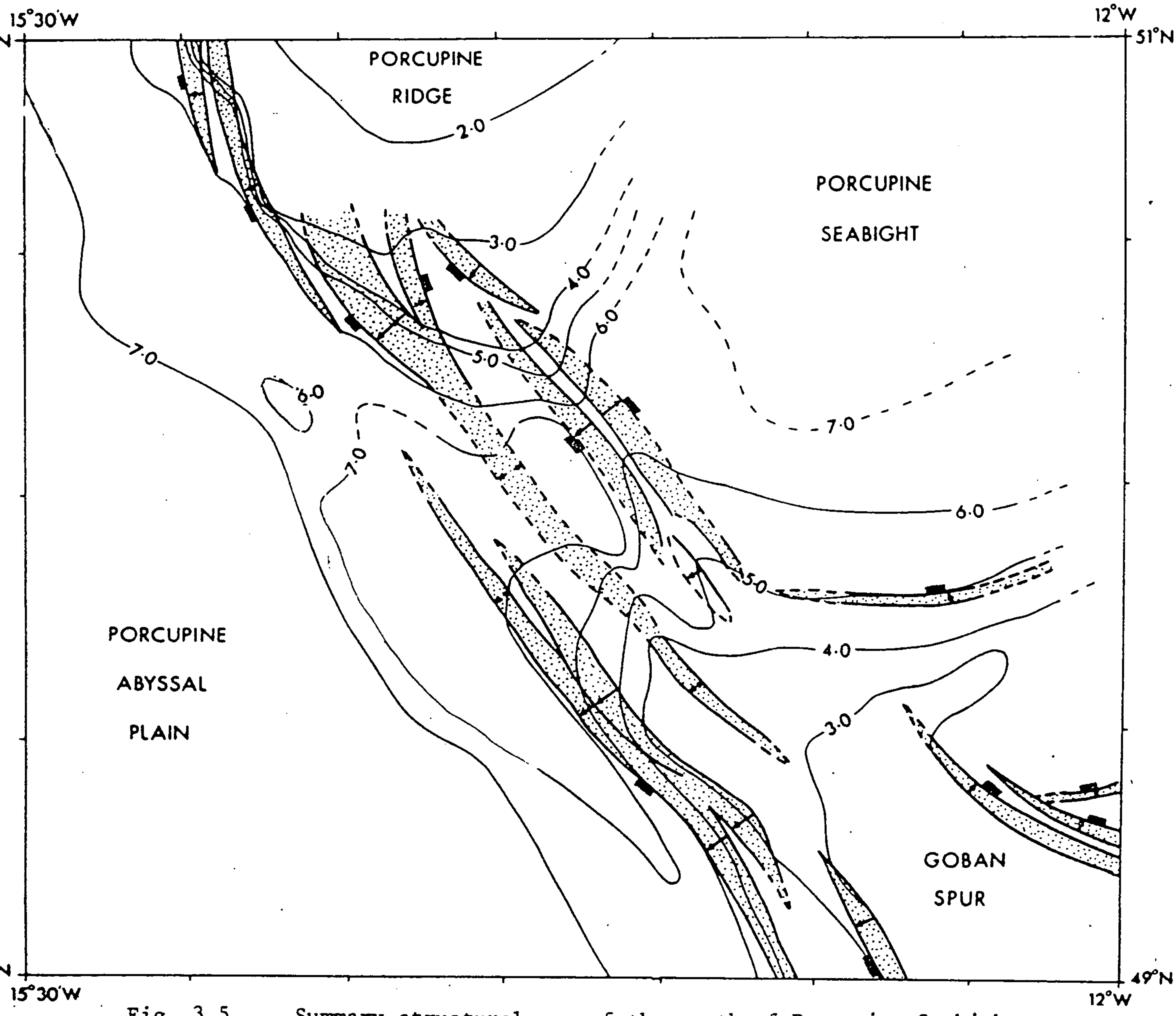


Fig. 3.5 Summary structural map of the mouth of Porcupine Seabight. Values on contours are two way times to basement.

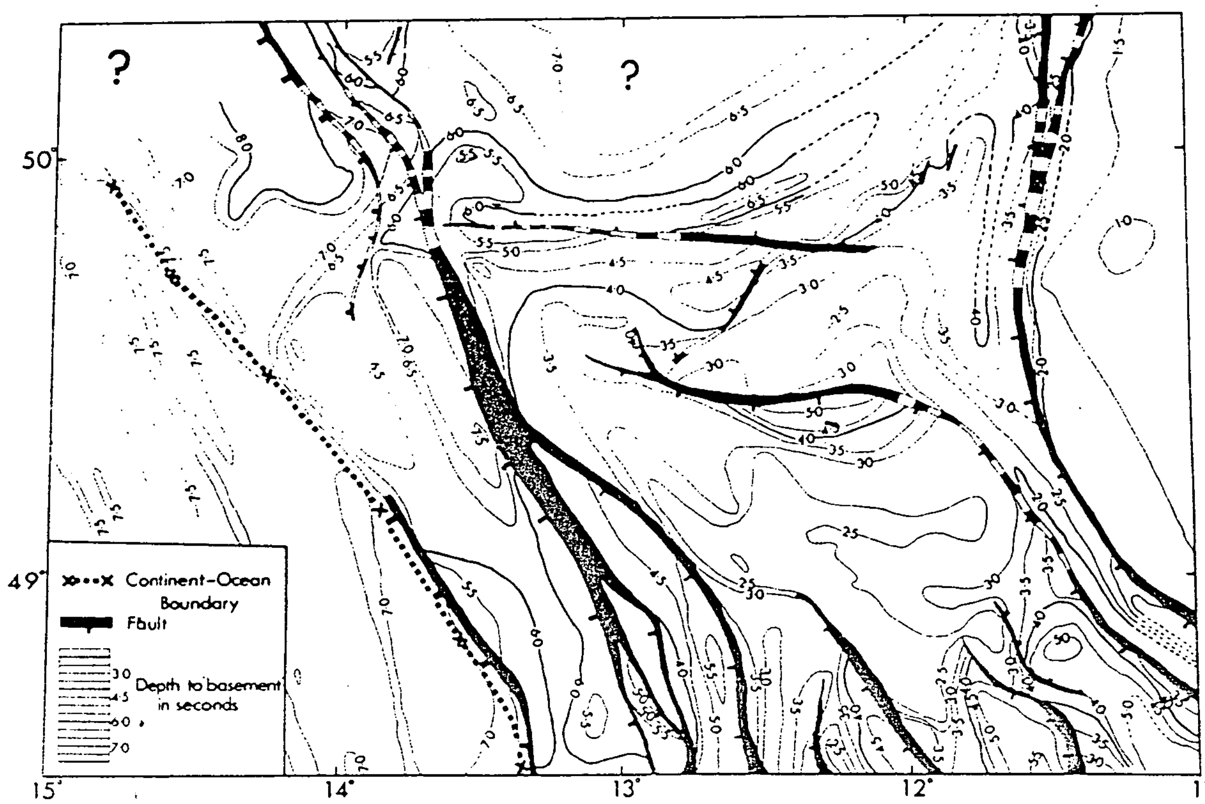


Fig. 3.6 Tectonic pattern of the Goban Spur, Roberts et al.,(1981(a))

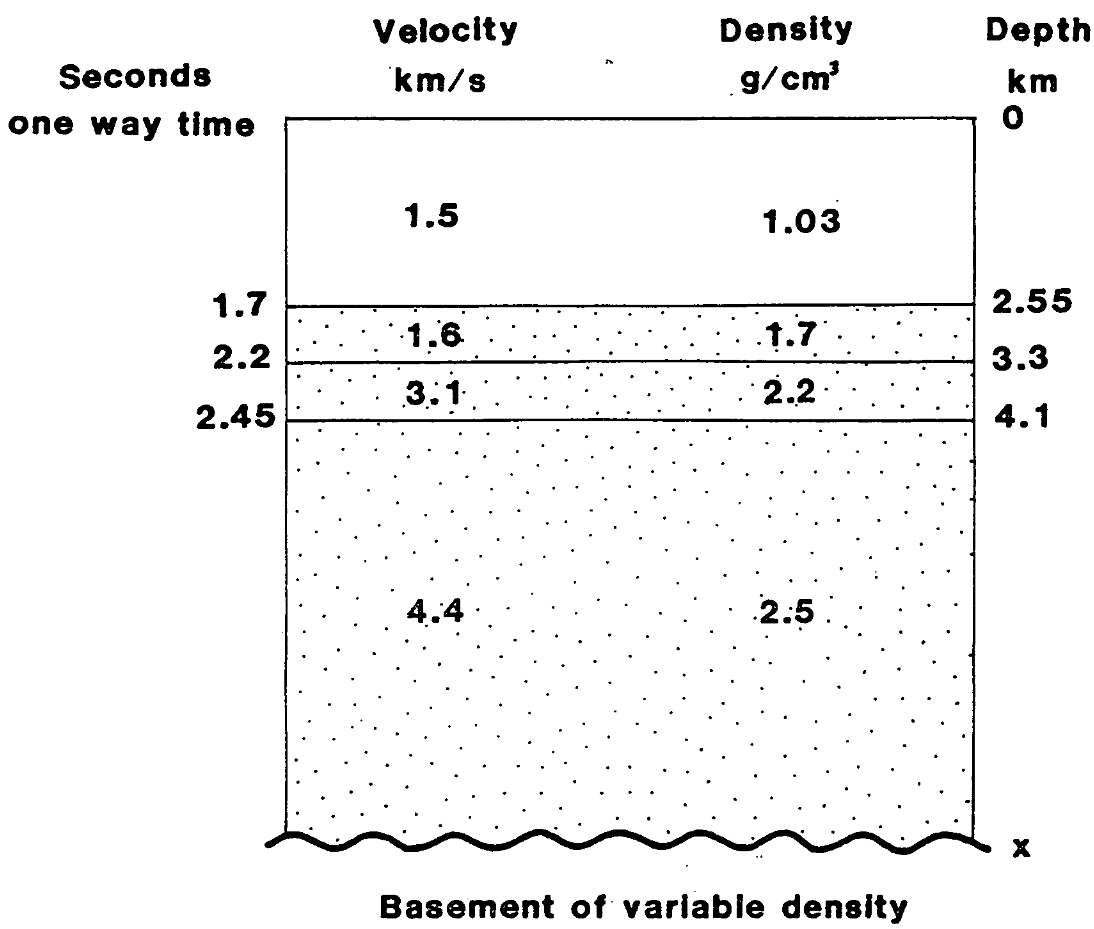


Fig. 3.7 Velocity/depth structure and density/depth structure for the sedimentary column in southern Porcupine Seabight

Two two-way times to basement were chosen; 7 seconds and 8 seconds, which are equivalent to basement depths of approximately 9 km and 11 km respectively.

Crustal density was treated as a variable, and crustal thicknesses for varying densities calculated for the two crustal depths (Fig. 3.8). The calculation was made by using the crustal model in Fig. 2.8.

Errors will be introduced by the conversion from velocities to densities. Furthermore, the crustal thicknesses were calculated for a deeper part of the Seabight (2.55 km water depth) than the eastern area where the refraction line was positioned.

The results show that if basement is as deep as 11 km (8 seconds), the crust under the centre of the basin must be very thin or absent.

A crustal density of 3.05 g/cc was estimated by Brooks and Chinner (1975) for the axial zone of the East African Rift. This is the highest crustal density described in the literature, and gives an upper limit to the crustal thicknesses in fig. 3.8.
Le Pichon (1981) found continental crustal thicknesses of 2-4 km in the Armorican Basin, which he assumed was due to two phases of extension.

3.1.4 The Structure of the Mouth of Porcupine Seabight

a) Basement Ridges

Fig. 3.5 is a time to basement map produced from Edinburgh data.

Areas of deeper basement are mapped on Roberts et al.'s map (Fig. 3.6) as this was based on multichannel data. Basement is invisible under thick sedimentary cover on the Edinburgh lines (Fig. 3.9).

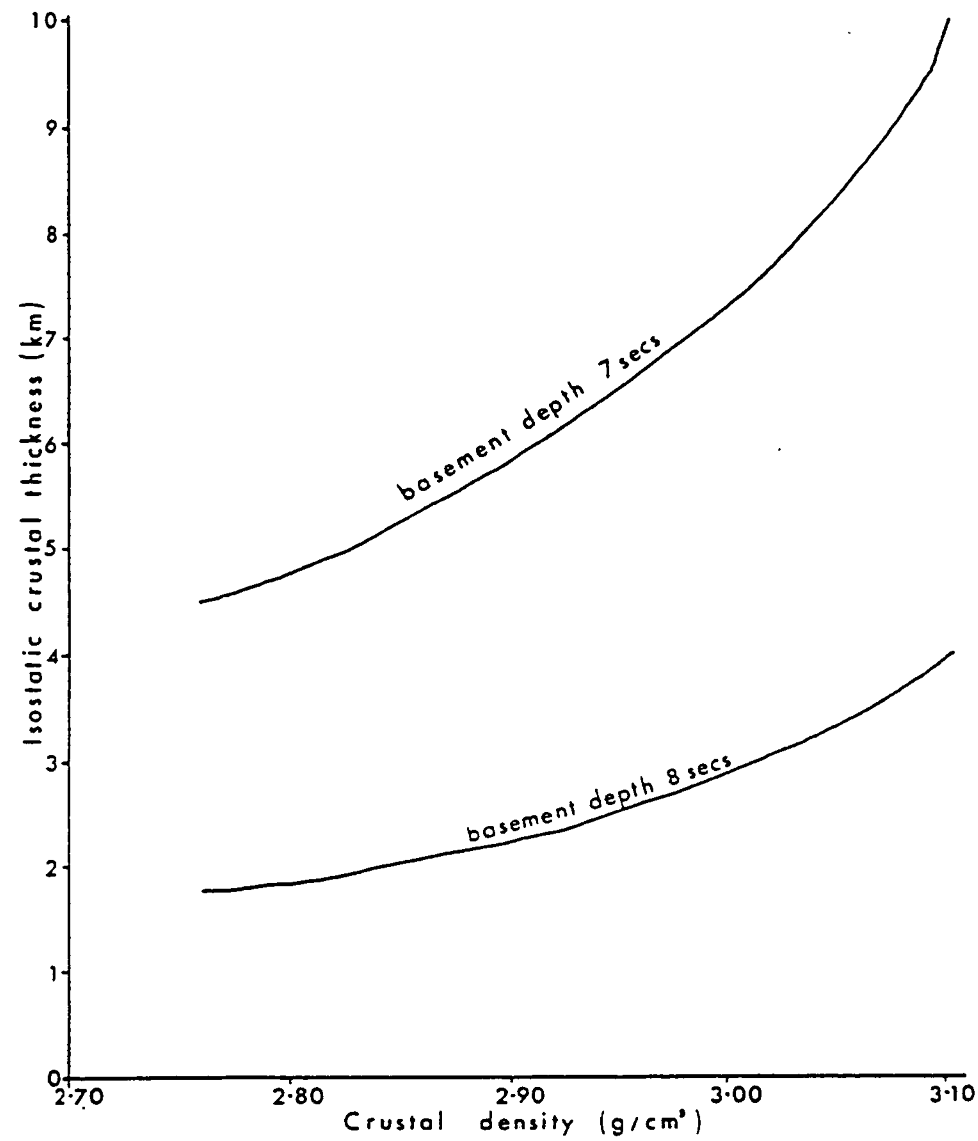


Fig. 3.8 Graph of isostatic crustal thickness vs crustal density for two different depths of crust in southern Porcupine Seabight. A time of 7 sec. (TWT) is assumed to be approximately equivalent to a depth of 9 km, and a time of 8 secs. equivalent to 11 km.

The major difference between the two maps is the NE-SW trend of a fault to the west of Goban Spur on Roberts et al.'s map. On the author's map this fault is presumed to continue to the NW under the mouth of Porcupine Seabight. This pattern of Goban Spur scarps foundering in the mouth of Porcupine Seabight but continuing on the same trend was arrived at after inspection of Edinburgh Goban Spur seismic lines.

Fig. 3.10 illustrates the splitting and foundering north-westwards of the basement high at the margin of Goban Spur.

The continuity of structure across the mouth of Porcupine Seabight from Goban Spur to Porcupine Bank does indicate a continuity of basement type. It would be impossible to suggest that oceanic crust in Porcupine Seabight could show the same continuity of structure.

This evidence for continental crust underlying the southern Seabight is supported by the flat magnetic field, and magnetic modelling of Shackleton 3/79 line 1.

b) Modelling of Shackleton 3/79 line 1

Gravity

Errors in the model were introduced due to

i) Invisibility of basement below six or seven seconds TWT

ii) The assumption that the hinge zone is underlain by a single solid basement ridge at approximately 5 seconds. The complex diffraction pattern under the hinge zone shows that instead this feature is pervasively faulted and broken up.

iii) the fact that two-dimensionality cannot be used. The eastern end of Shackleton 3/79 line 1 lies close to Goban Spur. Scrutton (unpub.) produced a 3-D gravity model of the broad features of the mouth of Porcupine Seabight. The gravity anomaly produced along the line of line 1 due to Goban Spur and Porcupine Bank was subtracted from the observed model, with the result that the line could be

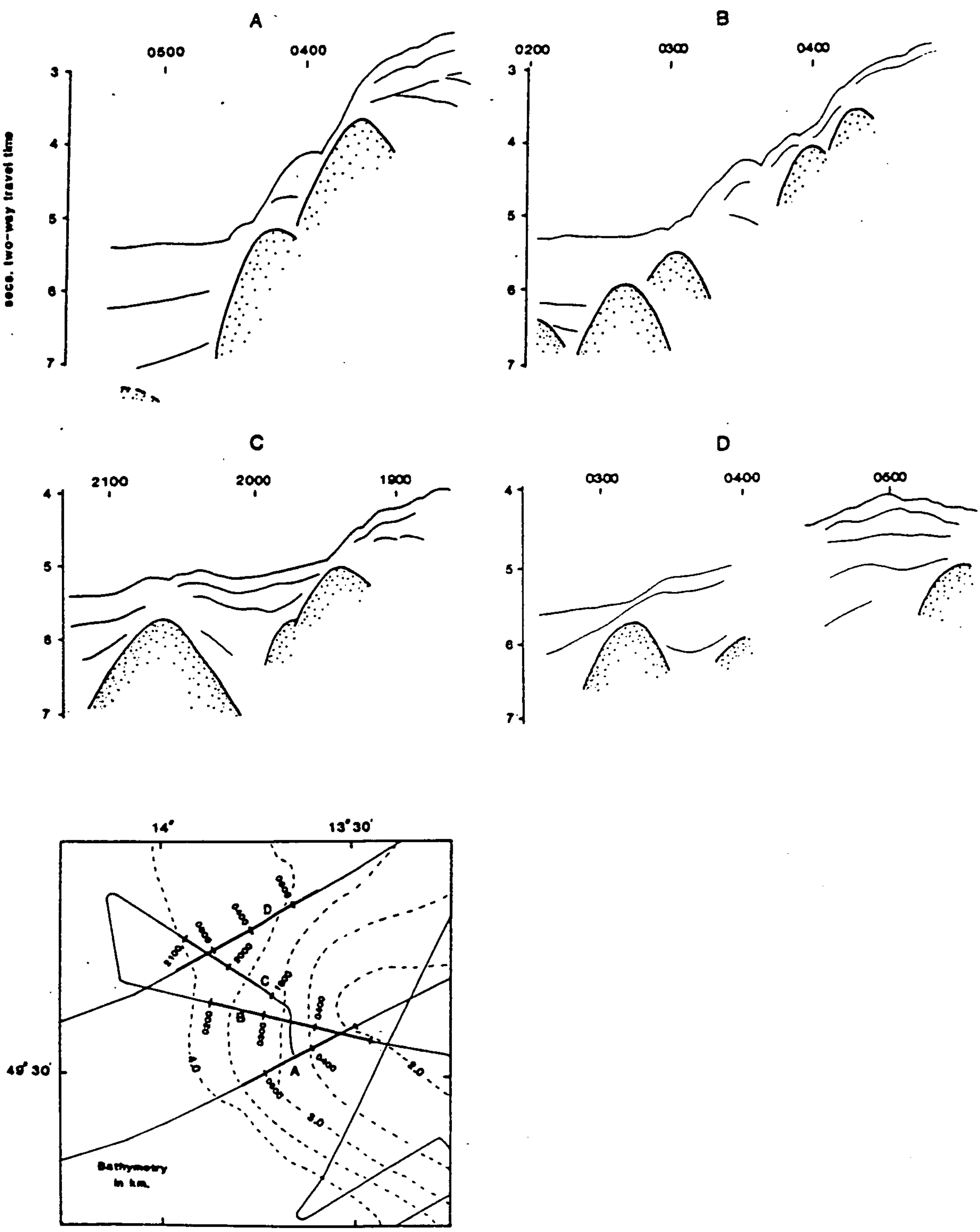


Fig. 3.10 The foundering of basement scarps from NW Goban Spur into Porcupine Seabight. Edinburgh seismic data

modelled in two dimensions.

The isostatic model (Fig. 3.11) shows close agreement with the observed values minus the 3-D values. Therefore the mouth of Porcupine Seabight is in isostatic equilibrium.

Magnetics

It proved difficult, with basement depths poorly constrained, to model the continent-ocean boundary on Shackleton line 1.

It was decided, from comparison of seismic character with Shackleton lines to the north, that boundary B at 0430 Z/090 is the most likely continent ocean boundary. Boundary A, which was initially modelled as the continent ocean boundary, is now interpreted as the junction between deeper continental crust to the east, and a ridge of more magnetic continental crust to the west.

3.1.5 Conclusions

The observed structural continuity from Goban Spur, across Porcupine Seabight, to Porcupine Bank, implies that Porcupine Seabight must be underlain by continental crust. This is supported by the magnetic anomaly field over Porcupine Seabight.

It is thought probable that the broad history of southern Porcupine Seabight was as follows:

i) Late Caledonian wrench faulting producing Devonian and Carboniferous basins

ii) Permo-Triassic basin formation

iii) Intermittent faulting throughout the Jurassic and the Cretaceous until the Campanian, when spreading started to the west of Porcupine Seabight.

iv) Early Tertiary igneous activity and Late Tertiary subsidence.

These cumulative episodes of crustal stretching produced an unusually thin pre-Devonian crust.

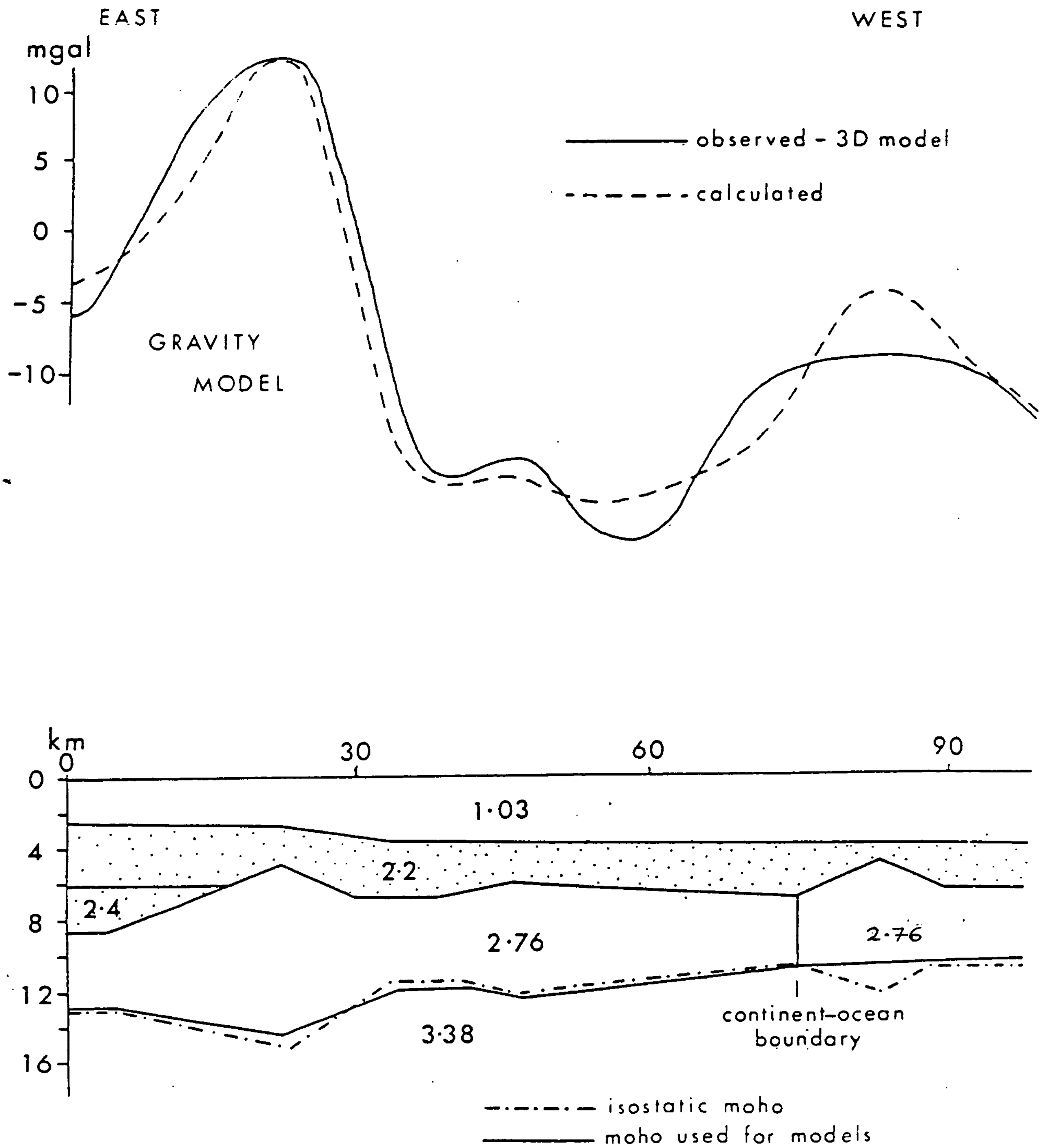


Fig. 3.11 Gravity model along Shackleton line 1 from Porcupine Seabight to Porcupine Abyssal Plain.
Figures are densities in g/cm^3

3.2 SOUTHERN PORCUPINE BANK

Southern Porcupine Bank is defined as that part of Porcupine Bank to the south of the Clare Trend (Fig. 3.1). There is a very limited amount of new data as the eastern parts of the Shackleton 79 lines penetrate only as far as 40 km into Southern Porcupine Bank. This section is limited to a seismic interpretation of these Shackleton lines, and an interpretation of the trans-Atlantic correlation of this part of the Porcupine Bank from magnetic data.

3.2.1 Structural Position and Magnetic Signature

Southern Porcupine Bank is linked to Goban Spur by a high in the mouth of Porcupine Seabight (Fig. 3.5).

Its most prominent feature is a 500 gamma magnetic high over the western part of the Bank. The magnetic high tends NW-SE to WNW-ESE (Fig. 3.1).

a) Trans-Atlantic correlations: the Avalon zone and Iapetus Suture.

The Avalon platform of southern Newfoundland represents the southern margin of Iapetus, formed in the Precambrian and Lower Palaeozoic. (Howarth and Keen, 1979; Howarth and Lefort, 1979; Rast et al. 1976, etc).

A volcanic arc model has been proposed for the Avalon zone (Rast et al, 1976), which consists of thick Precambrian volcanics and associated sedimentary basins (Fig. 3.12). The Avalon zone is separated by the Dover fault, the Iapetus Suture equivalent, from the platform which was to the north of the Iapetus ocean, the Grenville platform. The Avalon zone is characterised by high magnitude sinuous magnetic anomalies which can be traced offshore into the Orphan Basin (Fig. 3.13). These magnetic anomalies are bounded to the north by the Dover fault, and to the south by the Collector anomaly (Haworth and Lefort, 1979). Haworth and Lefort consider that the western limit of the Charlie Gibbs Fracture Zone

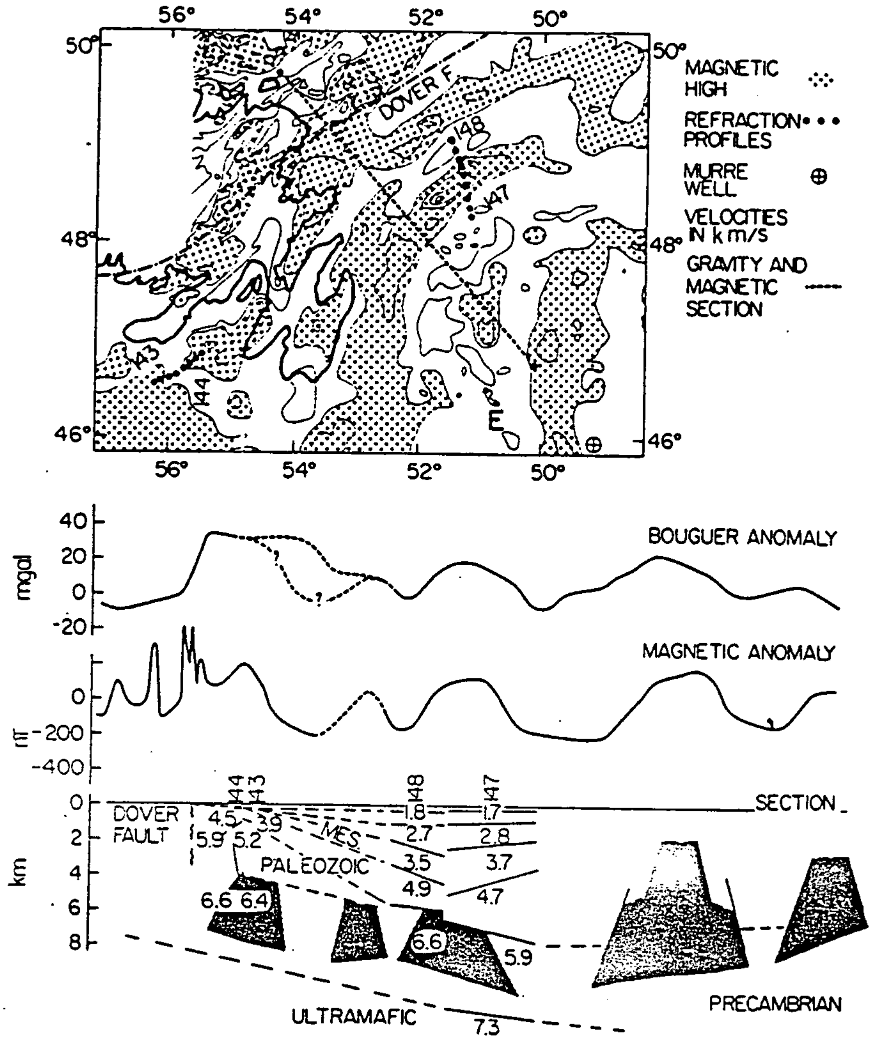


Fig. 3.12 Geophysical and corresponding structural section across the Avalonian structures of the Grand Banks of Newfoundland Magnetic anomalies are contoured at 200 gamma intervals. From Lefort and Haworth, 1979

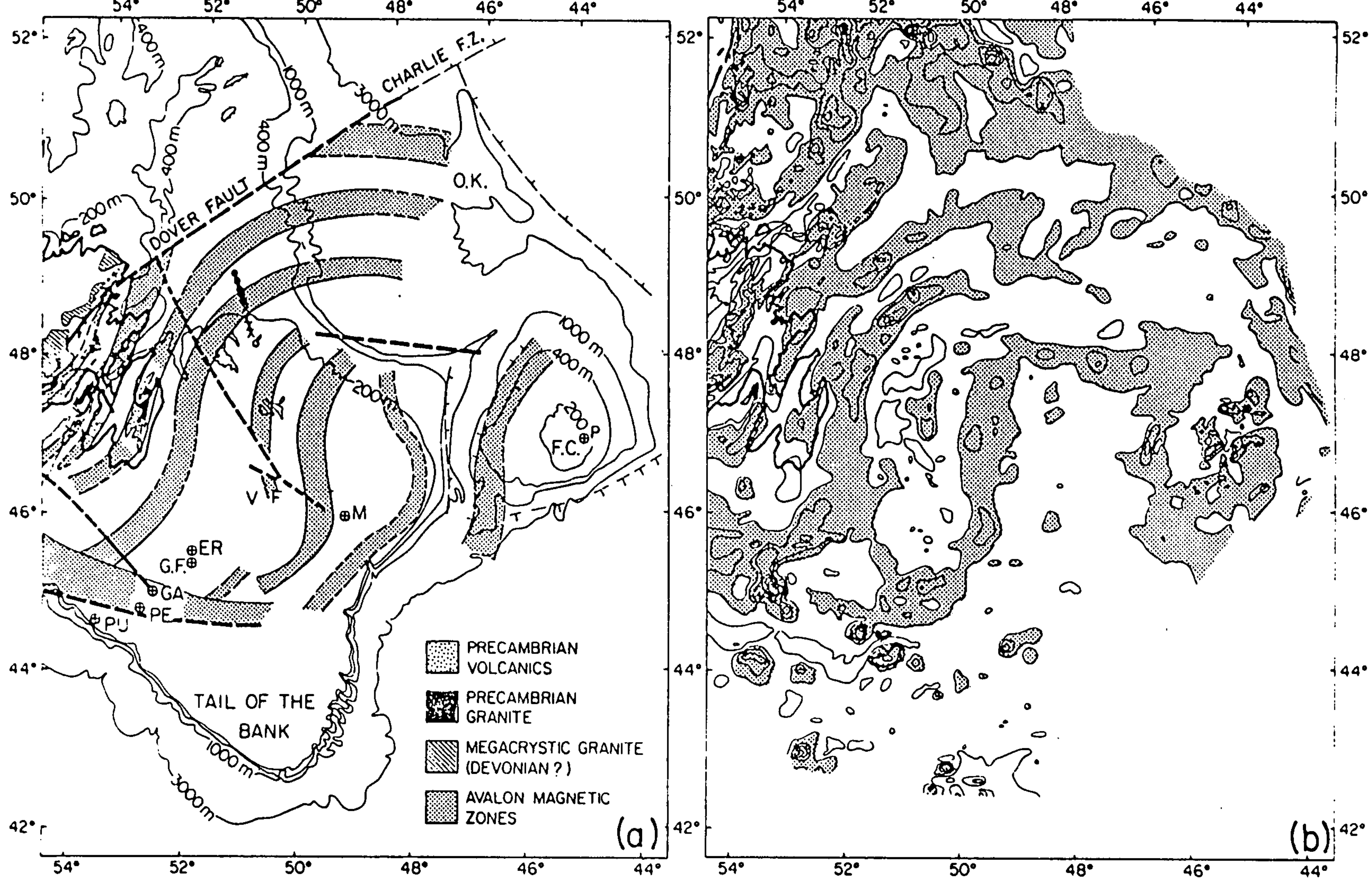


Fig. 3.13 The Avalon Zone and Grand Banks, Haworth and Lefort, 1979.

a) Structural trends

b) Magnetic anomalies

Solid and dashed lines represent faults. Magnetic contour interval is 200 gamma and positive anomalies are shaded.

is the intersection of the Dover Fault with the continent ocean boundary.

On an Atlantic reconstruction, the magnetic anomaly of southern Porcupine Bank is on the trend of the northernmost Avalon magnetic anomaly (Fig. 3.14). It is considered therefore that the junction of the eastern end of the Charlie Gibbs Fracture Zone with continental crust, i.e. the Clare Trend, also marks the position of the Iapetus Suture. (The structure of the Clare Lineament is discussed in greater detail in Chapter 5). The Iapetus Suture crosses Porcupine Bank as the Clare Trend, separating crusts of very different geology to north and south as shown by the magnetic signature and trends of the Bank scarps. The Iapetus Suture divides southern Porcupine Seabight from the northern part, and swings to the NE in Ireland.

3.2.2 Reflection Seismic Data

The Edinburgh seismic lines are supplemented by one other line (Clarke et al., 1971). These lines run from the westernmost part of Southern Porcupine Bank into Porcupine Abyssal Plain (Fig. 3.15).

At the very south of southern Porcupine Bank, basement deepens towards Porcupine Seabight, and is more broken up. Sediment cover is 0.5 seconds to >1.5 seconds, and basement is not exposed. Shackleton line 3 (Fig. 3.16) illustrates this buried basement and grabens within Porcupine Bank.

To the north, basement is exposed at the steeper continental slope. Attempts to correlate these basement 'scarps' from line to line suggest that these scarps are en echelon, and swing round in trend from NW-SE in the south to NNE-SSW in the south. This northerly trend is orthogonal to the Clare Lineament and Clare Trend.

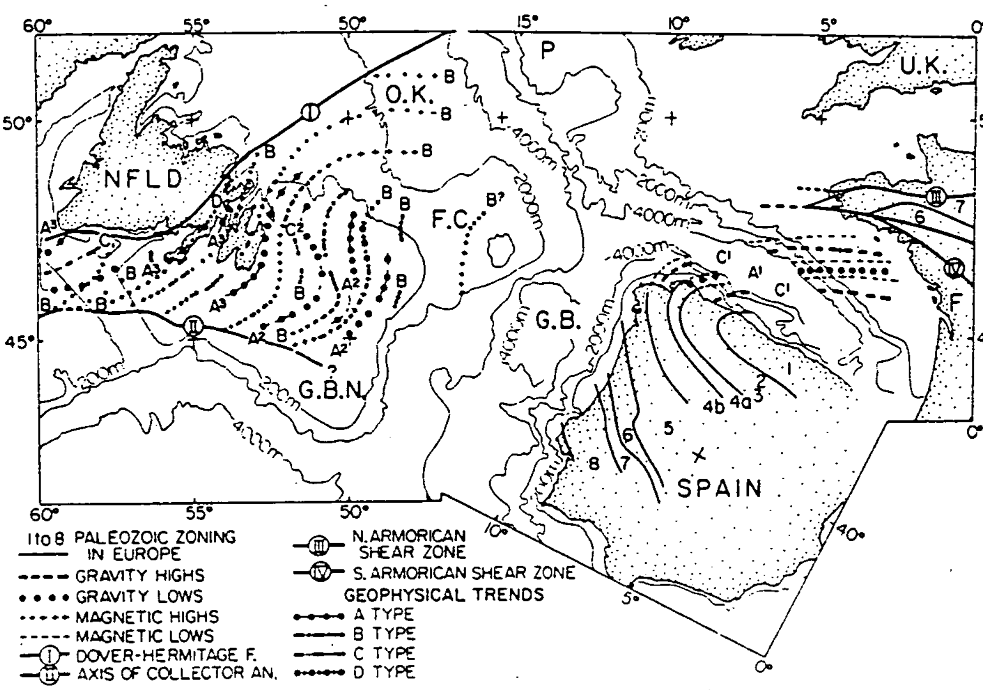


Fig. 3.14 Avalon magnetic zones on an Atlantic refit, from Lefort and Haworth, 1979. The most northeasterly Canadian magnetic high is directly on trend with the southern Porcupine Bank magnetic high

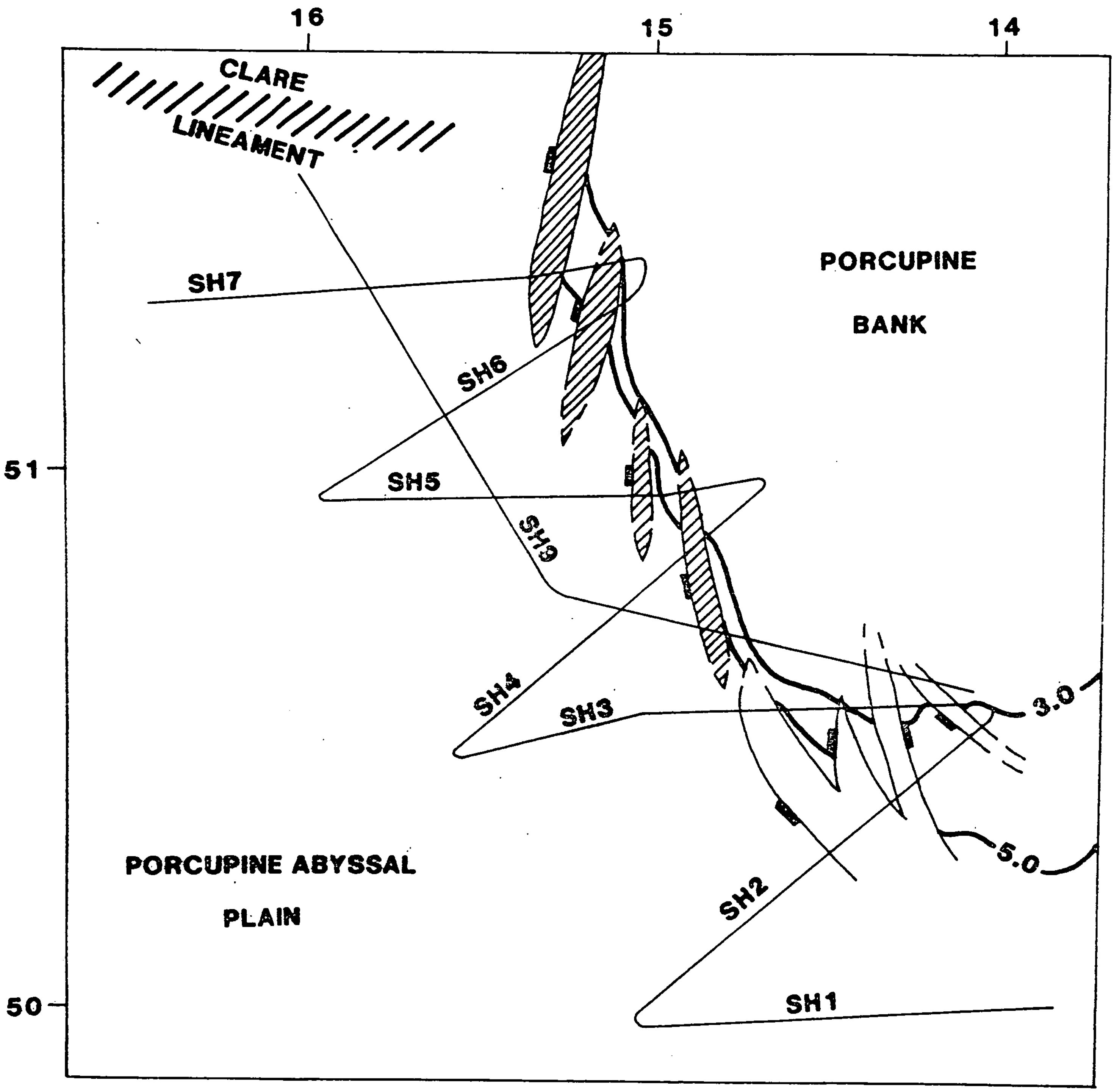


Fig. 3.15 Edinburgh seismic lines (Shackleton 79 cruise) over Porcupine Abyssal Plain and southern Porcupine Bank. The 3.0 seconds (TWT) and 5.0 seconds time to basement contours are shown. The stippled features are exposed basement scarps; unstippled are sediment covered.

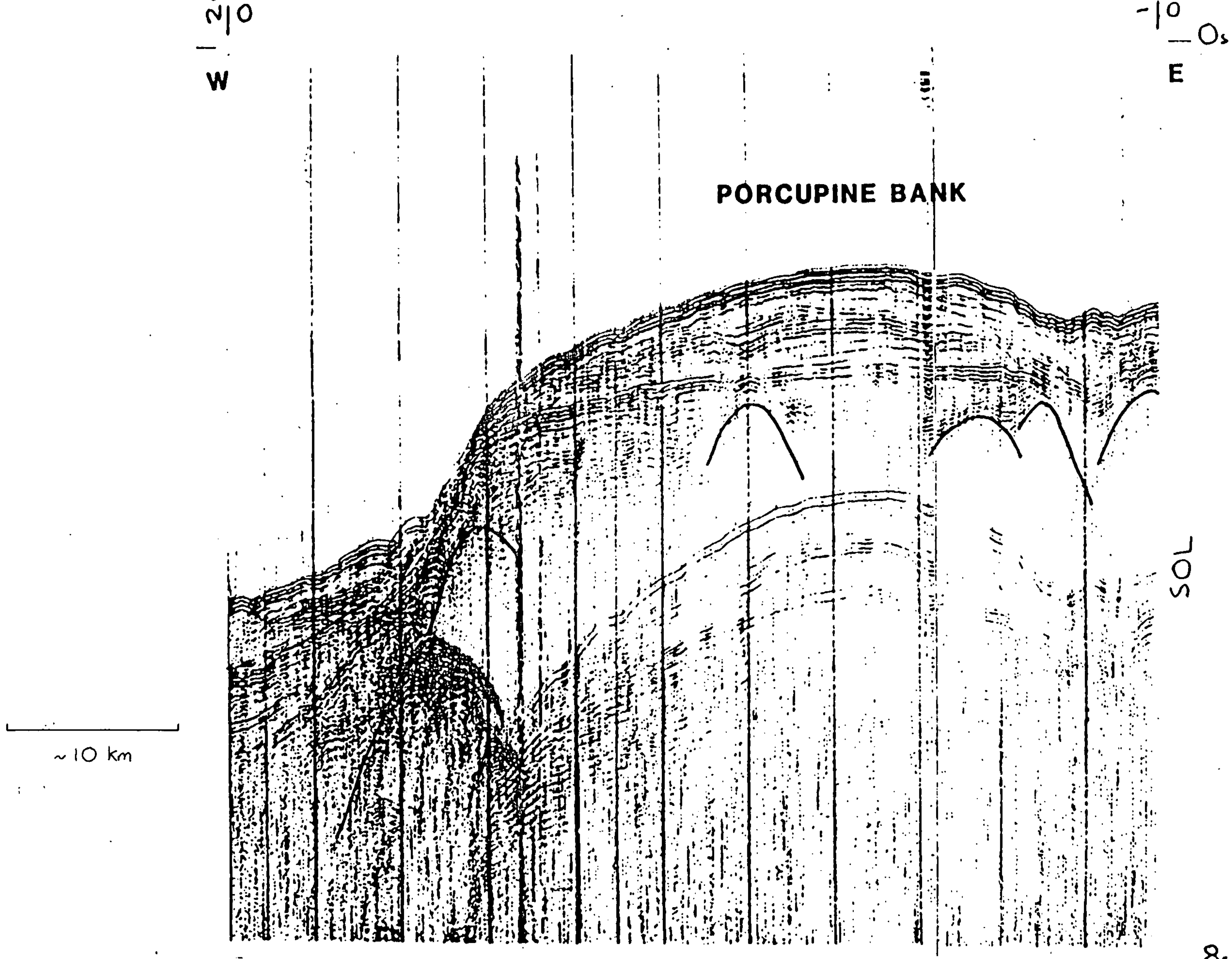


Fig. 3.16 Interpretation of Shackleton line 3; a section over the southern extremity of southern Porcupine Bank. Note slumps at foot of slope. Vertical exaggeration approximately 11x

Shackleton line 6 (Fig. 3.17) shows exposed basement at the continental slope. To the north, Shackleton line 7 (Fig. 3.18) crosses a tilted basement fault block and the main scarp.

3.3 PORCUPINE ABYSSAL PLAIN

3.3.1 Gravity, magnetic, and seismic profiles.

a) Seismic Profiles

The Edinburgh seismic profiles show a graben at the foot of the continental slope, with basement invisible beneath 2 seconds or more of sediment. Beneath Porcupine Abyssal Plain basement highs are common, with up to 2 seconds of topography. The sediment cover is 0-2.5 seconds thick.

A generalised time to basement map is shown in Fig. 3.19.

Basement highs are difficult to correlate from line to line and it is probable that highs are broken up and arranged en echelon, in a similar way to the bounding scarps of the southern Porcupine Bank.

b) Magnetic Profiles

Magnetic anomaly profiles are shown in Fig. 3.20. The Edinburgh profiles do not extend as far as the oldest previously confirmed ocean spreading anomaly i.e. anomaly 32 (34). The cause of the rather complex and uncorrelatable magnetic signature of the area covered by the Edinburgh profiles is discussed in the section on the continent-ocean boundary.

c) Gravity Profiles and a Gravity Model of Shackleton line 4.

Free air gravity profiles are given in Fig. 3.21. A crustal gravity model of Shackleton line 4 was constructed (Fig. 3.22). This model predicts 4 km of sediment in the graben at the foot of the slope. Crustal thicknesses are approximately 22 km under the

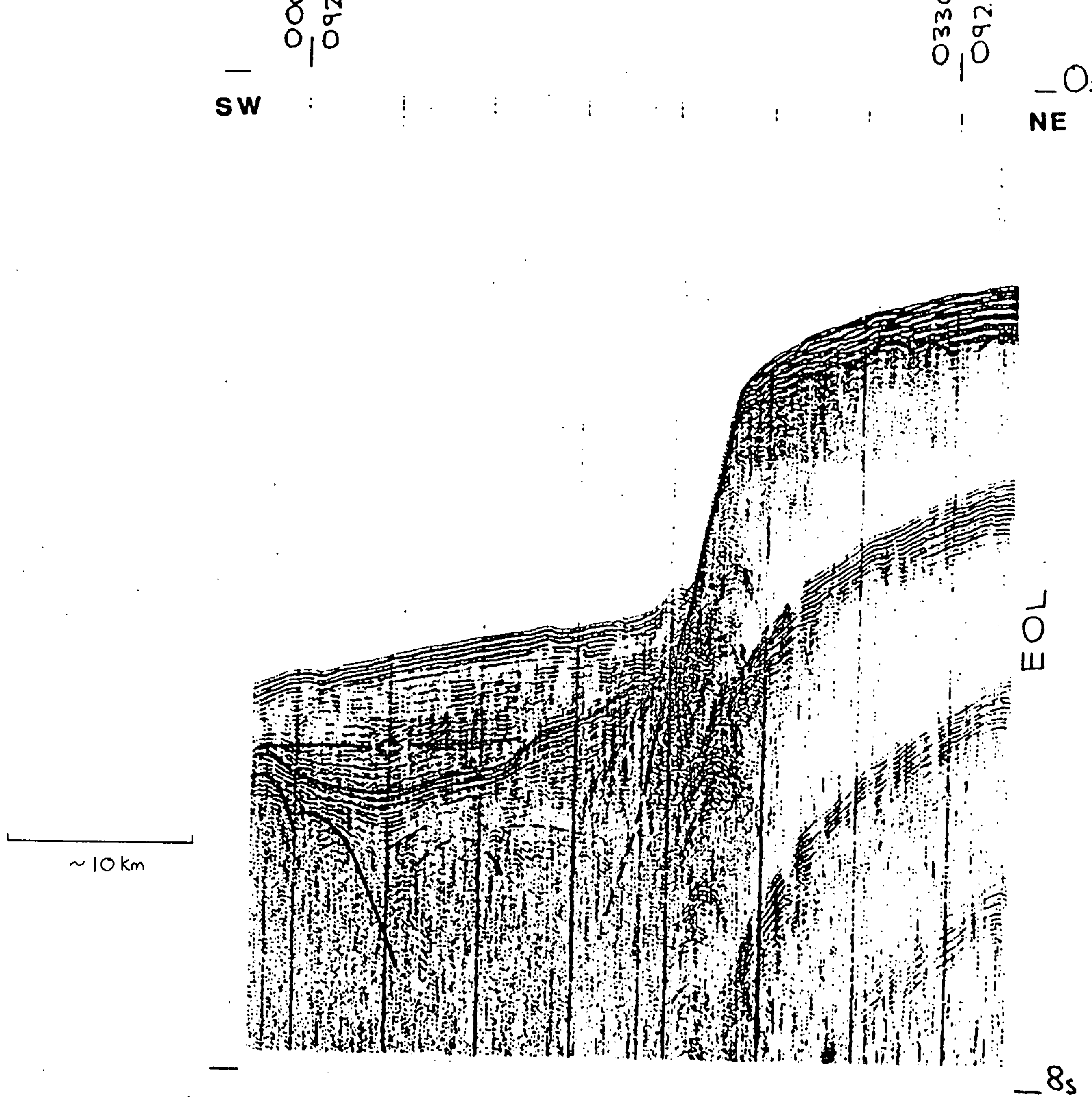


Fig. 3.17 Interpretation of Shackleton line 6 showing exposed basement at the continental slope

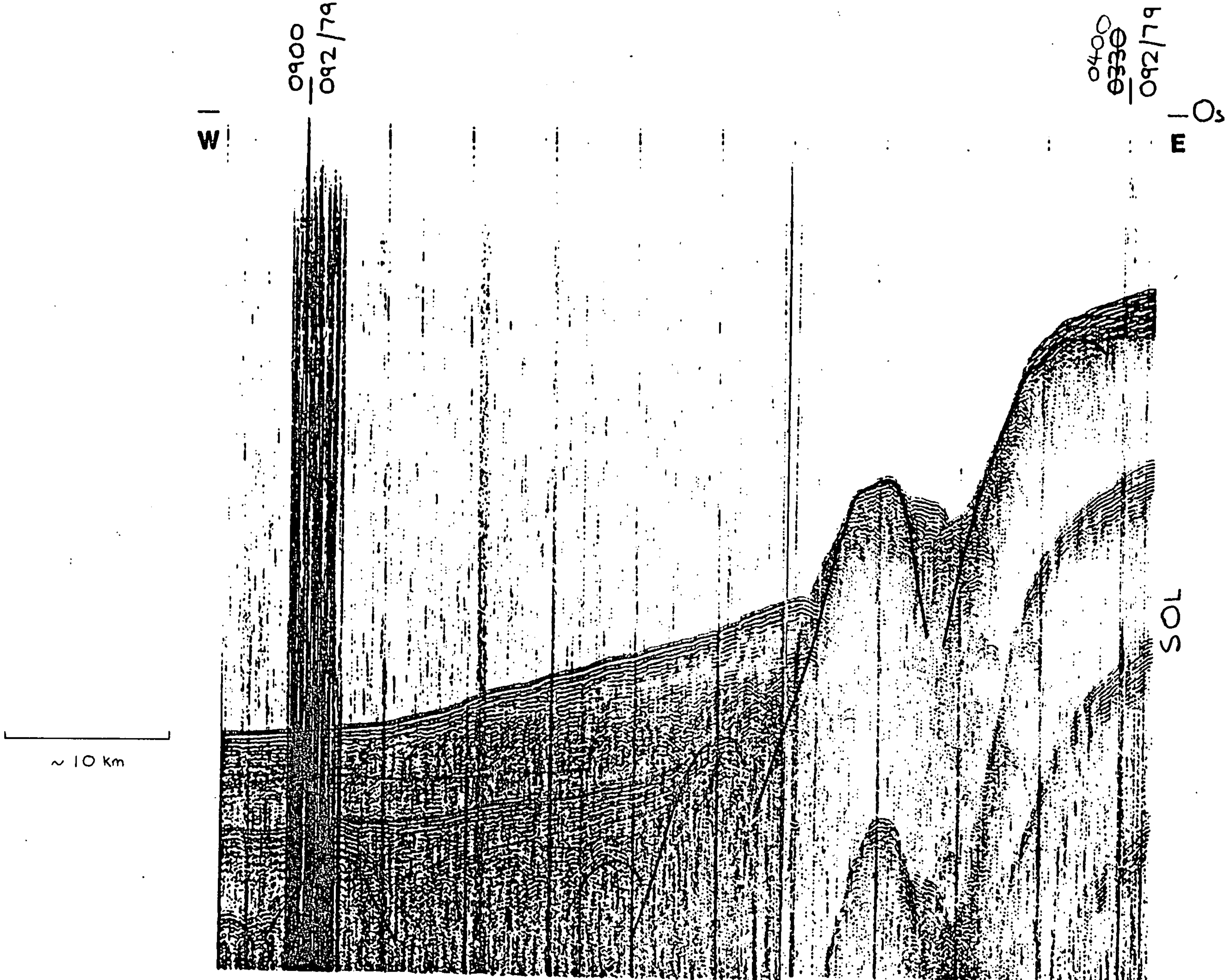


Fig. 3.18 Interpretation of Shackleton line 7

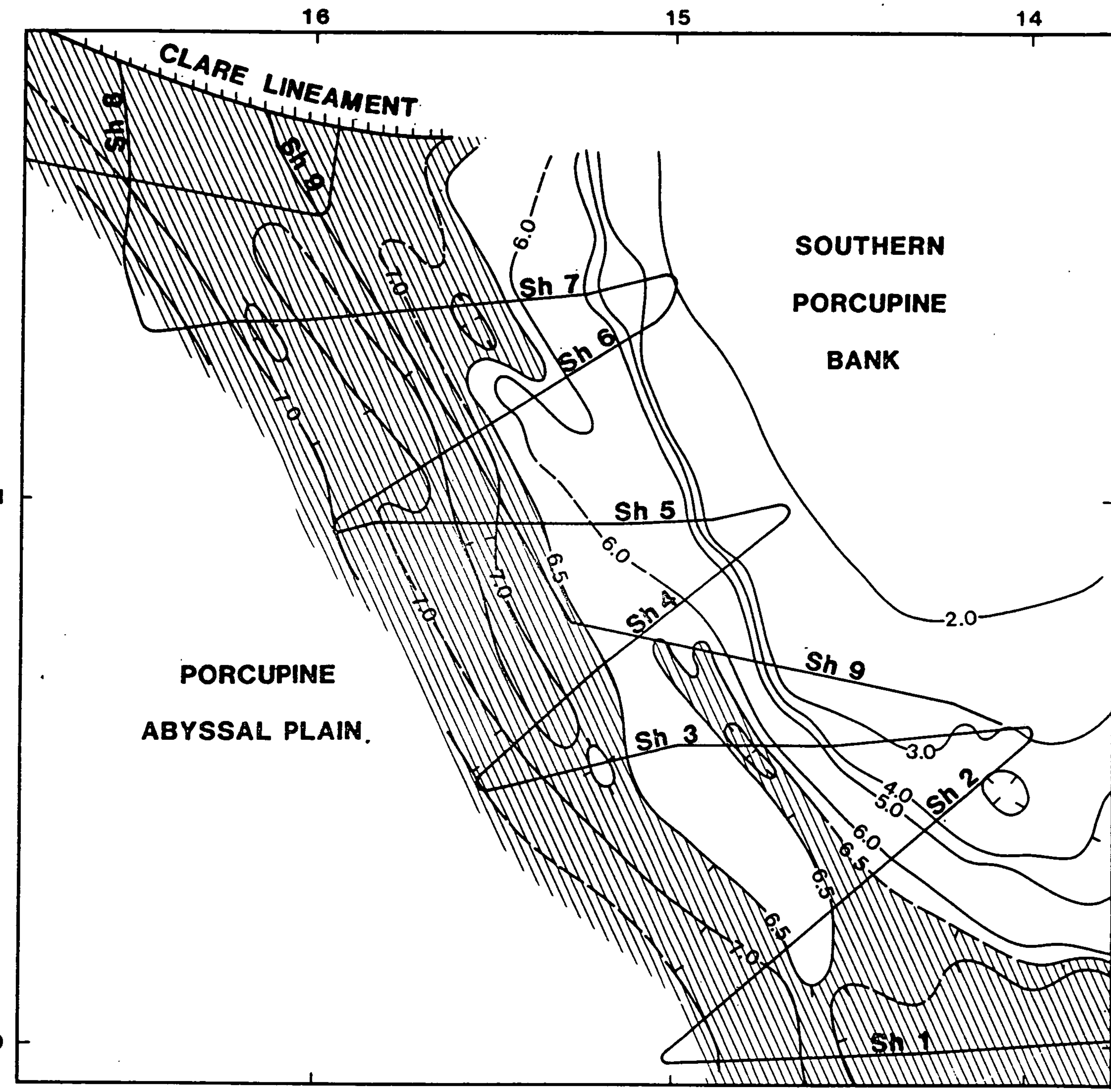


Fig. 3.19 Generalised time to basement from Edinburgh seismic lines. Times to basement adjacent to southern Porcupine Bank are greater than shown as acoustic basement is invisible

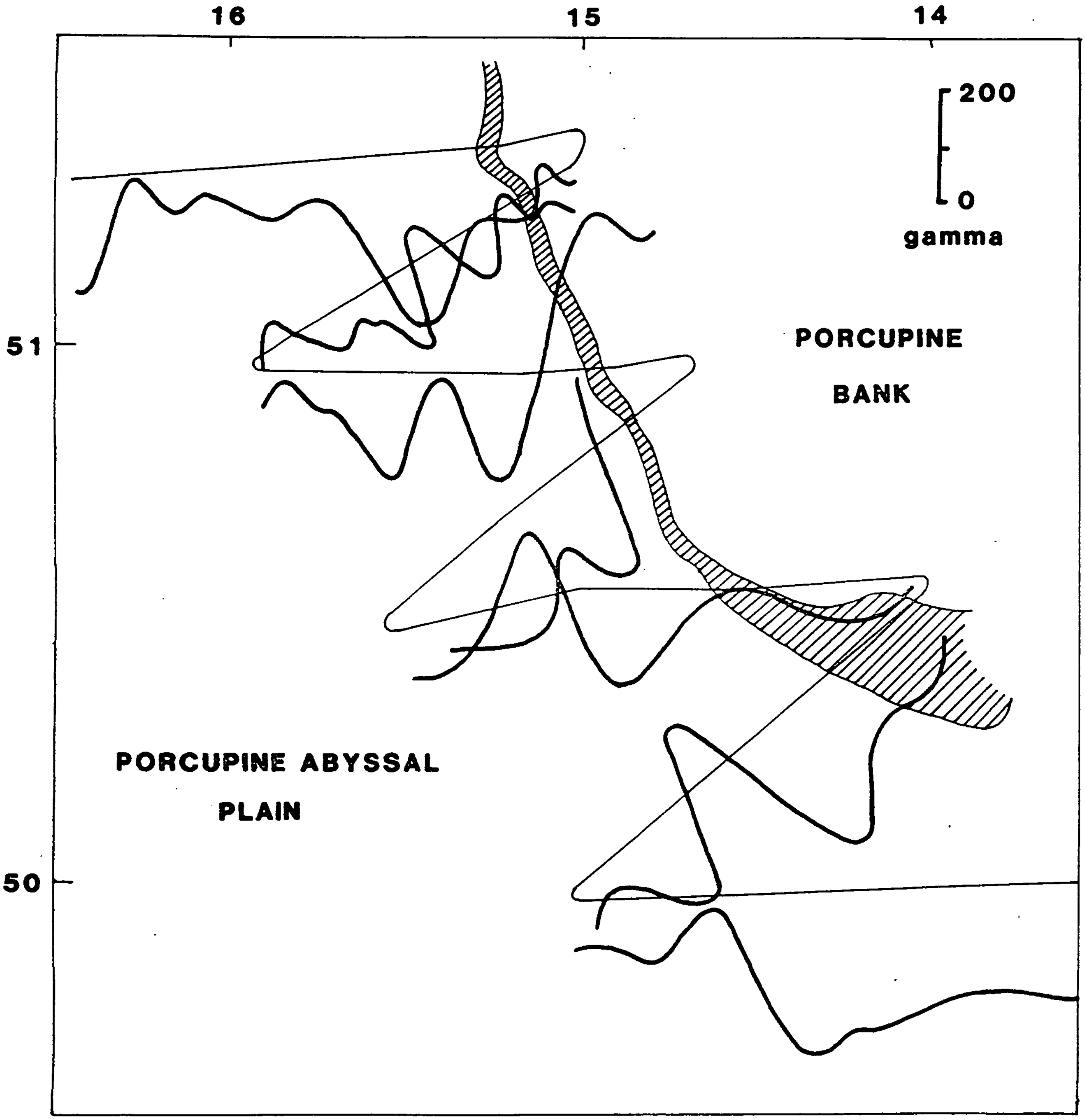


Fig. 3.20 Magnetic anomaly profiles: Edinburgh data

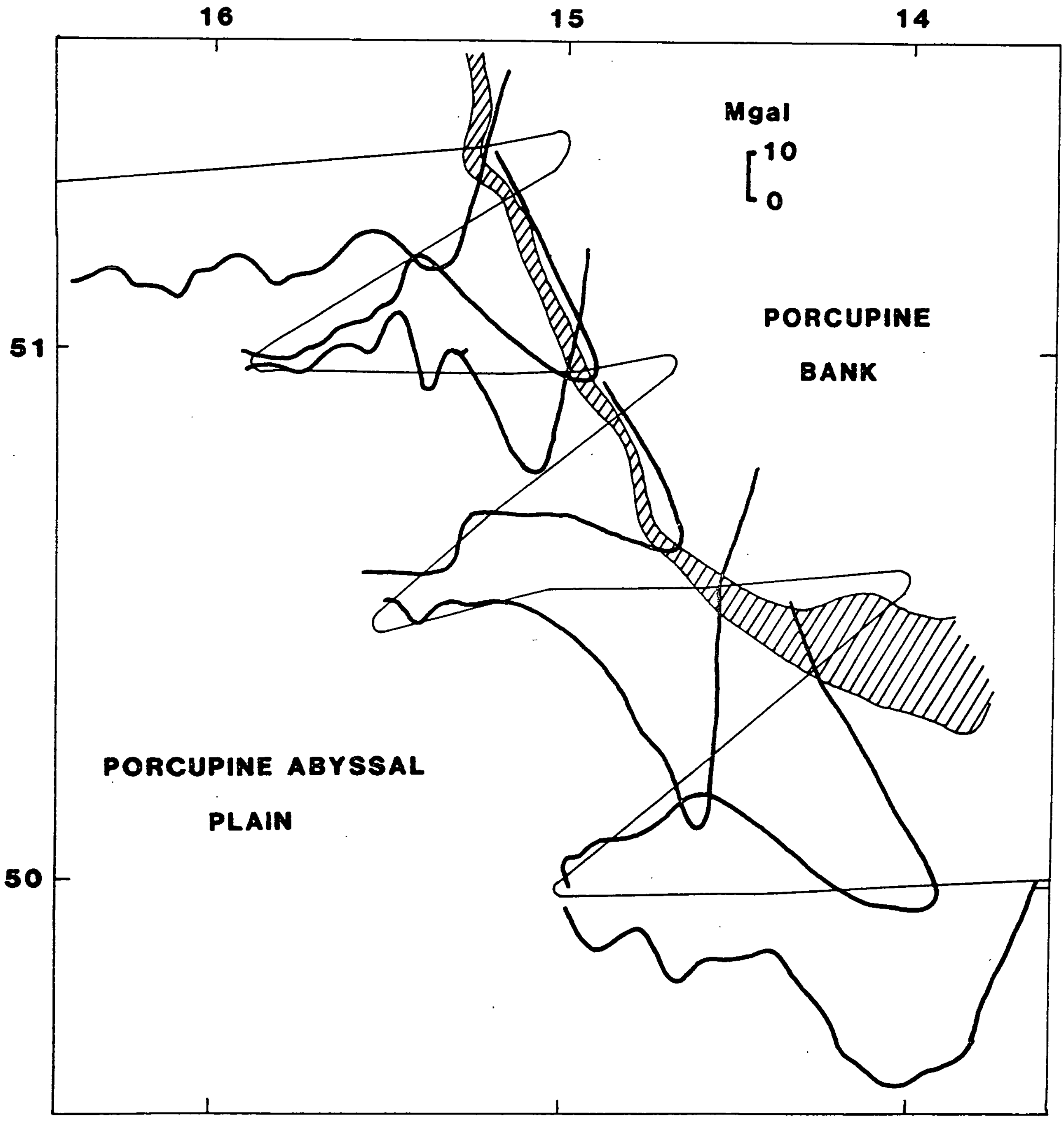


Fig. 3.21 Free air gravity profiles: Edinburgh data

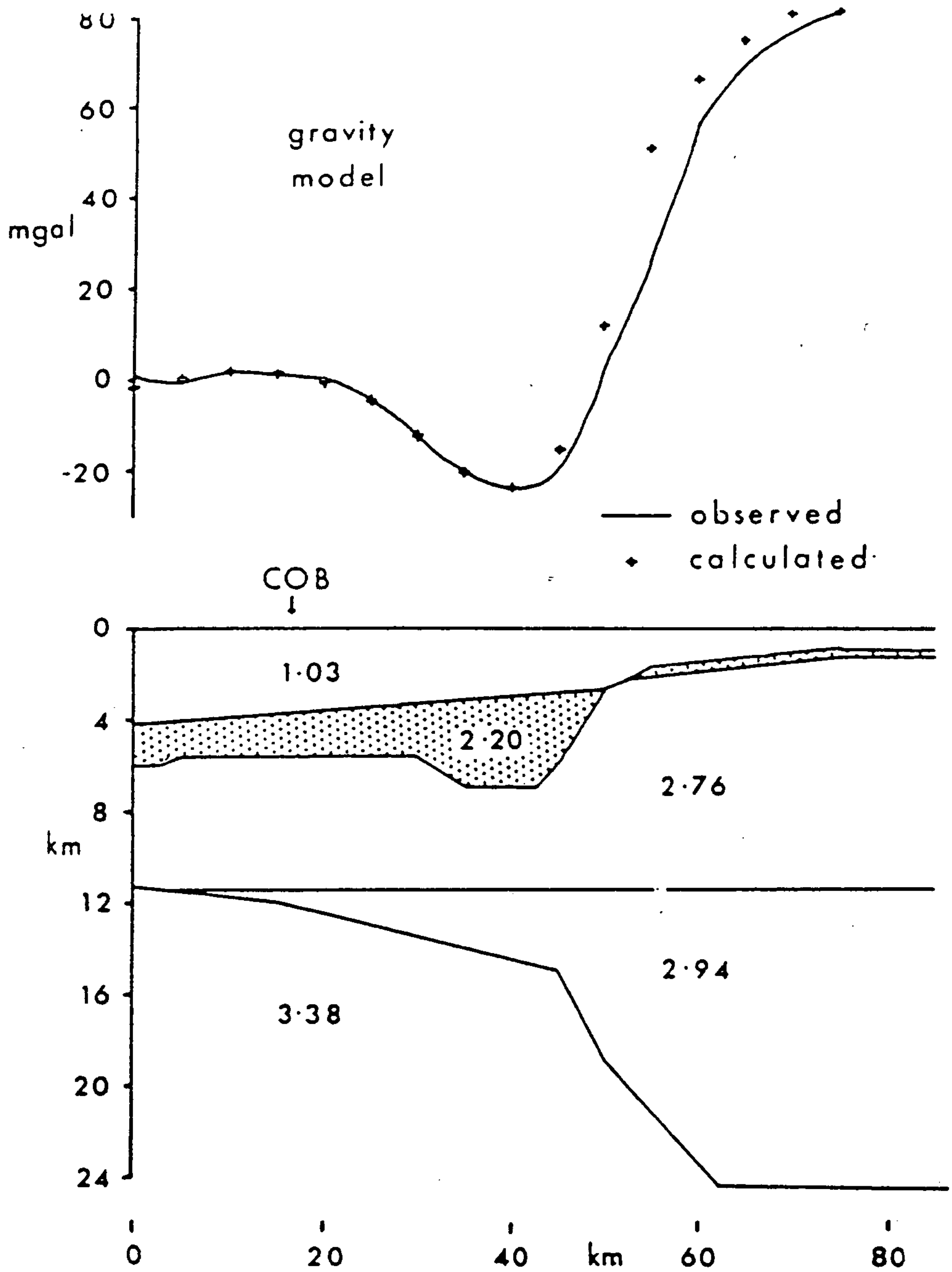


Fig. 3.22 Gravity model of Shackleton line 4; southern Porcupine Bank to Porcupine Abyssal Plain. Densities in g/cm³

western margin of Porcupine Bank, and 6 km under the Abyssal Plain.

3.3.2 The Continent-Ocean Boundary

a) Discussion

The problem of mapping the continent-ocean boundary off Porcupine Bank is approached in two different ways; by magnetic modelling of sea-floor spreading anomalies, and by investigation of the Edinburgh seismic lines. The two methods give the same continent-ocean boundary position. The continent-ocean boundary is shown overlain on a magnetic anomaly chart of part of the margin (Fig. 3.23).

Appendix A discusses the problems of mapping continent-ocean boundaries. It is to be expected that the magnetic method would be difficult to apply in this area, as the continental crust of southern Porcupine Bank is magnetic. This is no quiet zone adjacent to southern Porcupine Bank, although there is an area of lower magnitude, confused magnetic anomalies (anomalies W, X, Y and Z in Fig. 3.23).

b) Magnetic Modelling

Anomaly 32 is the oldest generally recognised sea-floor spreading anomaly off Porcupine Bank. Anomaly sequence 26-32 north of the Azores-Gibraltar Ridge was reidentified as anomaly sequence 28-34 by Cande and Kristoffersen (1977). Anomaly 32 is now usually designated anomaly 32(34) (e.g. Roberts et al., 1981(b)), although Hailwood et al.,(1979) believe that the oldest anomaly at DSDP 48 sites west of Rockall Plateau is anomaly 32, and not 34 as Cande and Kristoffersen believe.

Srivastava (1978) identified two older semi-lineated anomalies, 33 and 34, off Orphan Knoll and Flemish Cap. Kristoffersen (1978) on the other hand identifies the low amplitude anomalies outside anomaly 32(34) as noise within anomaly 34.

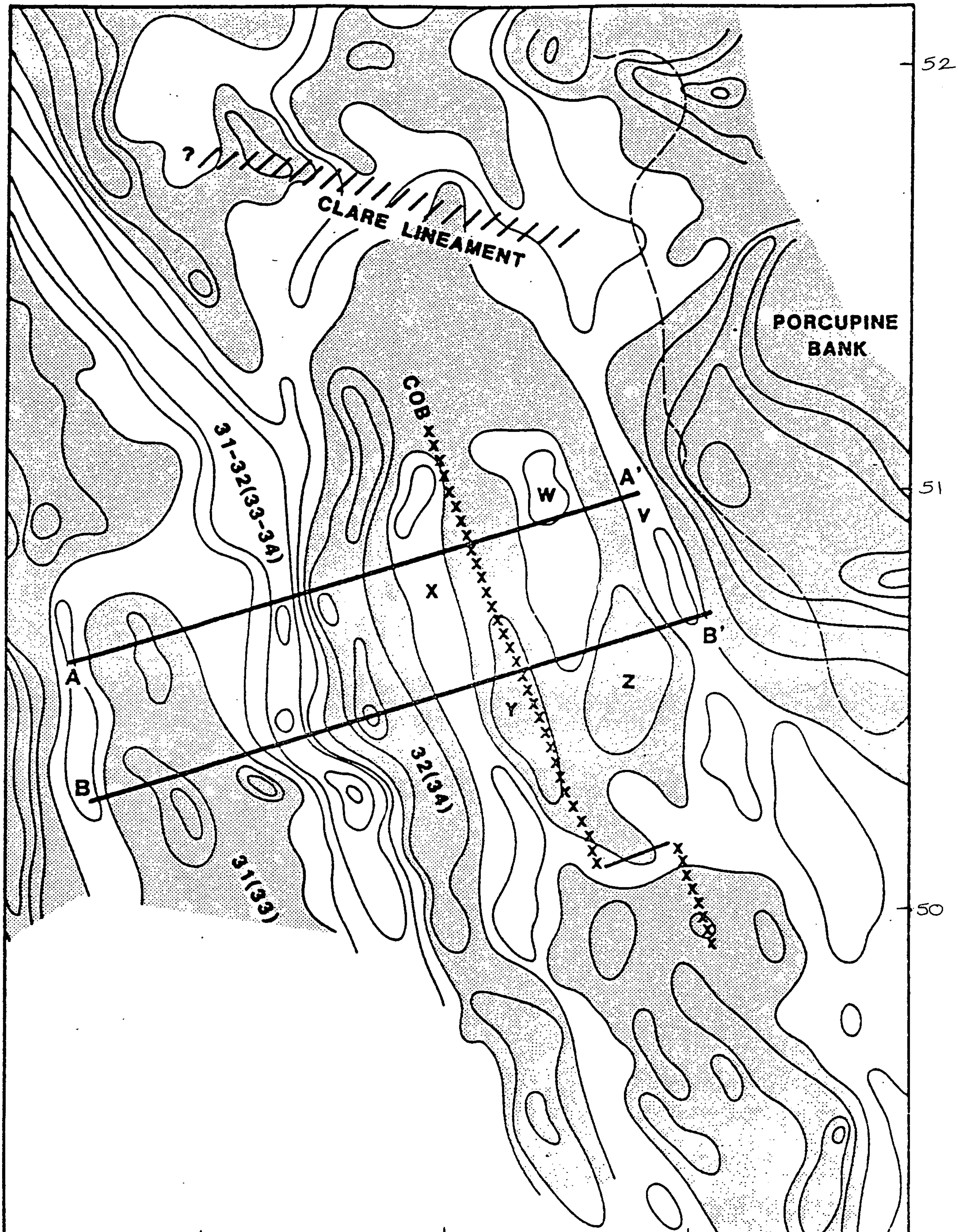


Fig. 3.23 Magnetic anomaly chart of Porcupine Abyssal Plain, Roberts and Jones 1975. Contour interval 100 nT. Shaded areas positive.. Continent ocean boundary position : this thesis.
The continent ocean boundary position from Roberts et al (1981 (b)) crosses anomalies W and Z

It was decided to model a magnetic reversal sequence to attempt to fit a reversal pattern to profiles A-A' and B-B' (Fig. 3.24). The reversal sequence used was one constructed by D. Smythe (unpub.) (Fig. 3.26). The model used a spreading rate of 11 mm/yr, after Srivastava's spreading rate for anomalies 30-32 immediately to the north of the Charlie Gibbs Fracture Zone.

Fig. 3.27 shows the results of this modelling. With the above spreading rate, the fit between the observed and calculated profiles is good as far east as negative anomaly X, and possibly the poorly defined magnetic high immediately to the east. This identifies X as anomaly 32-33. Interestingly, anomaly X is the first continuous anomaly to the west of southern Porcupine Bank.

Normally magnetised continental crust of 500 gamma magnetisation was modelled immediately to the west of anomly 32-33. This model is compared with profile A-A' in Fig. 3.28. Anomalies to the east of the continent-ocean boundary would be caused by arbitrary - in the modelling sense - variations of magnetisation within continental crust.

This modelling suggests that immediately to the east of anomaly 32-33 (X) is a break in the sea-floor spreading history of this margin. The crust to the east could be continental crust. Alternatively some of it could be oceanic crust which spread at a much slower rate, giving rise to chaotic magnetisation and topography.

An inspection of the seismic profiles indicates that this swathe of crust is more likely to be of continental origin.

c) Seismic Evidence

Shackleton line 6 (Fig. 3.29)

Fifteen km eastwards from the western end of Shackleton line 6, there is change in acoustic basement appearance. To the west, basement is at 7 seconds and is hummocky. Eastwards, basement is very variable in depth, and less reflective. It is suggested

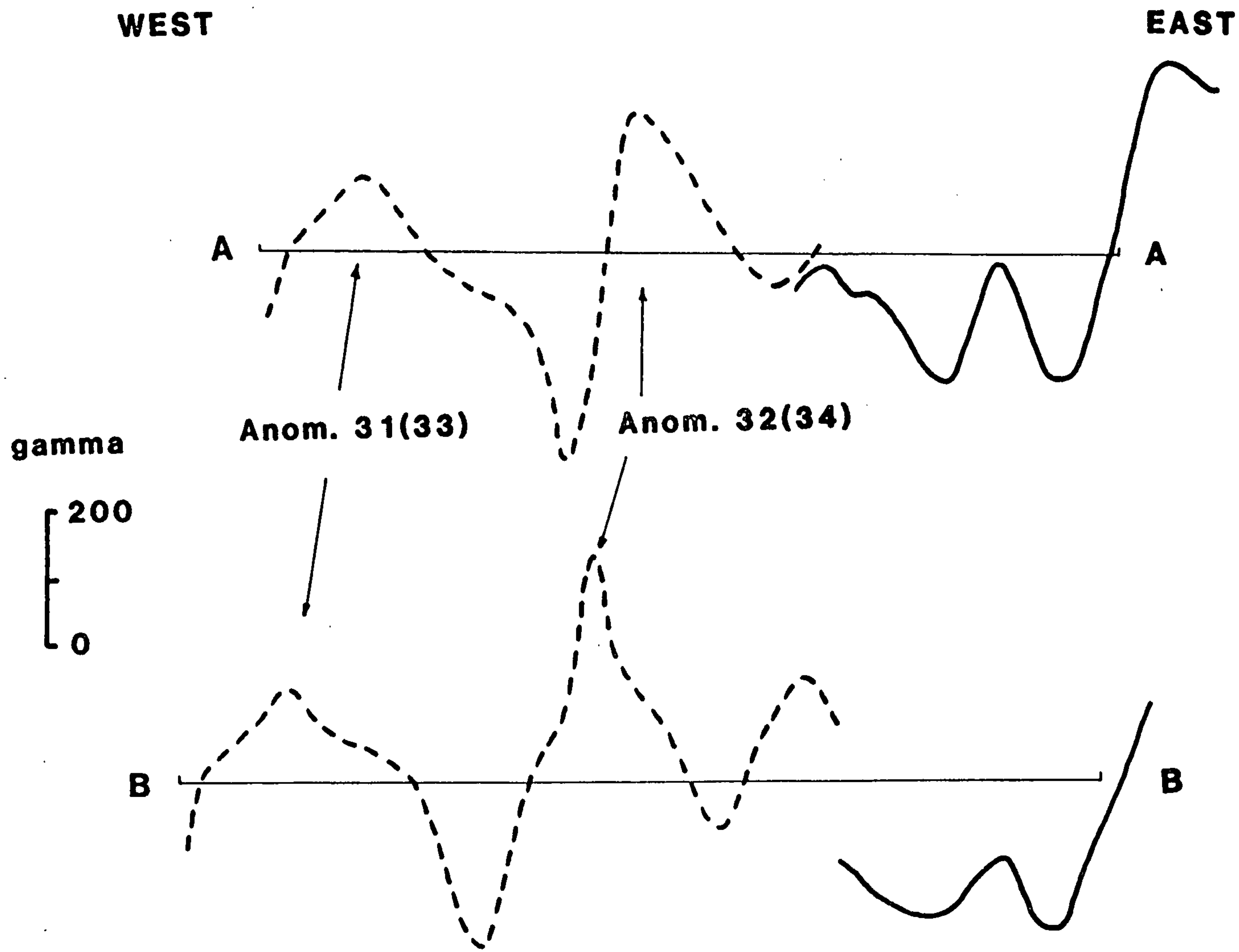


Fig. 3.24 Magnetic profiles A-A' and B-B' (Fig. 3.23).
Solid curve : Edinburgh data
Dashed curve: from Roberts and Jones 1975

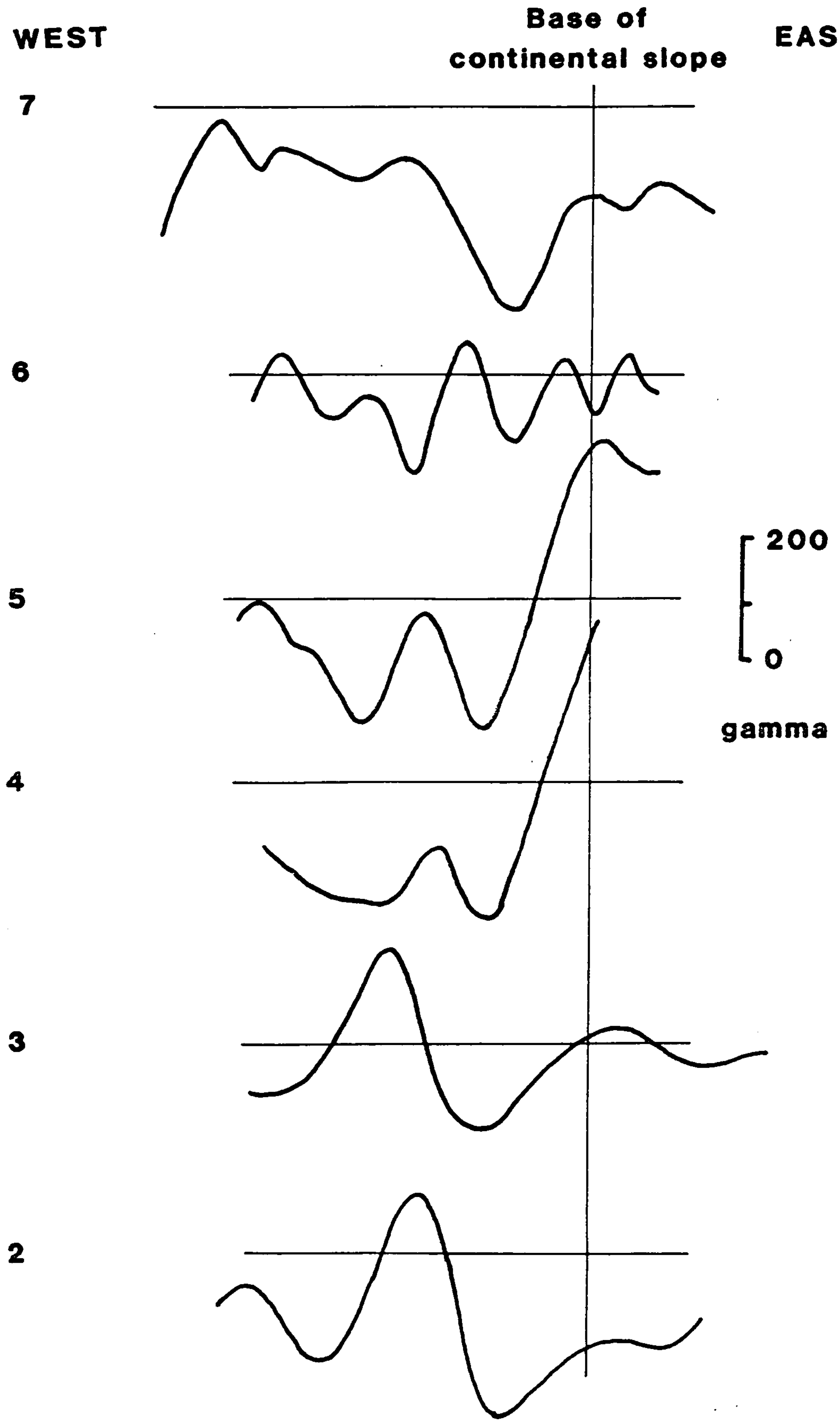


Fig. 3.25 Magnetic profiles over Shackleton lines aligned with respect to the base of the continental slope of southern Porcupine Bank

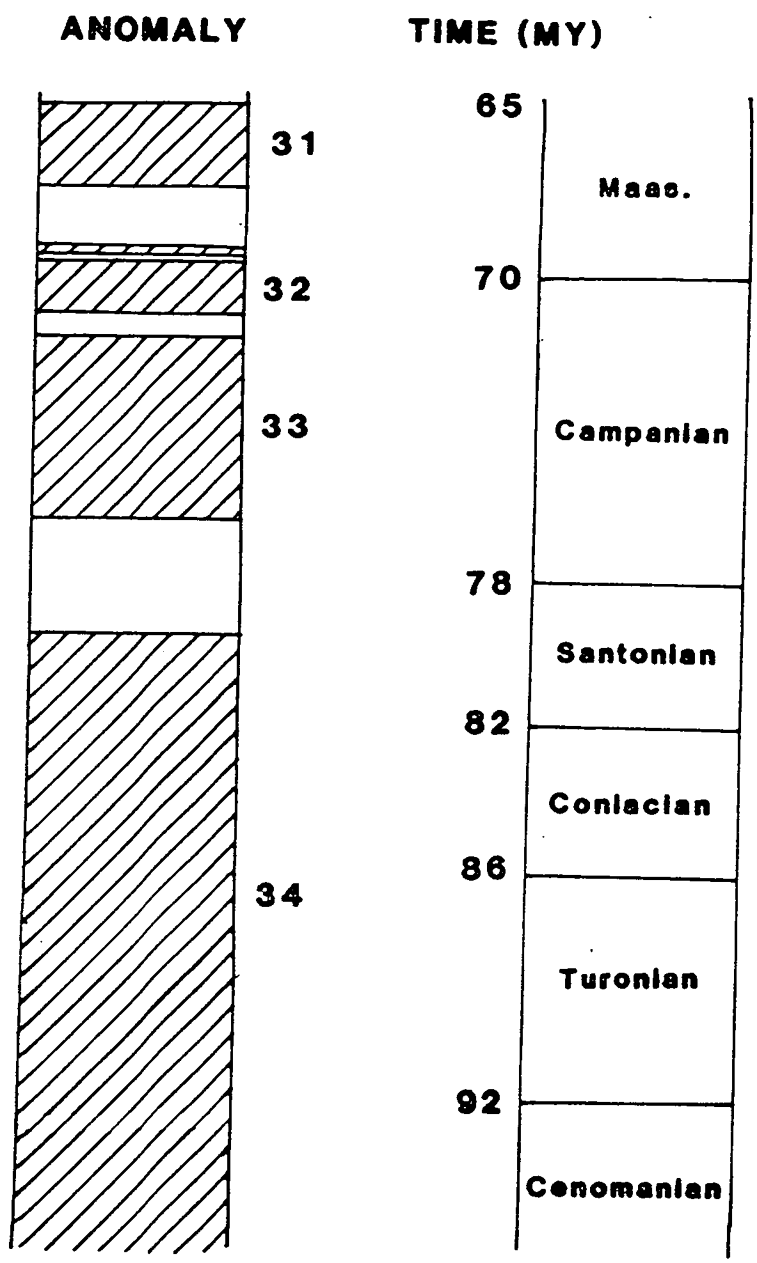


Fig. 3.26 Reversal sequence for anomalies 31-34 from Smythe (unpub.)

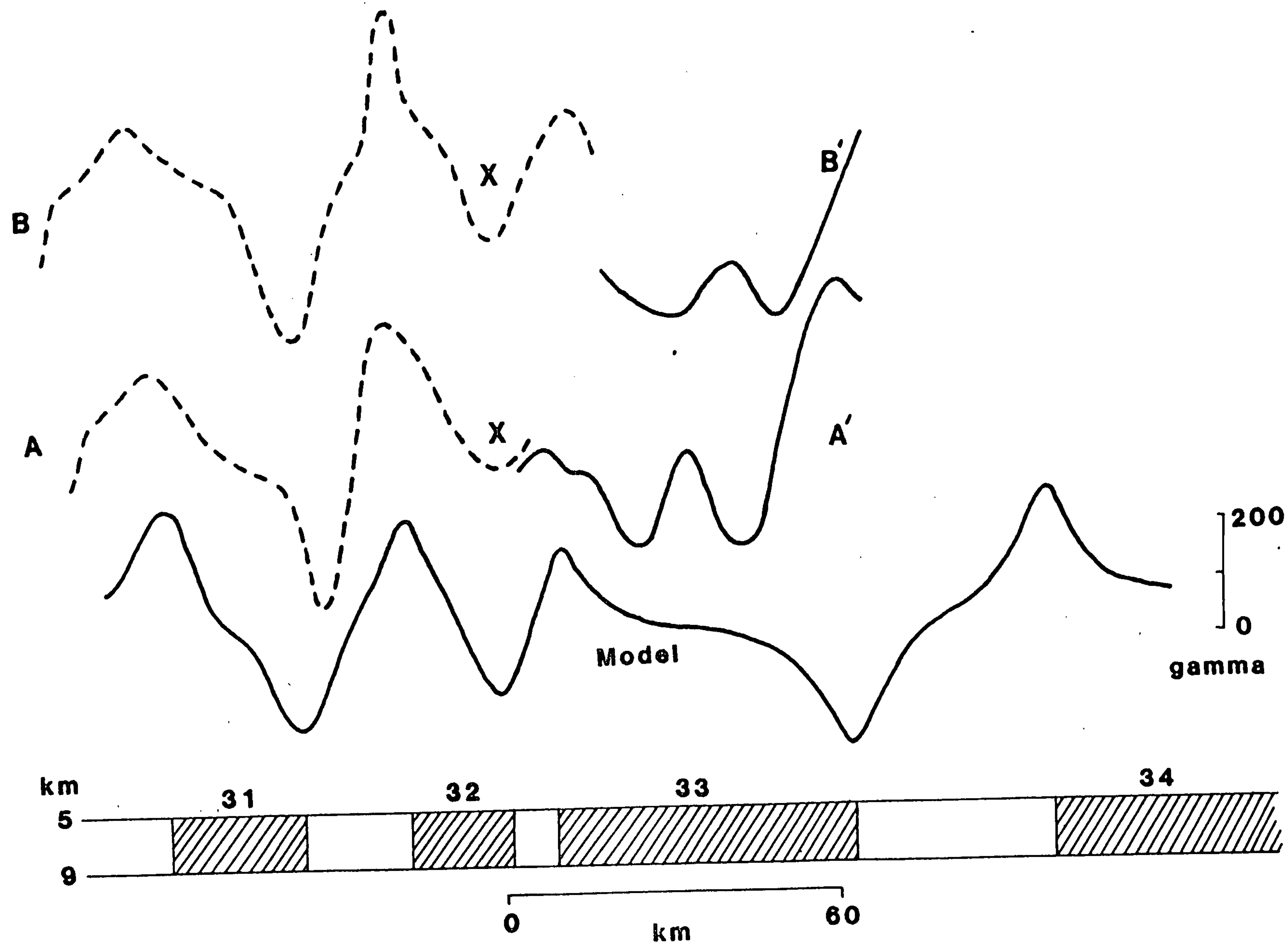


Fig. 3.27 Reversal sequence model compared with profiles A-A' and B-B' (Fig. 3.24)

Spreading date = 11 mm/yr

Magnetisation = 800 gamma

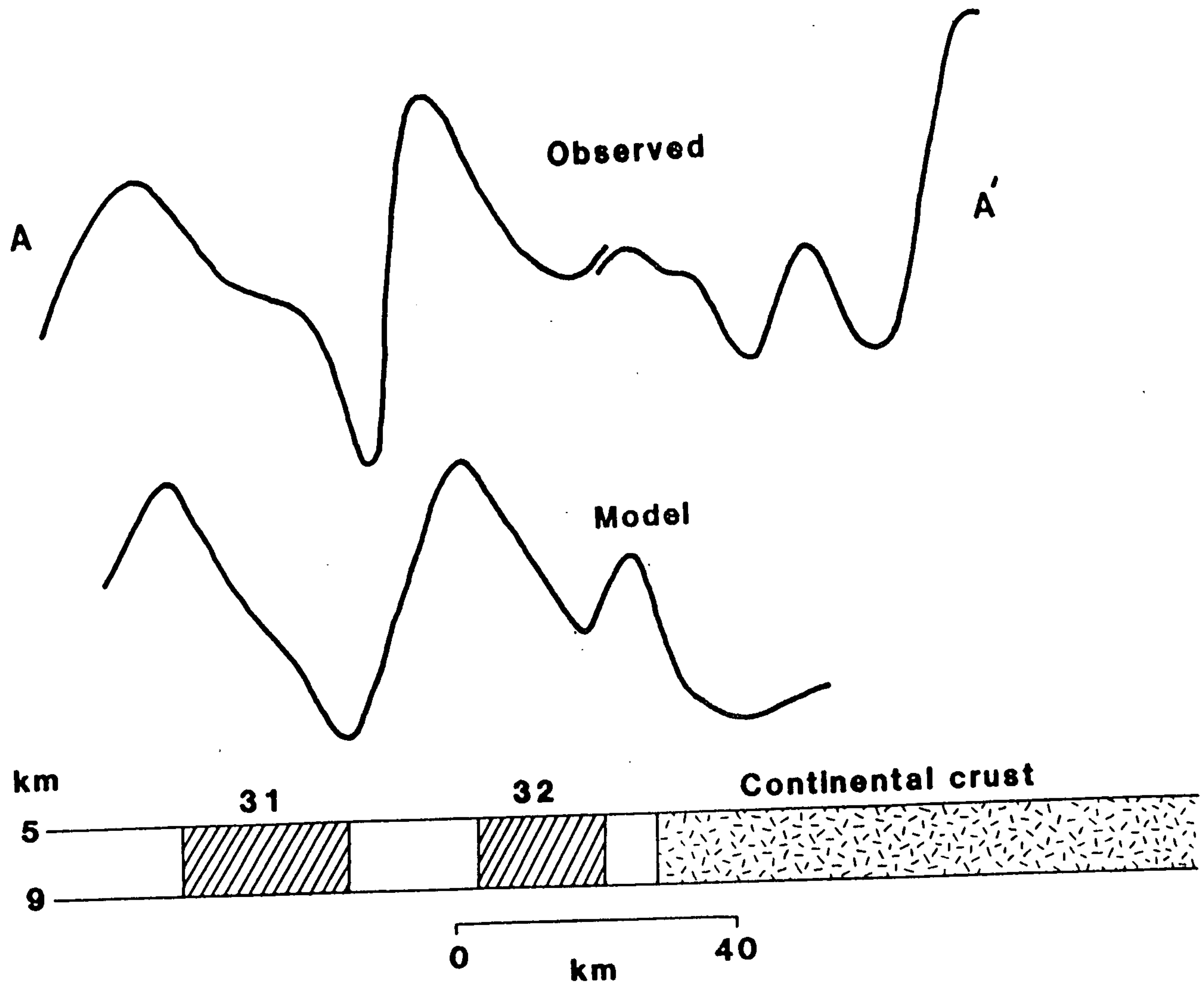


Fig. 3.28 Modelled magnetic anomalies over reversals (parameters as in Fig. 3.27) with a continent ocean boundary to the east of anomaly 32-33. Continental crust is normally magnetised with a magnetisation of 500 gamma

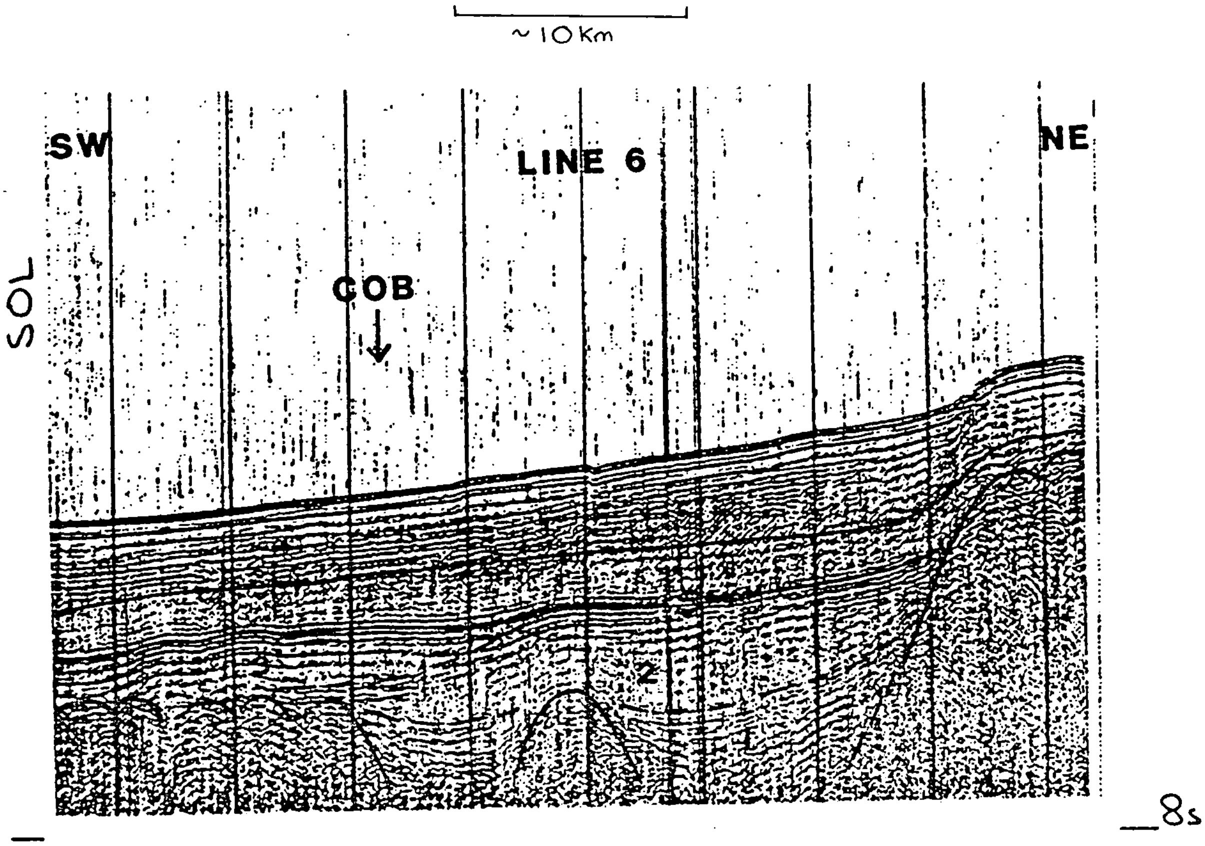


Fig. 3.29 Interpretation of the western end of Shackleton line 6, showing the continent-ocean boundary and relationship between sedimentary units 1, 2 and 3 and the Shackleton and Challenger horizons

that the COB is at 2045/091. Pre-Shackleton horizon sedimentary units are provisionally labelled units 1, 2 and 3. It appears that units 1 and 2 are pre-oceanic crustal age (i.e. pre 72 my, from Smythe), and that unit 3 is of the same age as and younger than the earliest formed oceanic crust.

A similar relationship exists between units 1, 2 and 3 on the other seismic lines, and this is used as an aid to picking the COB. Lines 4 and 5 are shown in Fig. 3.30.

The continent ocean boundary as defined seismically in this manner coincides with the boundary defined by magnetic modelling.

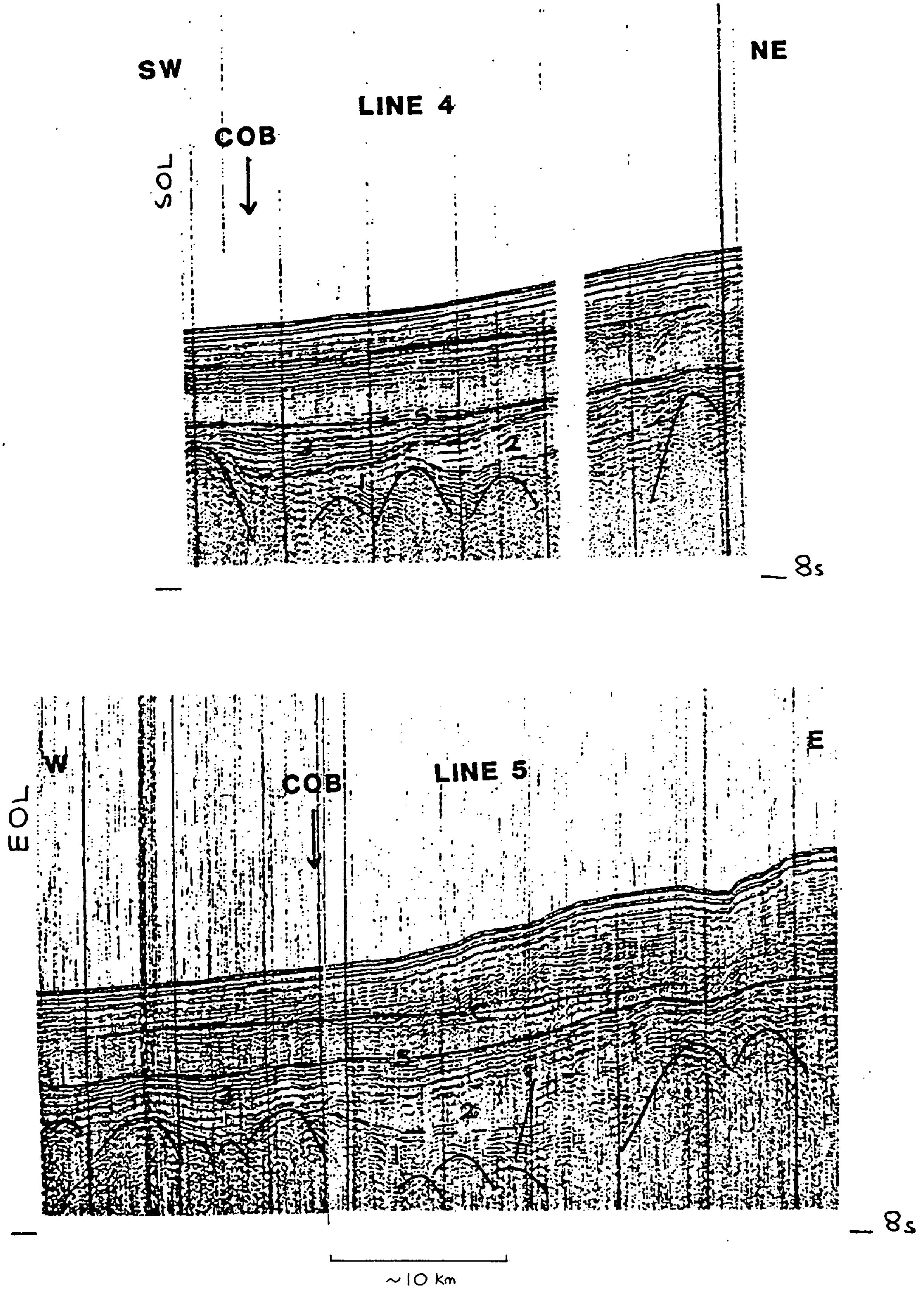


Fig.3.30 Interpretation of western ends of Shackleton lines 4 and 5

3.3.3 Conclusions

Using seismic and magnetic modelling evidence, it is concluded that the continent-ocean boundary in the Porcupine Abyssal Plain lies approximately 55 km from the base of the slope of southern Porcupine Bank.

The 55 km of a thinned continental crust probably forms a complex en echelon pattern of tilted fault blocks and horsts. The trends of these features presumably mirror the changing trends of the basement scarps which form the slope of southern Porcupine Bank, i.e. NW-SE to the south and NNE-SSW to the north.

If anomaly 32(34) is in fact anomaly 32, the first formed oceanic crust is approximately 72 my old (early Maastrichtian). If in fact this first anomaly is shown to be anomaly 34, first formed oceanic crust is older than 80 my, i.e. of probable Santonian age.

In the anomaly 34 case, the seismic evidence only can be used.

CHAPTER 4

NORTHERN PORCUPINE BANK AND SOUTHERN ROCKALL TROUGH

4.1 NORTHERN PORCUPINE BANK

In Chapter 3, Northern Porcupine Bank was defined as that part of Porcupine Bank which is to the north of the Clare Trend (Fig. 3.1). Structurally and magnetically the two provinces are distinct.

Figure 4.1 is a schematic structural map of Porcupine Bank based on data described in this chapter.

4.1.1 Database

a) Magnetics

i) Tertiary igneous centres

The magnetic anomaly field of northern Porcupine Bank is a fairly flat field superimposed on which are six short wavelength high amplitude anomalies. These short wavelength anomalies trend northeast-southwest. All are positive anomalies except one at 53° 40'N (Fig. 4.1). It is likely that these represent the magnetic field due to ?Tertiary igneous centres. They may be intruded along faults of Caledonoid trend. (Refer to Riddihough and Max, 1976, for discussion of Caledonian trends over Porcupine Bank). Alternatively they may be associated with the stress system of Cretaceous age which formed the Clare Lineament and the NE-SW trending grabens in northern Porcupine Bank.

ii) The 53°N Flexure Zone

Bailey (1975 a and b) named an east-west trending '53°N Flexure Zone' across northern Porcupine Bank. This zone was considered to be a continuation of the structural trend at the same latitude across northern Porcupine Seabight (refer to Fig. 1.14).

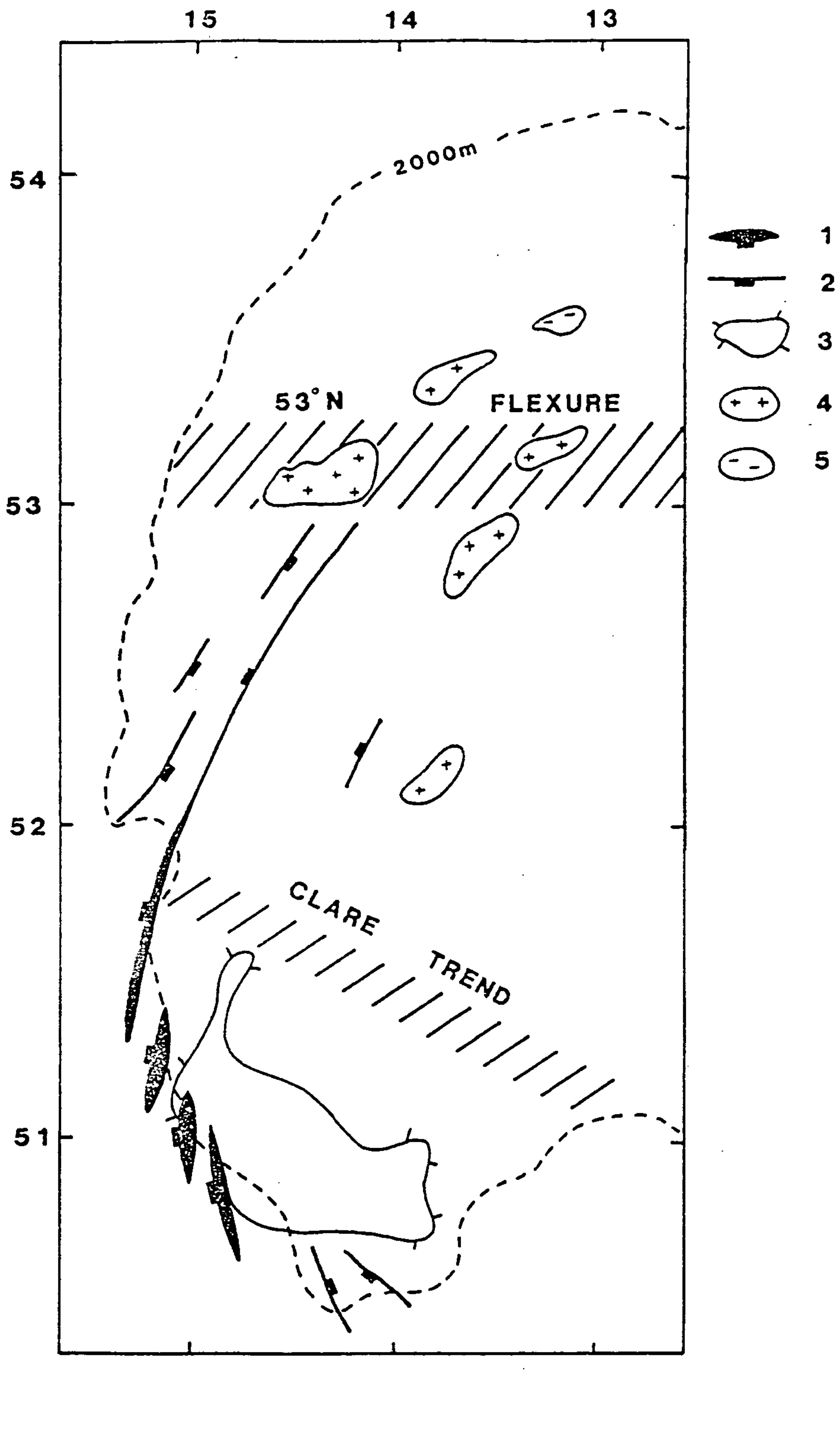


Fig. 4.1 Schematic structural map of Porcupine Bank

1: exposed basement scarps

2: major faults within Porcupine Bank

3: 200 gamma contour of southern Porcupine Bank magnetic high

4: ?Igneous centres - normal magnetisation

5: ?Igneous centres - reversed magnetisation

Riddihough and Max (op. cit.) recognised an east-west trend at 53° 15' from magnetic data. They consider that this lineation is the continuation of the Fair Head - Clew Bay line in Ireland, which is the continuation of the Highland Boundary Fault.

On the magnetic anomaly map (Fig. 3.1) this east-west trend is more striking than on the total field magnetic map. In addition it can be seen that the short-wavelength anomalies, or possible igneous centres, are concentrated spatially around the latitude of this trend.

This zone, which is named the '53°N Flexure Zone' on Fig. 4.1, after Bailey, may be a Caledonian discontinuity and a zone of weakness which subsequently allowed the formation of Porcupine Seabight Trough to the south.

b) Gravity

Chart 2 shows free air anomalies over Porcupine Bank. In this area, data from University College of North Wales and from Edinburgh University cruises are not strictly compatible. The contouring shown is approximate. It appears that the gravity field over northern Porcupine Bank is considerably flatter than that over southern Porcupine Bank.

A gravity model over an Edinburgh line is presented in section 4.1.2. (b).

c) Seismic

Tracks of all available seismic lines over northern Porcupine Bank are shown in Fig. 4.2.

The DSDP leg 12 track over Porcupine Bank is virtually coincident with the southern Charcot track, and is not shown.

Times to acoustic basement were mapped from Edinburgh and Charcot data, with limited use of the Clarke et al.,(1970) profiles.

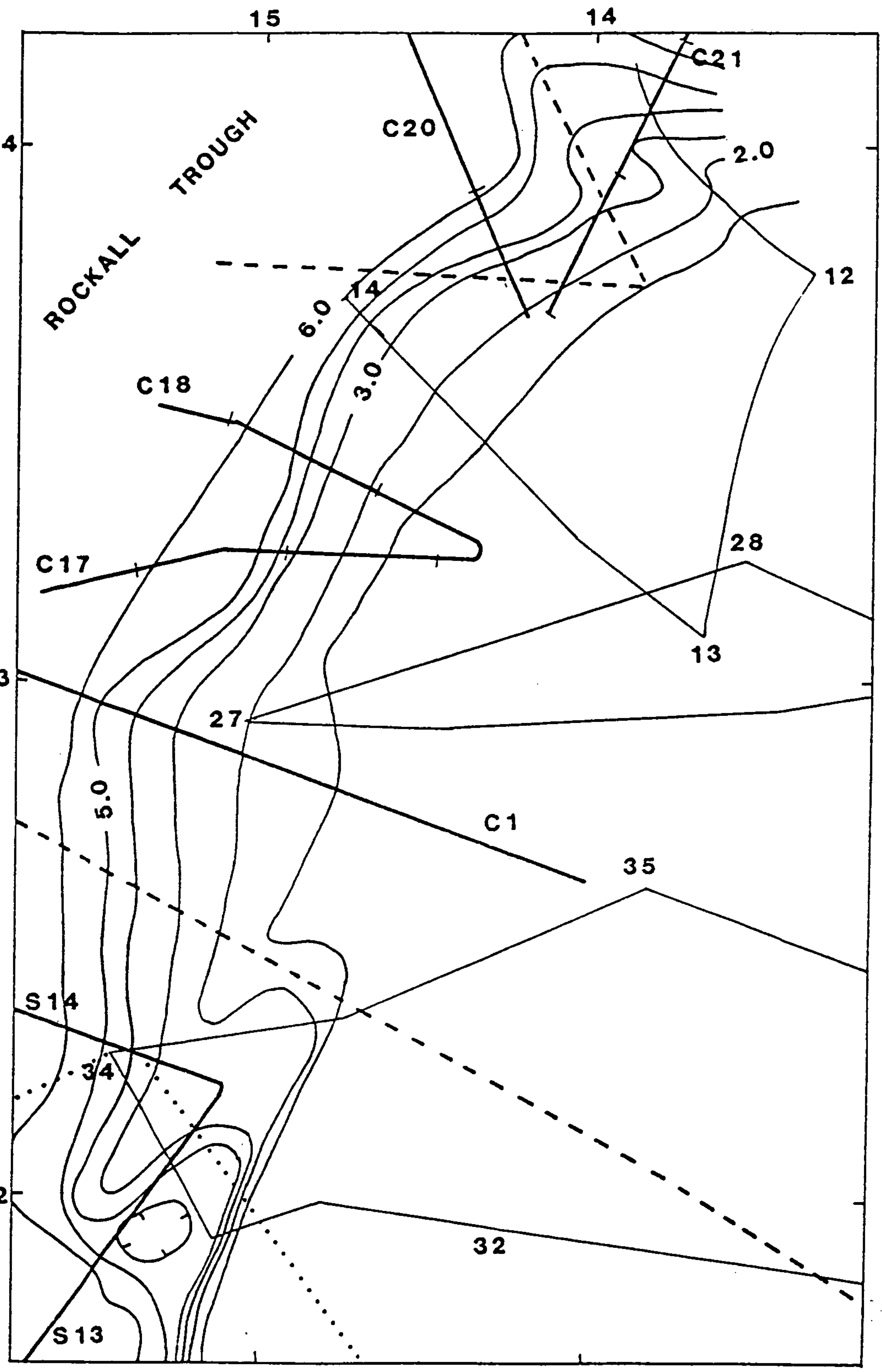


Fig. 4.2 Available seismic lines over northern Porcupine Bank, approximate time to basement contours from 1.0 to 6.0 seconds TWT

Heavy lines: Edinburgh profiles

Light lines: profiles from Clarke, Bailey, and Taylor Smith (1970)

Dashed lines: Charcot profiles

Dotted line: profile from Roberts (1975)

No attempt was made to map basement shallower than 1.0 second, as the data are completely obscured by water-bottom and bubble pulse multiples. Bailey, Buckley and Clarke (1971) contoured depth to basement over the shallower areas of Porcupine Bank using sparker data. Bailey (1978) discusses the thin, locally absent, sedimentary sequence over northern Porcupine Seabight.

The time to basement chart illustrates the strong contrast between the basement gradients at the western margins of northern Porcupine Bank and southern Porcupine Bank. (Compare with Fig. 3.19).

4.1.2 The NE-SW Trending Graben System

There is a prominent reentrant in the continental slope of western Porcupine Bank at the position at which it is intersected by the Clare Lineament and Clare Trend. This reentrant is the expression of a NE-SW trending graben system which can be traced as far north as 53°M (Fig. 4.3). Gray and Stacey commented upon the reentrant in 1970.

A gravity low overlies this graben, which locally contains over 2 seconds (TWT) of sediments. The graben shallows northeastwards.

Profile M-L-K (Fig. 4.4) from Roberts (1975) shows that in the southern part of the graben, the eastern bounding fault is very steep. This eastern scarp appears to be a continuation of the northernmost steep exposed basement scarp of southern Porcupine Bank.

a) Gravity Modelling

Edinburgh survey line Challenger 1 (Fig. 4.5, and line C1 in Figs. 4.2 and 4.3) crosses the graben system at 52° 45'N. Fig. 4.6 shows the crustal model that was constructed for the whole of Challenger line 1. Crustal thickness under northern Porcupine Bank was modelled as 27 km. The fit over northern Porcupine Bank is poor, and a more detailed gravity model over this area is shown in Fig. 4.7.

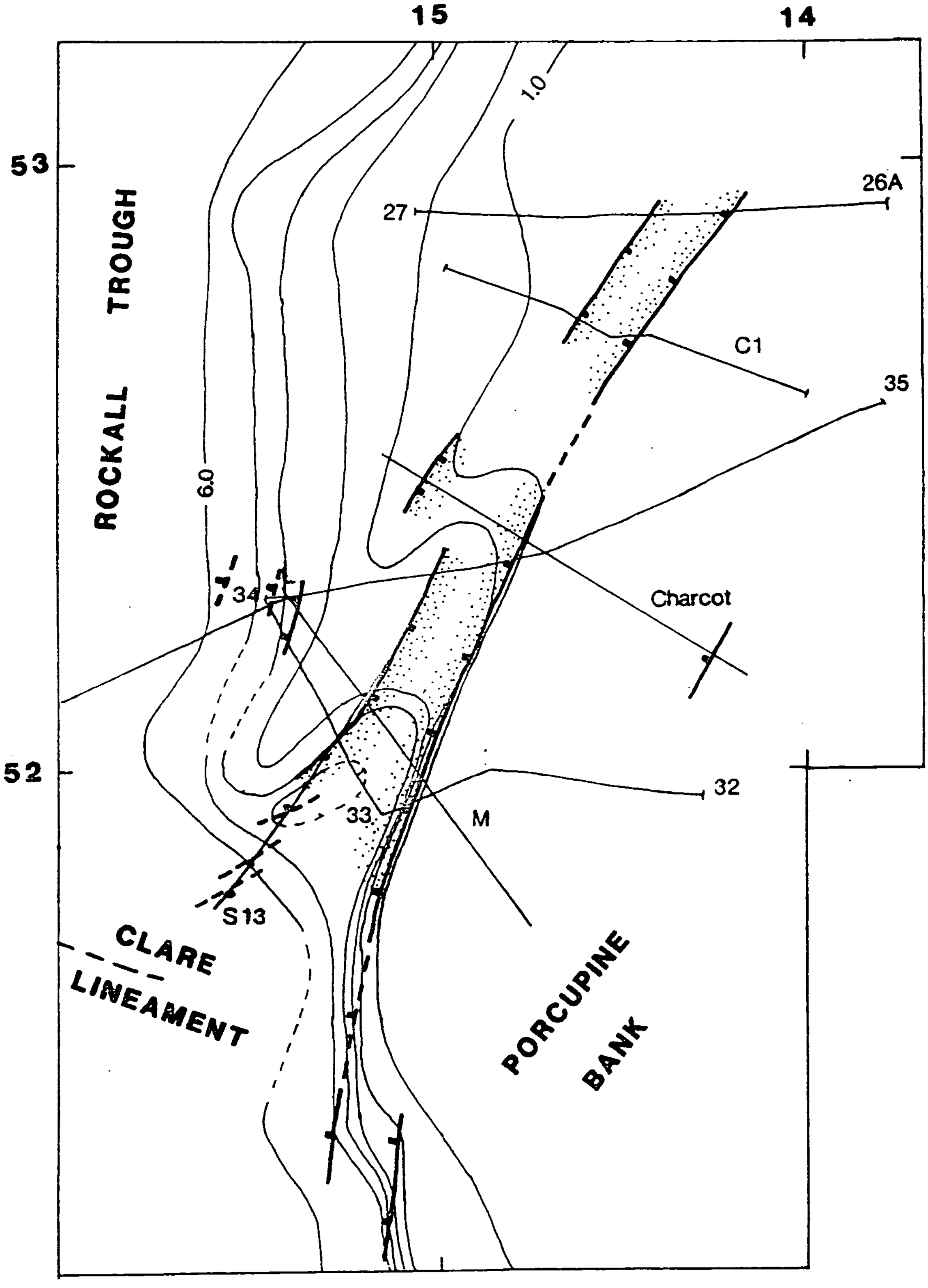


Fig. 4.3 The NE-SW trending graben system in northern Porcupine Bank, and seismic lines across the graben. For surveys, refer to Fig. 4.2

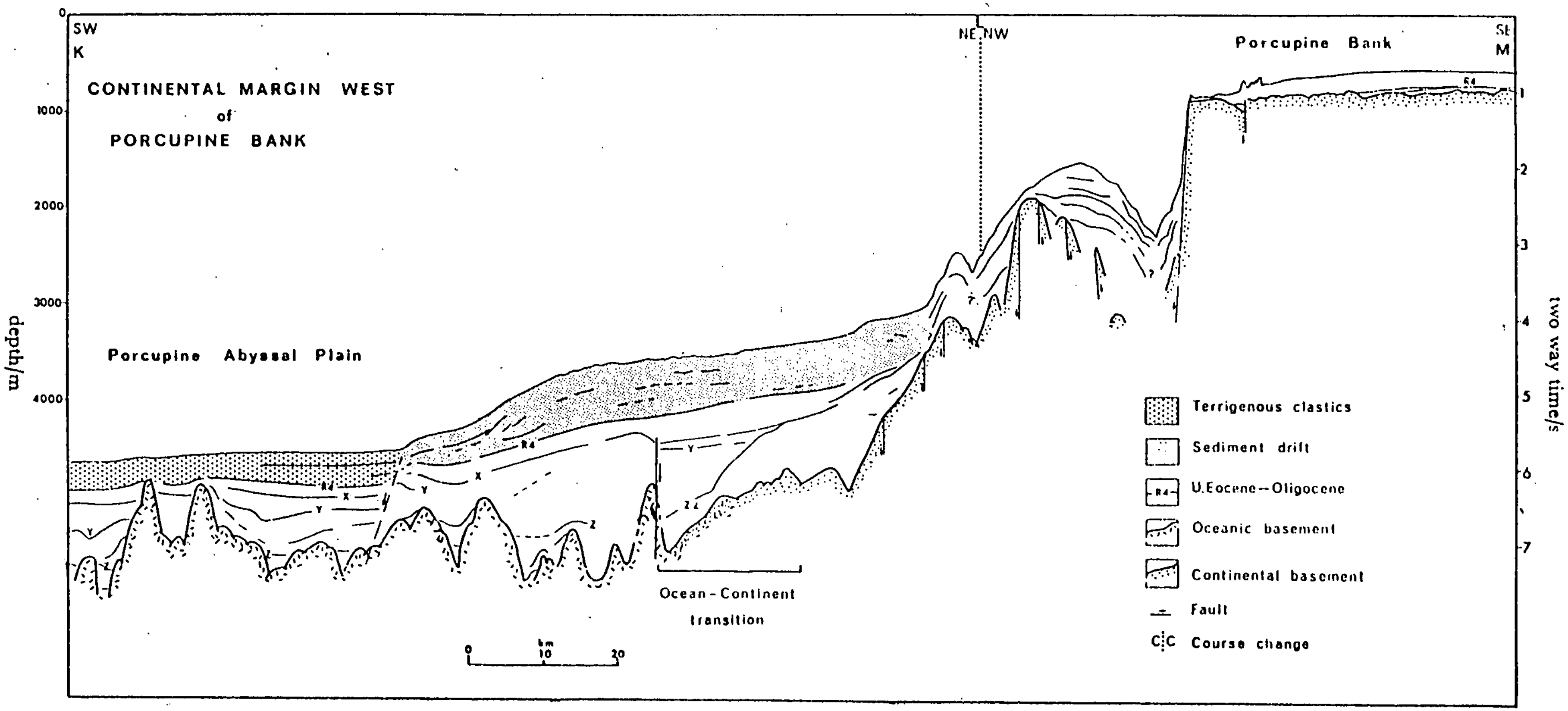


Fig. 4.4 Profile M-L-K, from Roberts (1975). Refer to Fig. 4.3 for position

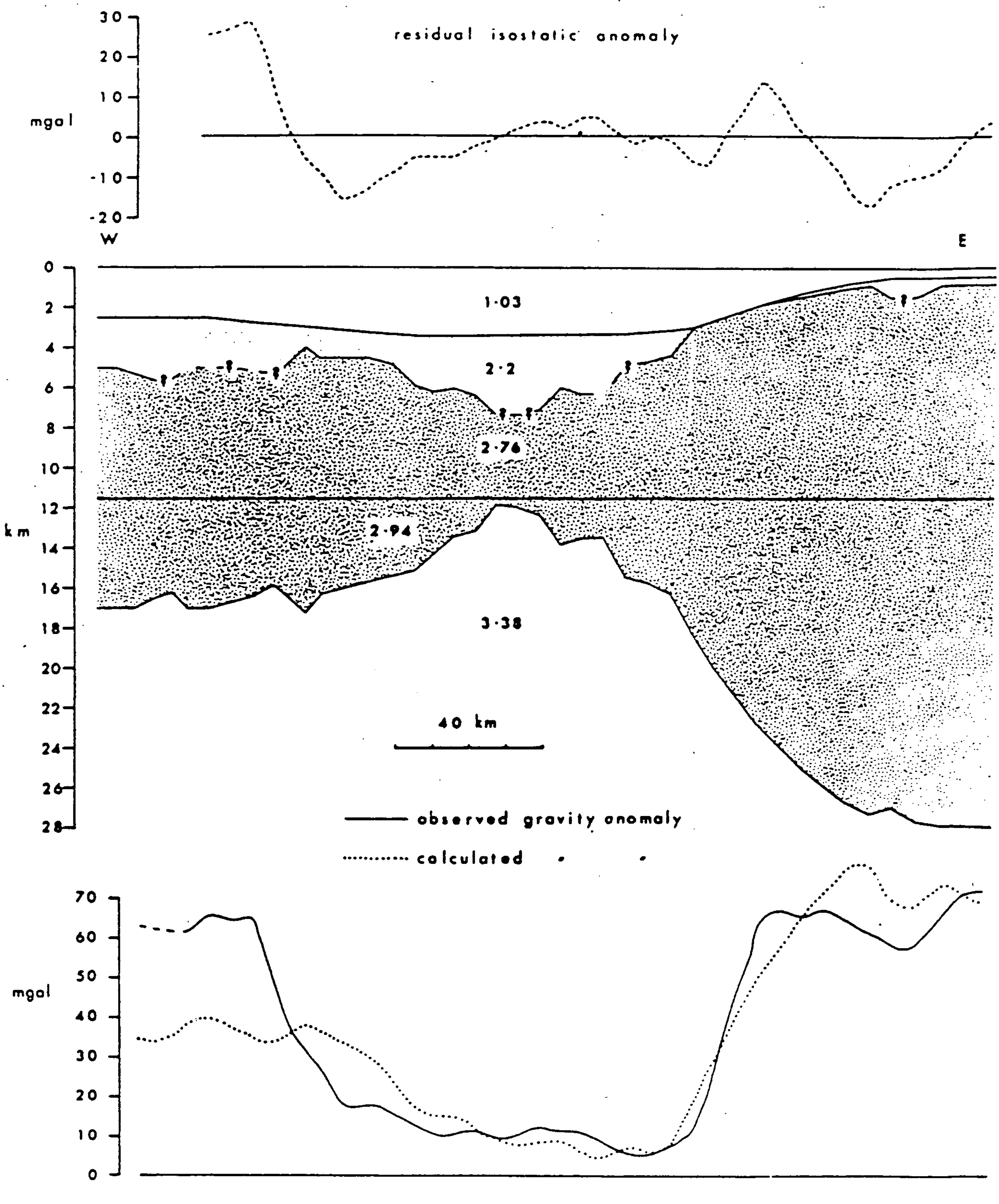


Fig. 4.6 Isostatic gravity model of Challenger line 1. Densities in g/cm³

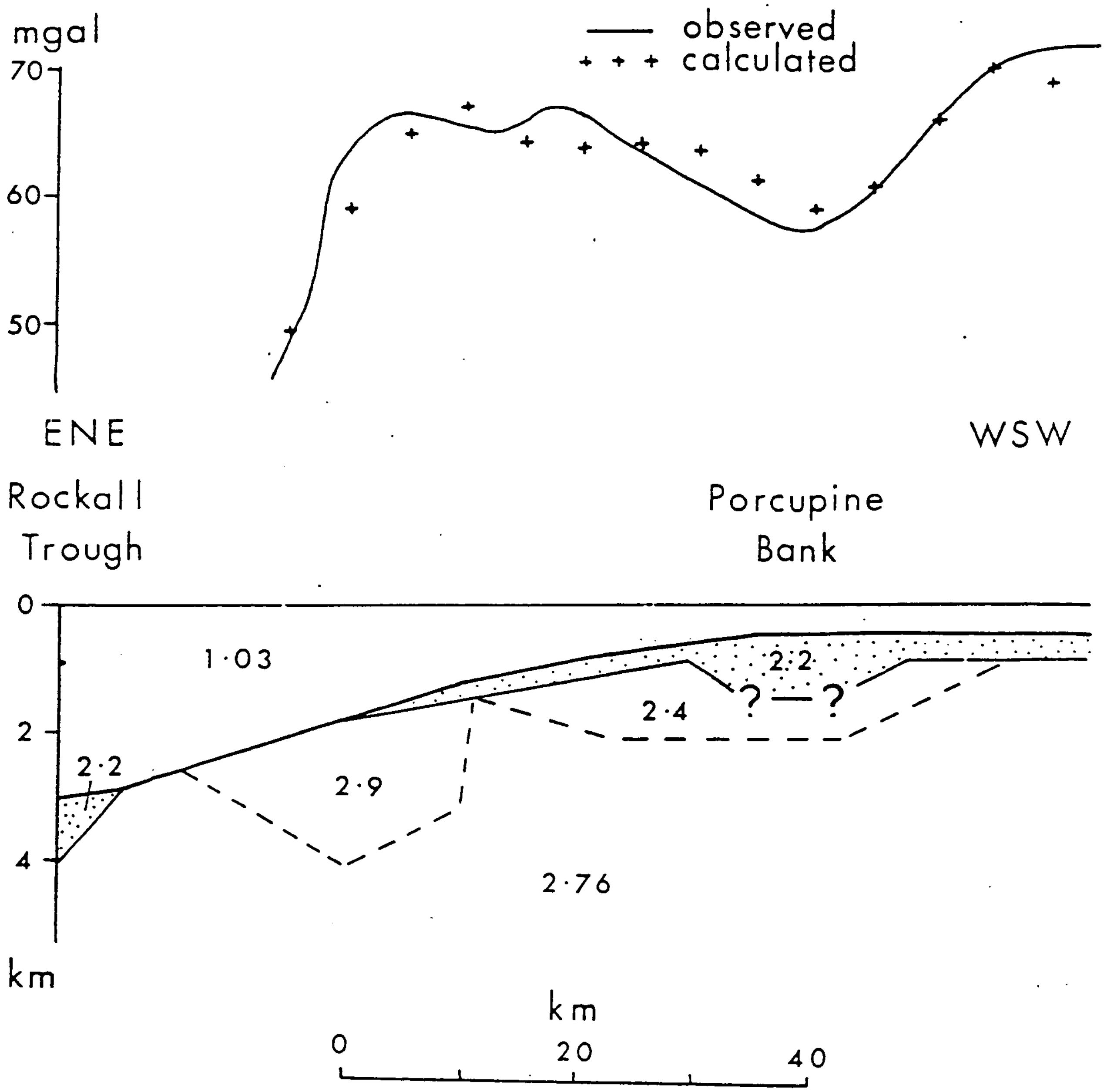


Fig. 4.7 A more detailed gravity model of Challenger line 1 over Porcupine Bank. Densities in g/cm^3. Line C1, Fig. 4.2

A graben approximately 2 km depth roughly satifies the observed gravity anomaly. Acoustic basement is difficult to map from the seismic data, and the structure of the graben remains unknown.

b) Position of the Graben System with Respect to the Structural Elements of Northern Porcupine Bank.

The graben system cannot be mapped on strike to the north of the 53°N Flexure.

The NE-SW trend of the graben is the same trend as that of the Slyne-Erris Trough system to the northeast of Porcupine Seabight Trough. Both graben systems may be a result of tensional stress reactivating Caledonian faults. Alternatively, the Porcupine Bank graben could be a consequence of the WNW-ESE directed tensional stresses of Cretaceous rifting which produced the en echelon scarps of southern Porculine Bank.

4.2 SOUTHERN ROCKALL TROUGH

4.2.1 Objectives

Previous work on southern Rockall Trough is described in Hill (1952), Clarke (1973), Roberts (1975), and Roberts et al (1981(b)). Theories on the formation of Rockall Trough are:

i) The whole of the Trough is underlain by Cretaceous oceanic crust, of anomaly 34 age (Bott, 1978)

ii) The whole of Rockall Trough is underlain by thinned continental crust (Talwani and Eldholm, 1972 and 1977).

iii) The Trough is underlain by Permian oceanic crust (Russell and Smythe, 1978)

iv) The centre swathe of southern Rockall Trough is underlain by Cretaceous (pre-anomaly 32) oceanic crust, flanked by thinned continental crust (Roberts et al, 1981(b)). Fig. 4.8.

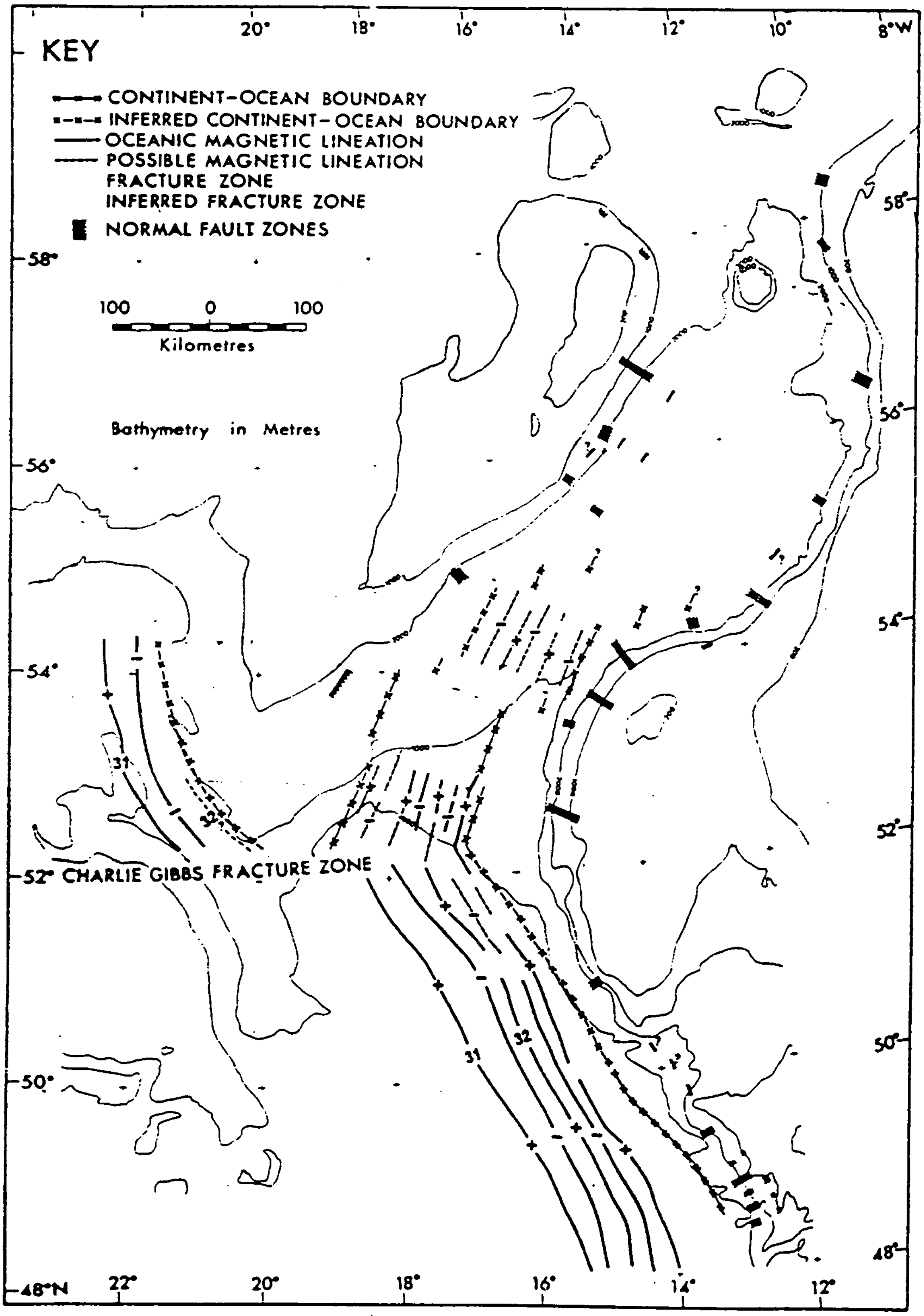


Fig. 4.8 Regional structural interpretation, southern Rockall Trough and surrounding area. Roberts et al 1981(b)

The objectives of this section are to describe the approximately 3,500 km of new data in southern Rockall Trough, and to discriminate between the above theories of formation, as far as is possible. The conclusion arrived at in this chapter is that southern Rockall Trough is floored by thinned continental crust with extensive central igneous intrusions.

The work undertaken has attempted to

a) delineate different crustal units, or the basis of magnetic, seismic, and gravity characteristics. Magnetic and gravity modelled was used for quantitative distinctions.

b) describe the age and evolution of southern Rockall Trough by drawing comparisons with other areas. A thermal maturation calculation is presented in chapter 6.

The extent to which conclusions drawn about the evolution of southern Rockall Trough can be applied to Rockall Trough north of 55°N is discussed in this chapter and in chapter 6.

4.2.2 Magnetic Data and Modelling

4.2.2.1 Noisy and Quiet Zones, and their Correlation with Seismically Distinct Basement Types.

Figure 4.9 shows the magnetic anomaly map over southern Rockall Trough. The interpretation of this area is given in Fig. 4.10. Terms used in this chapter for magnetic zones are added, and the position of the gravity high discussed in section 4.2.3.

The most prominent magnetic features are

i) The east-west trending positive anomalies of the Charlie Gibbs Fracture Zone. This is discussed in chapter 5.

ii) Positive anomalies of the central noisy zone. Up to + 600 gamma (nT) in amplitude, these circular or lens-shaped anomalies trend NE-SW immediately to the north of the Charlie Gibbs Fracture Zone, and NW-SE under Feni Ridge and Rockall Plateau. The anomalies link up with anomalies of similar amplitude and size over southern Rockall Plateau.

Negative anomalies within the central noisy zone are smaller in area and of lower amplitude than the positive anomalies.

iii) The eastern and western magnetic quiet zones are locally magnetically very flat, and have anomalies of zero to -200 gamma

iv) The northern quiet zone is magnetically varied, with a general NE-SW anomaly trend.

Correlation with basement types.

Every major positive anomaly in the central noisy zone is crossed by an Edinburgh line. It is of great significance that <u>every</u> positive anomaly is underlain by shallower basement of highly reflective appearance, and that all basement underlying negative

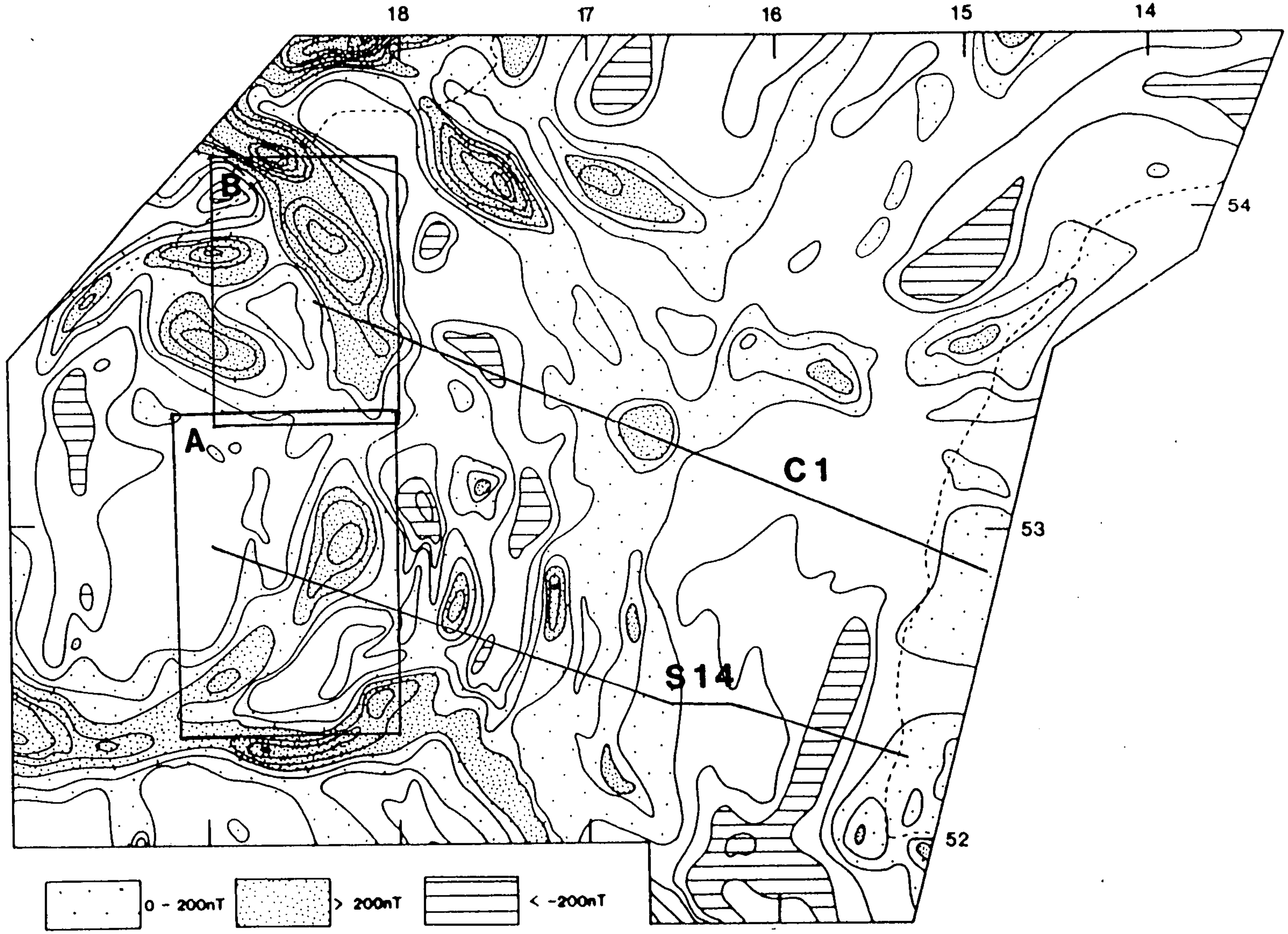


Fig. 4.9 Magnetic anomaly map of southern Rockall Trough. Challenger line 1 (C1) and Shackleton line 14 (S1) are shown. Boxed areas A and B are discussed in section 4.2.2.1. Dashed lines - 2,000 m isobath. 1nT = 1 gamma

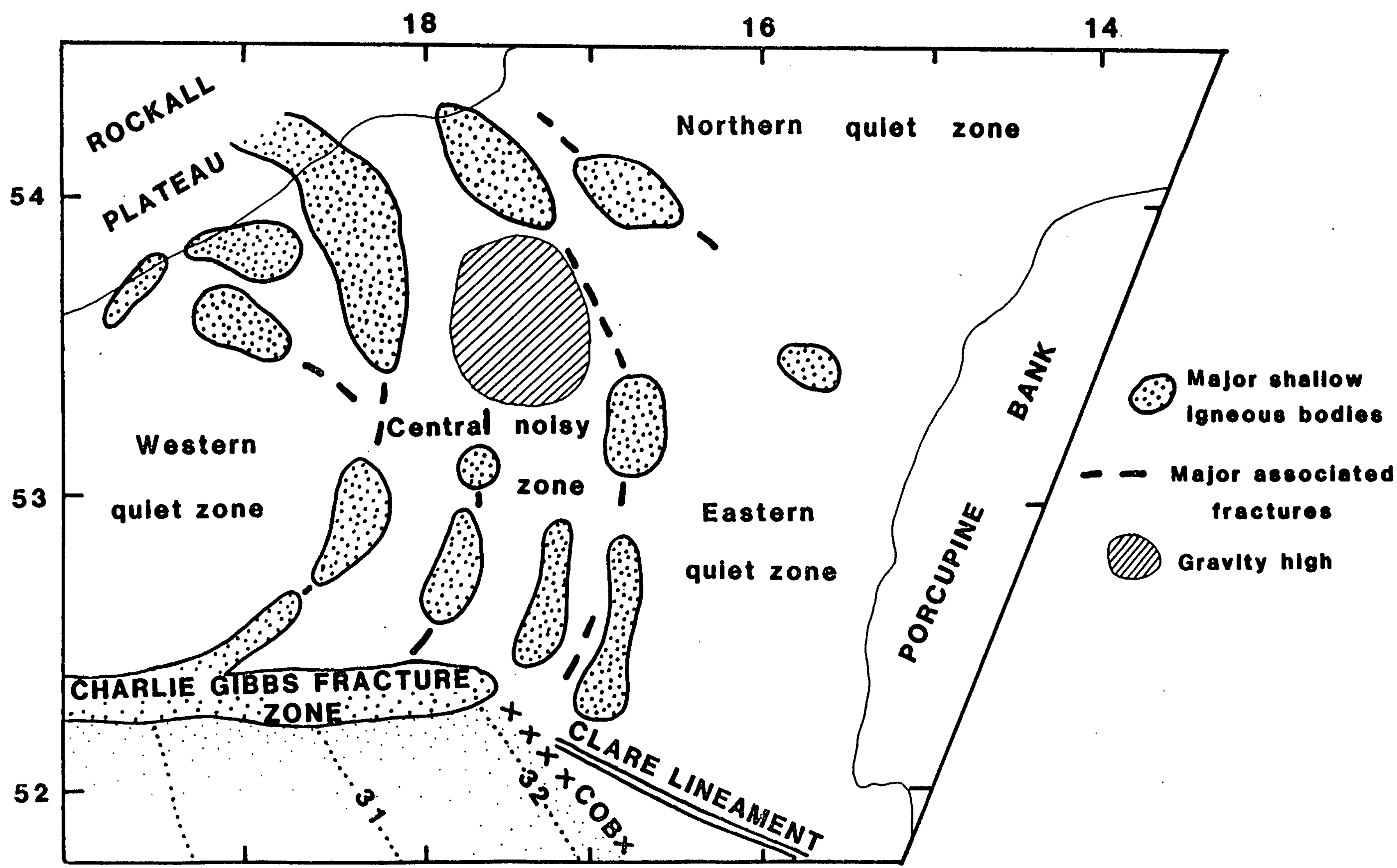


Fig. 4.10 Structural interpretation of southern Rockall Trough, to same scale as Fig. 4.9

anomalies of the central noisy zone, or underlying the quiet zones, is deeper and less reflective. There are no highly reflective basement highs which do not underlie magnetic highs, i.e. there is a one-to-one correspondence. This immediately implies a different basement type beneath magnetic highs, and this section and the following one develop the argument that all the magnetic highs are underlain by normally magnetised igneous bodies, and that the surrounding crust is thinned continental crust of low magnetisation.

Challenger line 1 (Fig. 4.5) and Shackleton line 14 (Fig. 4.11) show the change in basement reflectivity and height associated with positive magnetic anomalies. These two lines are treated in more detail in the magnetic modelling section.

a) Area A

Area A (Fig. 4.9) lies at the western boundary of the central noisy zone. It is shown in greater detail in Fig. 4.12.

Three lines - Shackleton 14 (S14), Challenger 7 (C7), and Challenger 5 (C5) cross the major NE-SW trending 400 gamma anomaly.

On Shackleton 14, this magnetic high is underlain by a prominent basement high at 1100-1130z/097. This ridge can be seen at a depth of 5 seconds (TWT) at 0430z on Challenger line 7, but is too deeply buried to be visible on Challenger line 5 (Fig. 4.13). The basement high is contoured on Fig. 4.12. Note how the basement ridge, where present, has controlled the deposition of the Feni Ridge to the west.

b) Challenger line 2

Challenger line 2 crosses the NW-SE trending magnetic anomalies which are the largest in the central noisy zone. The anomalies here are also directly underlain by basement highs, in this case with approximately 2 seconds of topographic expression (Fig. 4.14).

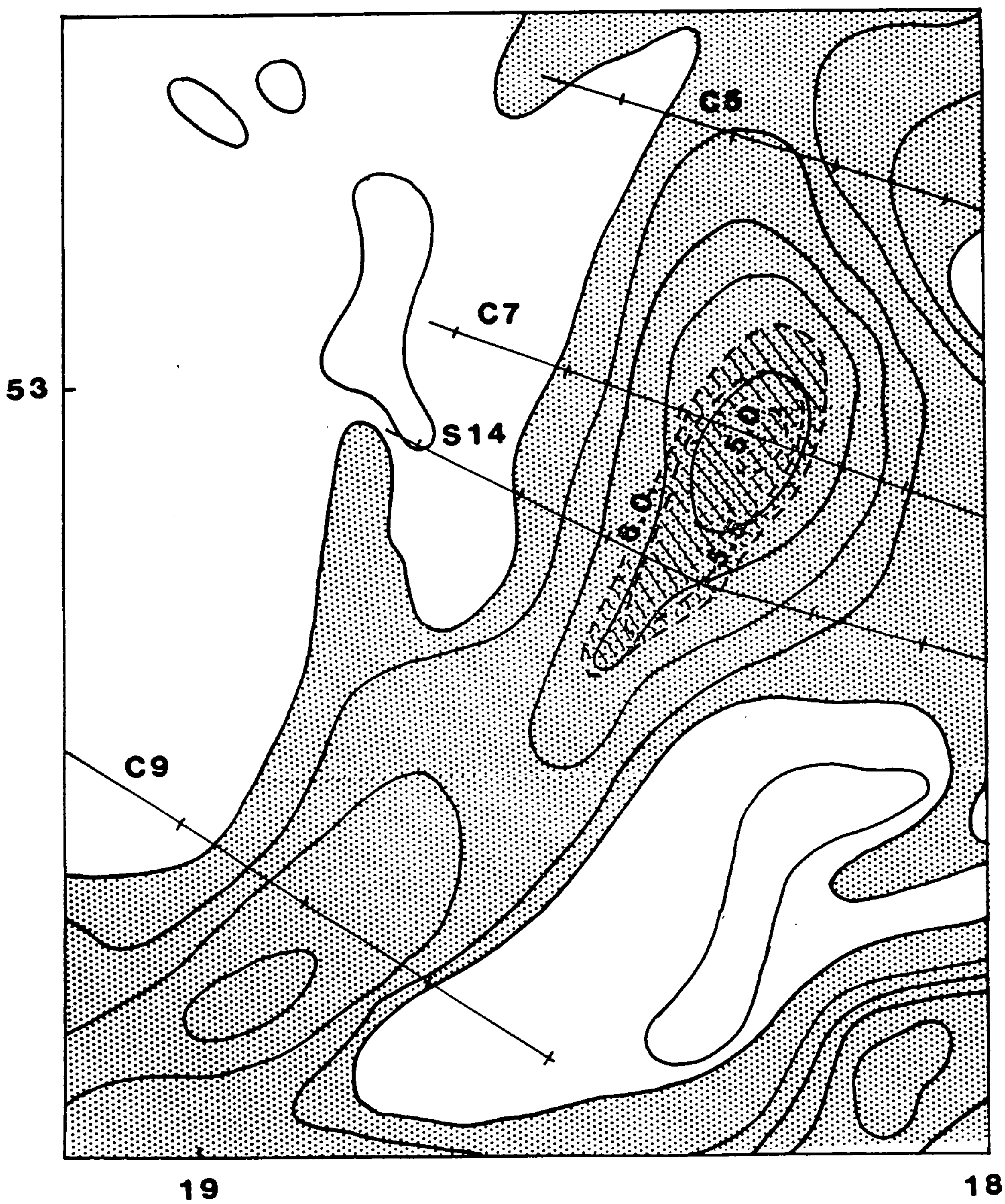


Fig. 4.12 Magnetic anomaly map of area A (refer to Fig. 4.9 for position) showing basement high (contours in seconds TWT) underlying the 400 gamma magnetic high

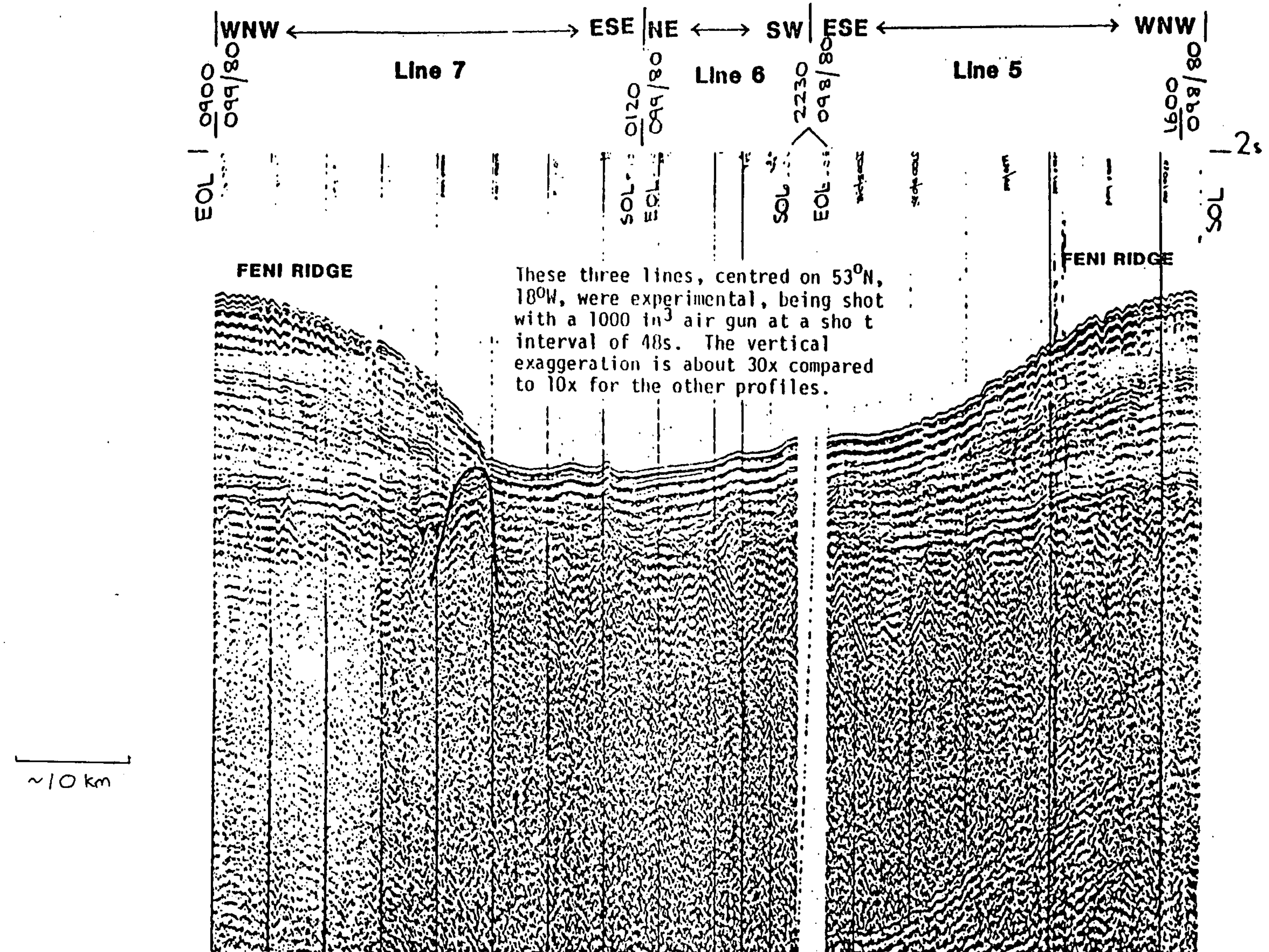


Fig. 4.13 Challenger lines 5,6, and 7. Refer to Fig. 4.12 for positions of lines 5 and 7

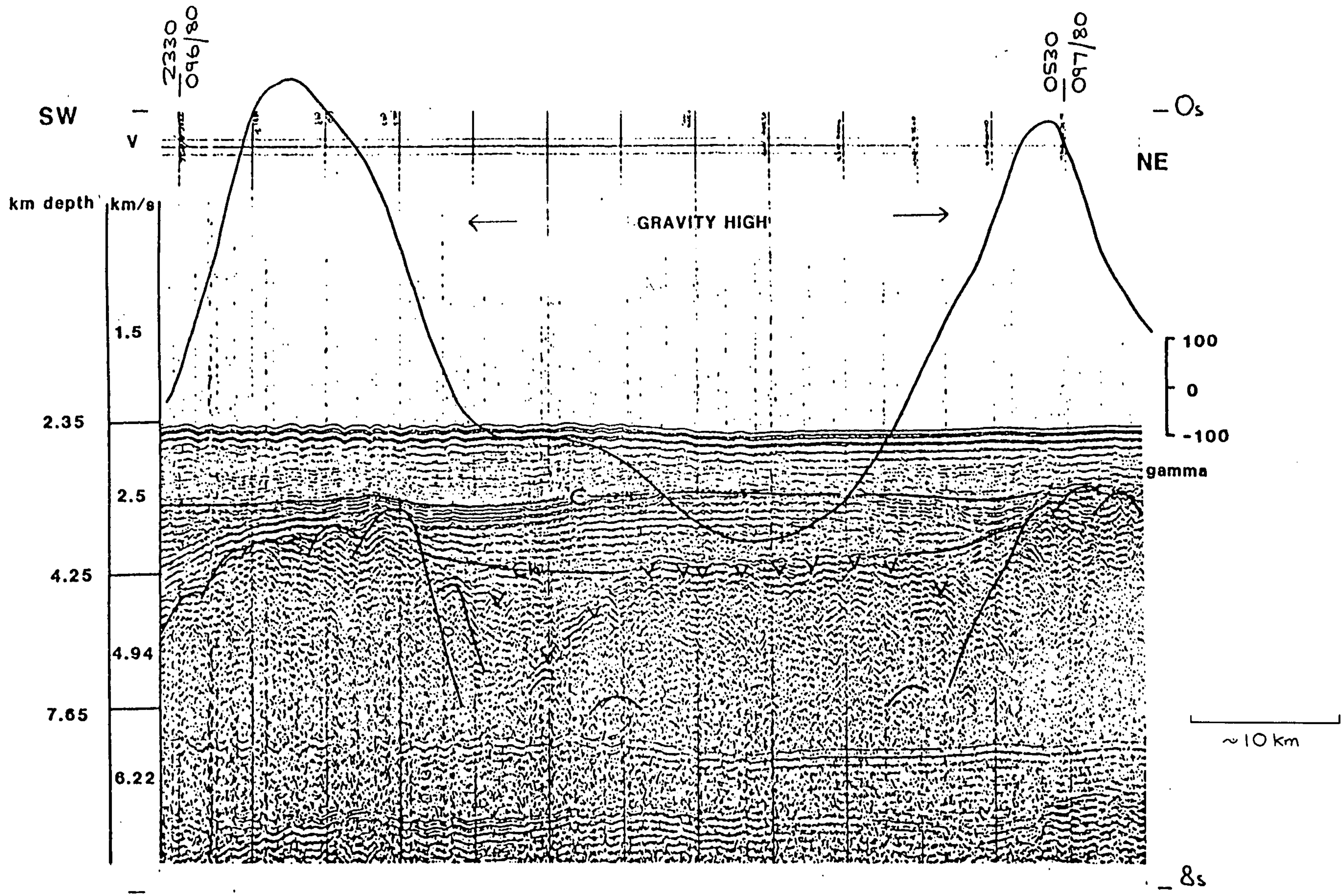


Fig. 4.14 The SW portion of Challenger line 2, with magnetic profile

C - Challenger horizon

Ch - Charcot horizon

V's show positions of sills

The results of Hills 1952 refraction experiment are shown (refer to section 4.2.4 c)

These basement highs can be seen to continue under the anomalies on Challenger line 4 (Fig. 4.21).

4.2.2.2 Magnetic Modelling

a) Roberts et al.,1981 (b)

The paper by Roberts et al. on southern Rockall Trough modelled normal and reversed magnetised blocks within the central noisy zone. The profiles for which modelling was done do not coincide with seismic lines, and the widths, topography, and magnetisations of the magnetised blocks were chosen arbitrarily. The distribution of magnetised blocks was not derived from a combination of spreading rate and the geomagnetic time scale, but was adjusted to give the best fit to the observed profiles.

All modelling parameters in this chapter are chosen after inspection of basement seismic character and depths.

b) Challenger line 1

The magnetic anomaly profile over basement topography derived from the seismic record is shown in Fig. 4.15.

Eastern Quiet Zone Modelling

There is no anomaly greater than 50 gamma in amplitude over the quiet zone and between the quiet zone and Porcupine Bank.

The topography of the quiet zone was used as input in a model to determine whether the underlying basement can be highly magnetic. Fig. 4.16 shows that a low magnetisation (G = 158 gamma) gives a good amplitude fit with the observed anomaly.

Modelling of a continent-ocean boundary between the quiet zone and the known continental crust of Porcupine Bank also indicated that a continent ocean boundary in this position would give a much larger than observed magnetic anomaly (Fig. 4.17).

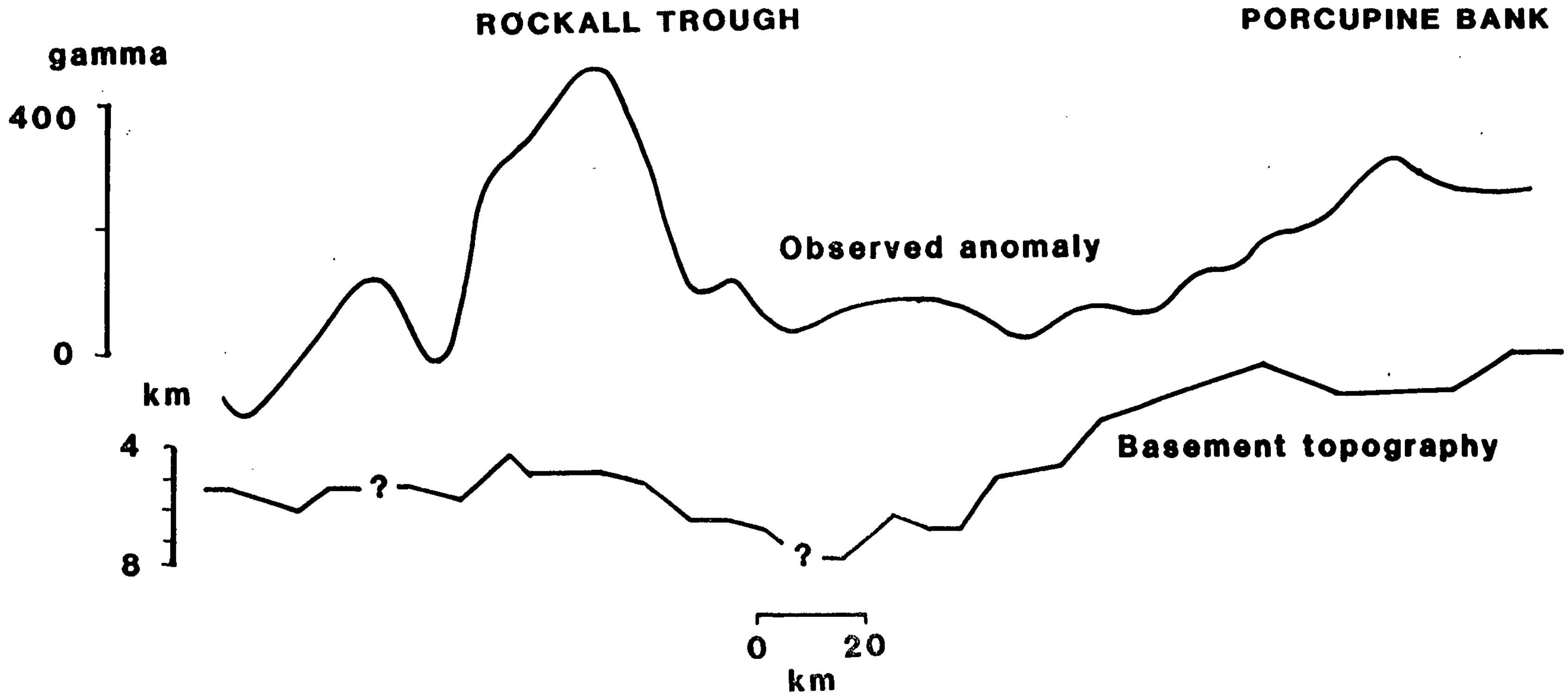


Fig. 4.15 Challenger line 1 modelled basement topography and magnetic profile

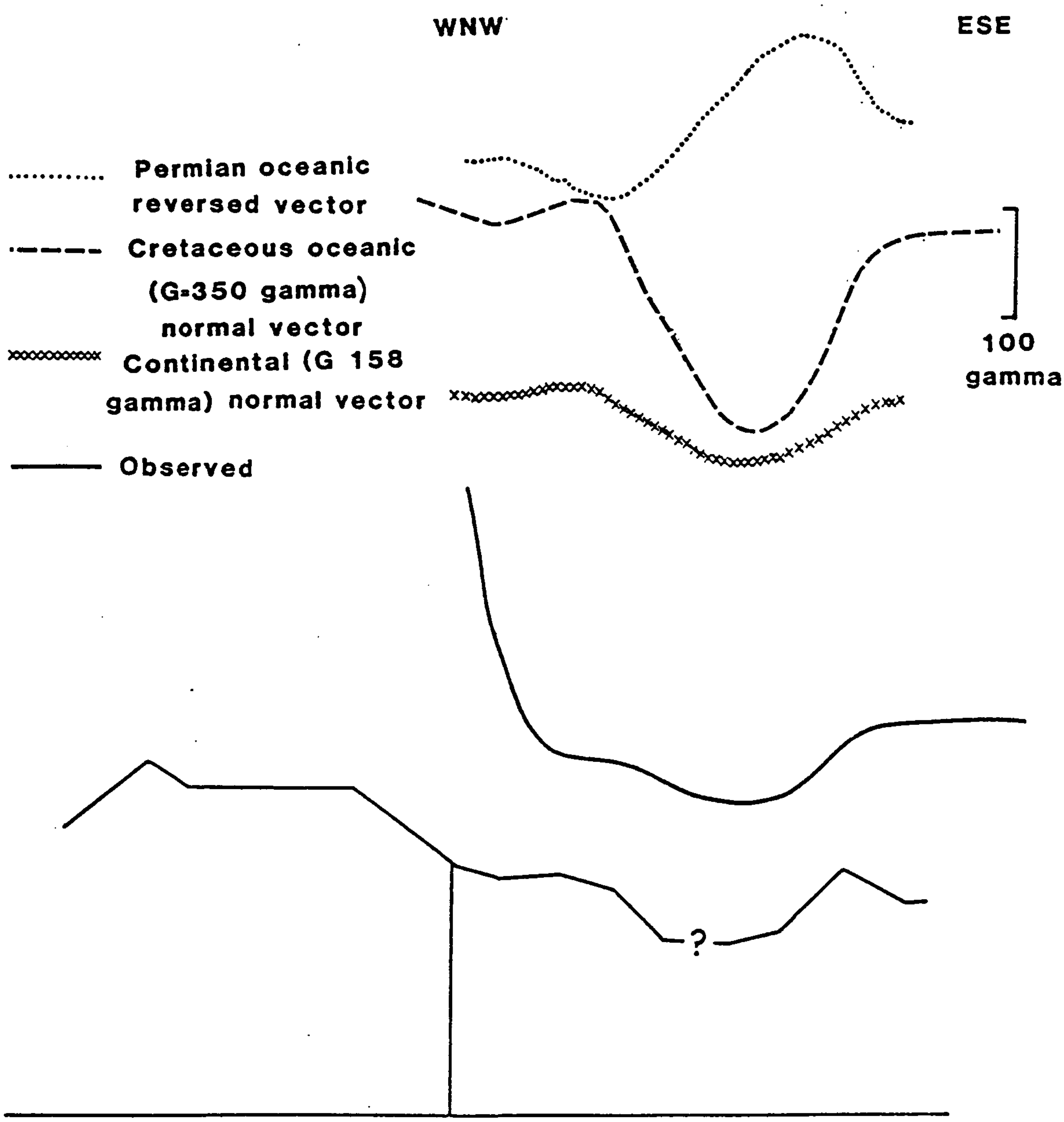


Fig. 4.16 Magnetic modelling of eastern quiet zone, Challenger line 1

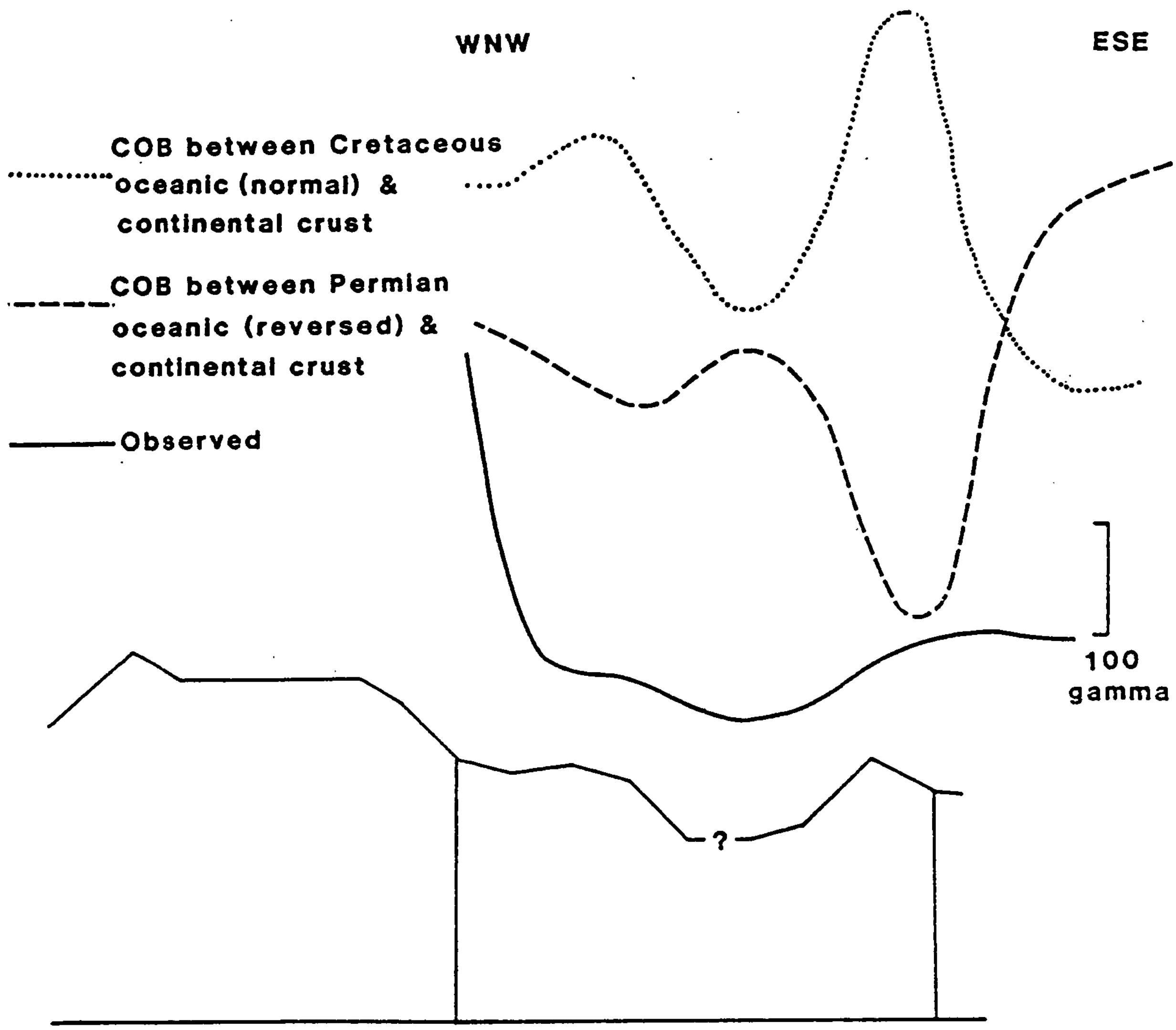


Fig. 4.17 Magnetic modelling of a continent-ocean boundary between the eastern quiet zone and Porcupine Bank, Challenger line 1. G = 350 gamma used for oceanic crust, and G = 158 gamma used for continental crust

Central Basement High

On Challenger line 1 basement is shallow and highly reflective between 1130z and 1400z/096 (Fig. 4.5). To the west of the high, basement appears to be at six seconds or deeper but is obscured by sills (discussed in section 4.2.4). This central basement high was assigned a magnetisation of 350 gamma in order to fit the observed curve (Fig. 4.18).

c) Shackleton line 14

There is a wider zone of high positive magnetic anomalies underlying Shackleton line 14, corresponding to a wider swathe of shallow and acoustically noisy basement.

The fit between observed and calculated curves is not as good as for Challenger line 1. However, a model which alternated deep, less magnetic prisms, with shallower, more magnetic prisms, again based on basement type and topography, produced a reasonable match of anomaly amplitude over the central noisy zone. (Fig. 4.19)

4.2.2.3 Summary

It has been demonstrated that magnetic reversals are not necessary in order to explain the anomaly pattern of central southern Rockall Trough. Indeed, the fact that all areas of negative magnetisation are underlain by deep basement and positive anomalies are underlain by shallow basement means that it is highly unlikely that reversals are present anywhere. The anomaly pattern is a result of a combination of magnetisation contrasts and topography.

The implication is that the whole of southern Rockall Trough can be considered to be underlain by thinned continental crust, affected by massive igneous intrusions. At a latitude of 53° 15'N, at the change of trend, the only igneous body is that modelled for Challenger line 1. Continental crust lies within the central noisy magnetic zone.

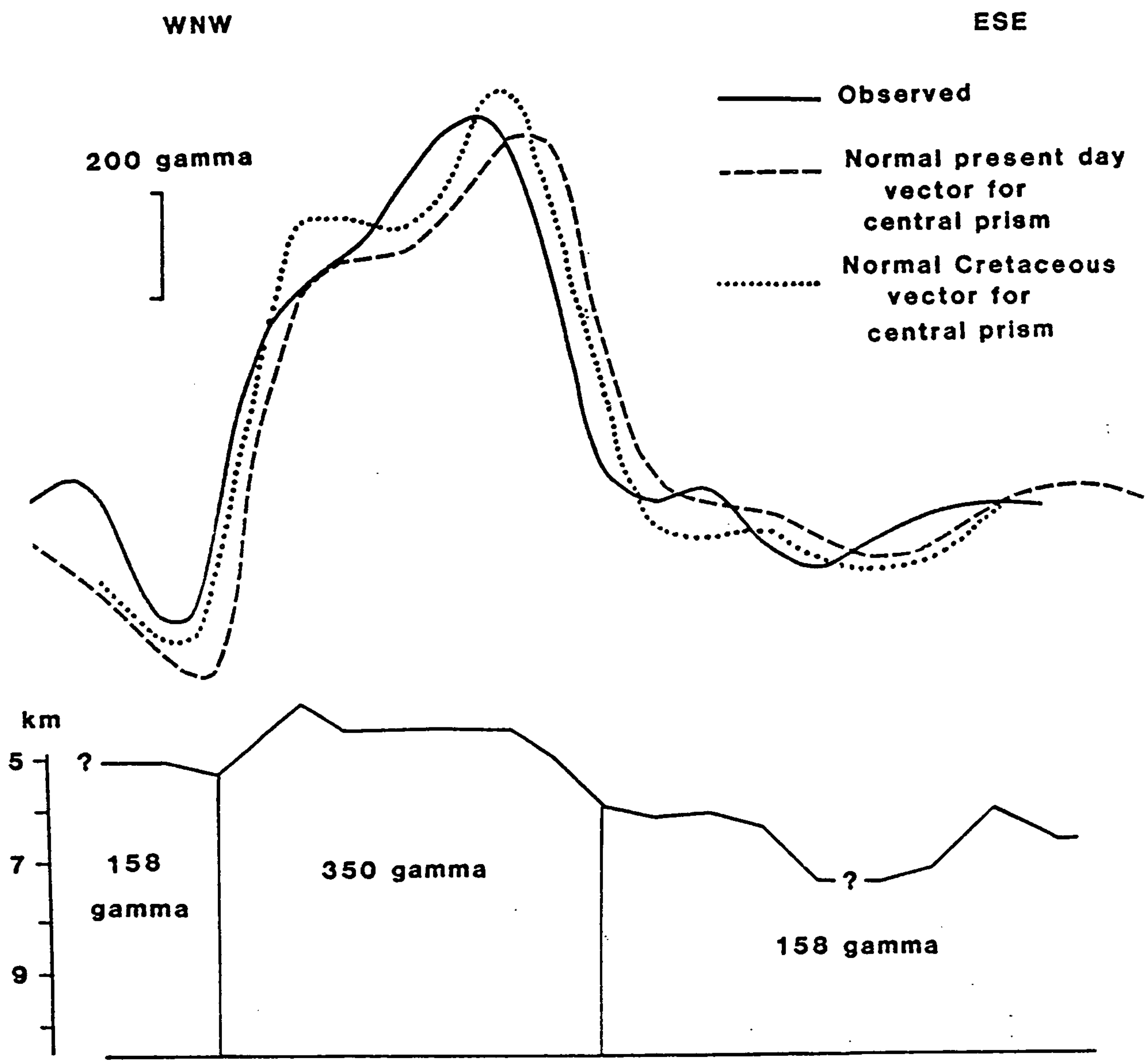


Fig. 4.18 Magnetic modelling of the central basement high and eastern quiet zone, Challenger line 1

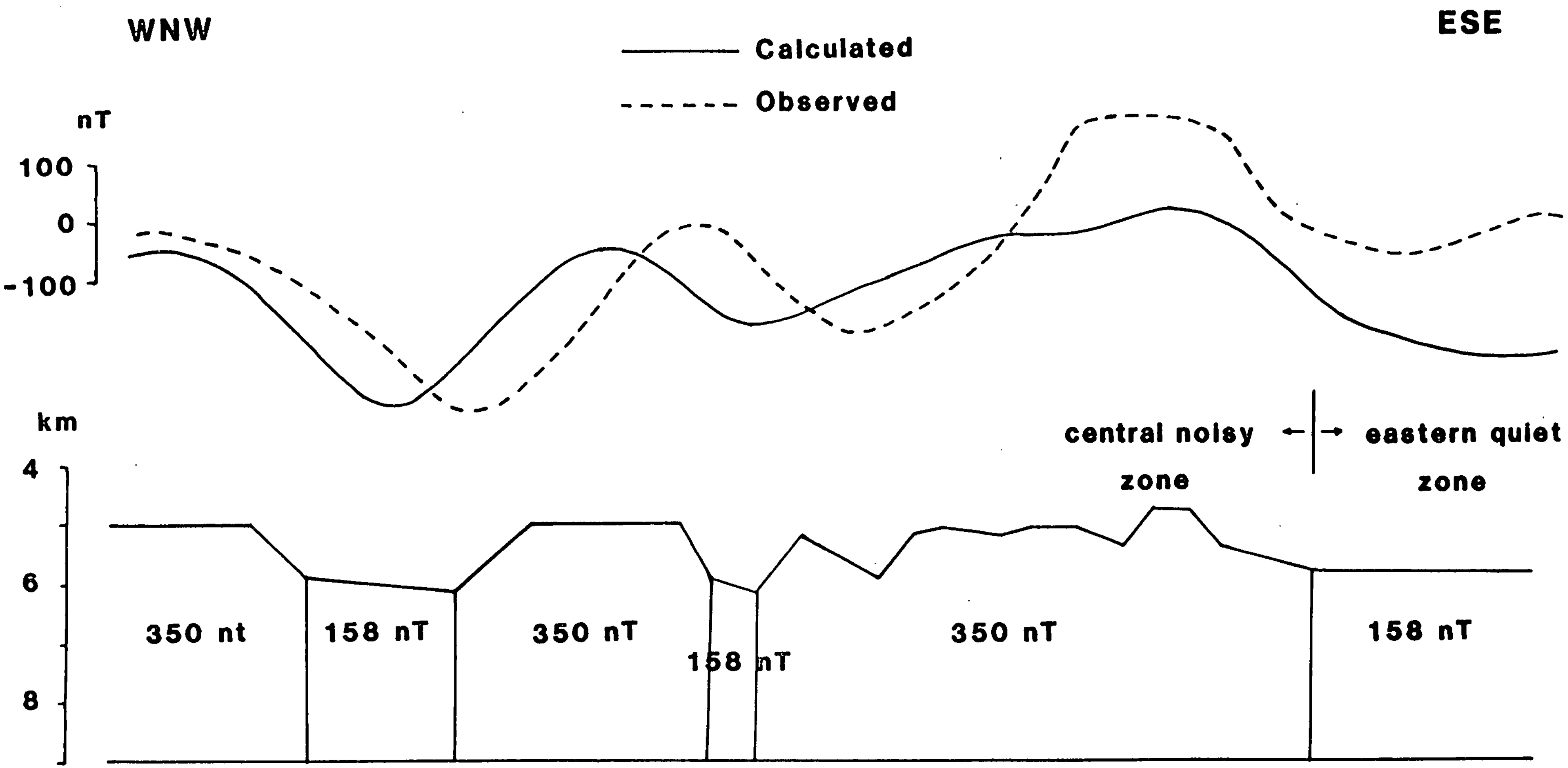


Fig. 4.19 Magnetic modelling of central noisy zone, Shackleton line 14

Comparisons could be drawn with the postulated igneous bodies which lie along the axis of the Faeroe-Shetland Trough (Ridd, in press).

The continuity of the magnetic anomalies with those of Southern Rockall Plateau further implies an origin by intrusion along fractures through continental crust.

4.2.3. Gravity Data

Figure 4.20 is the free air gravity anomaly map for southern Rockall Trough.

The most prominent anomaly is the 60-70 mgal anomaly beneath Challenger lines 1, 2, and 3.

a) Gravity Modelling and the 60 mgal Anomaly

An isostatic crustal model was constructed for Challenger line 1 (Figs. 4.6 and 4.7).

Basement depth is poorly constrained where deepest, but minimum crustal thicknesses of approximately 5 km are indicated.

The 60 mgal anomaly at the western end of Challenger line 1 is grossly out of isostatic equilibrium. The position of this anomaly is in the centre of the noisy central zone (Fig. 4.10). A portion of it underlies the basalts seen between the two igneous ridges on Challenger line 2 (Fig. 4.14).

The shape of the anomaly makes 2D modelling inapplicable, and it was not attempted. The source of the anomaly must be a roughly circular mantle 'bulge', and/or a high density intrusion within the crust. The intrusion must be at a depth below the Curie point isothem, as it has no magnetic expression. In either case, it is expected that the gravity anomaly is an expression of deep-seated asthemospheric breakthrough associated with the flanking massive igneous bodies. n/

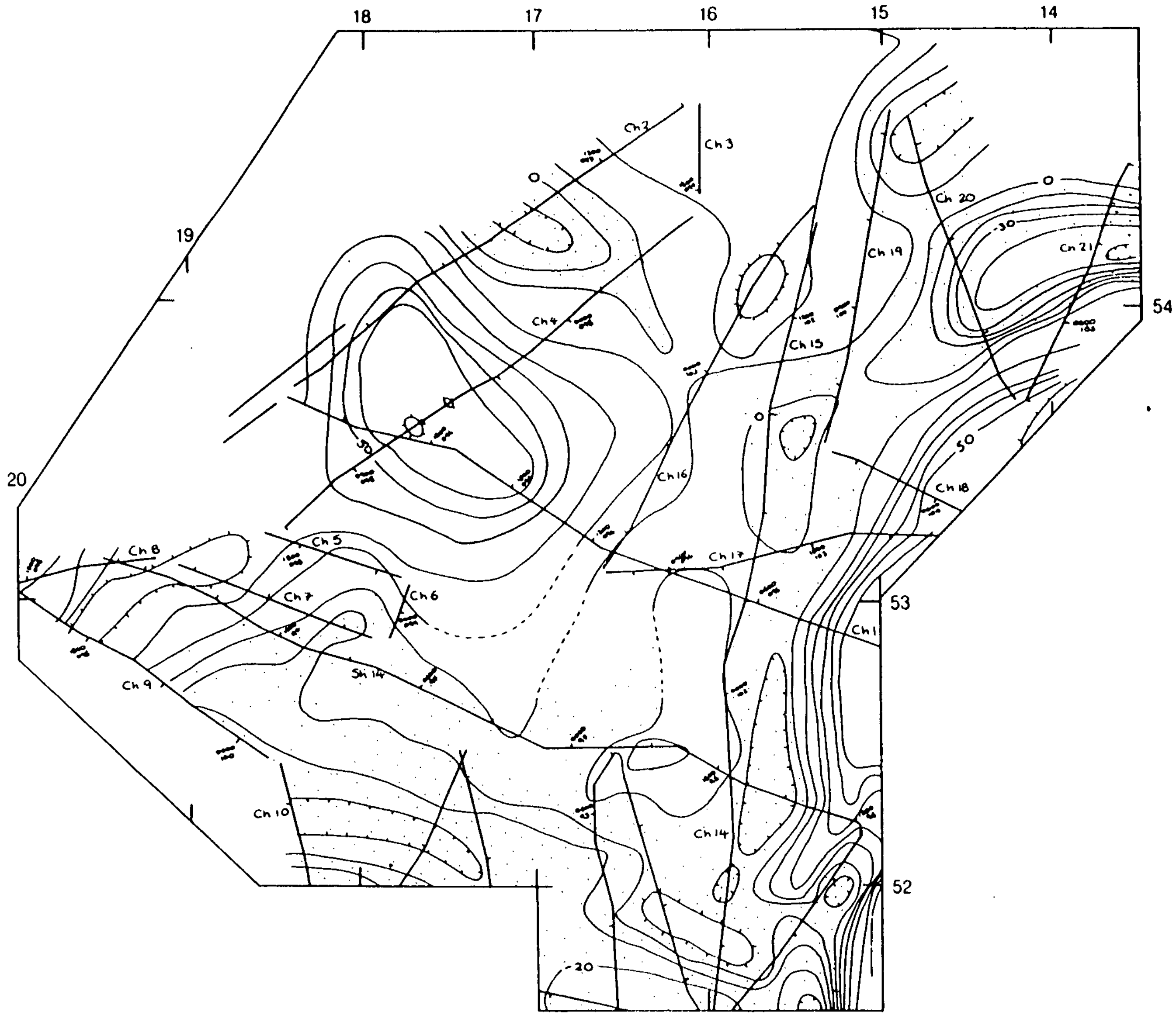


Fig. 4.20 Free air gravity anomaly chart of southern Rockall Trough. Edinburgh tracks superimposed. Contour interval 10 mgal. Area where FAA is +10 mgal or less is shown stippled

4.2.4 Seismic Data

Figure 4.2 1 is the time to basement chart of southern Rockall Trough, compiled from Edinburgh data. Points previously raised were the shallower basement beneath the central noisy magnetic zone, and the concentration of sills within the section over the gravity high. A sediment thickness chart is given in Dingle, Megson and Scrutton (Appendix B).

a) Sills

These are distinguishable from basement events by their bright and discontinuous nature. Commonly flat, the sills are occasionally steeply dipping. Diffractions from both ends of the sill give it a characteristic appearance. Where sufficiently discontinuous, true basement can be seen beneath the sills. Sediments beneath the level of the sills are obscured but show no sign of drape over a high.

A comparison is drawn with the sills described in the Faeroe-Shetland channel by Ridd (in press).

b) The northern and western magnetic quiet zones

Basement in these two areas is commonly invisible as it lies below the first water bottom multiple. Sediment thicknesses are greater than in the eastern quiet zone due to the sediment build-up of the Feni Ridge: locally more than 4 seconds (TWT) of sediment can be seen. It is believed that the evolution of these areas was roughly the same as that of the eastern quiet zone until the build up of Feni Ridge. Feni Ridge was built up by the flow of overflow water from the Norwegian basin in the ?Eocene. In the south of Rockall Trough these currents were controlled by the previously intruded igneous ridges.

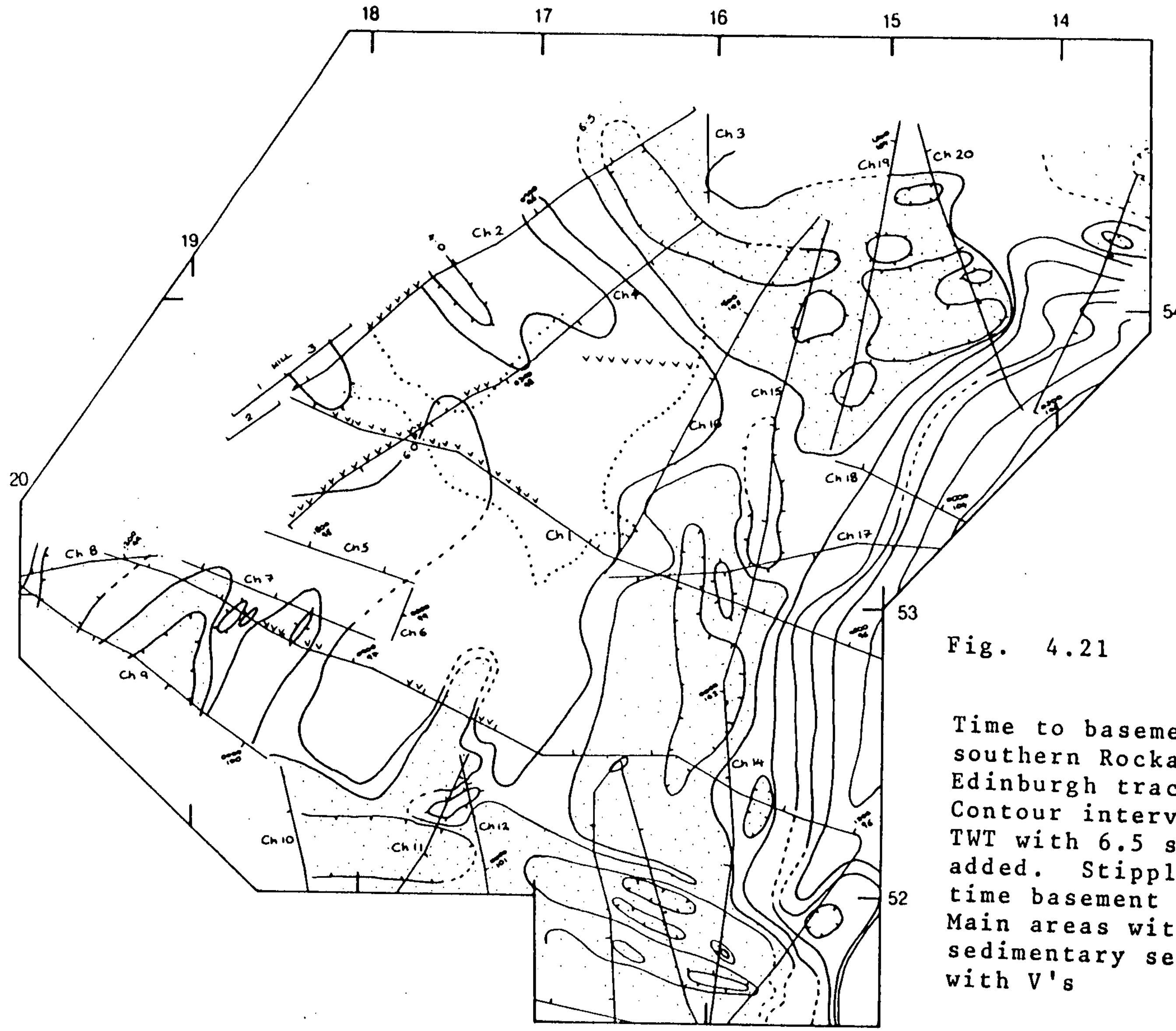


Fig. 4.21

Time to basement chart of southern Rockall Trough. Edinburgh tracks superimposed. Contour interval 1.0 seconds TWT with 6.5 seconds contour added. Stipple over area where time basement is 6.5 seconds. Main areas with sills within the sedimentary section are indicated with V's

c) Hill's Seismic Refraction Experiment

Three seismic refraction stations were occupied in southern Rockall Trough (Hill 1952). The positions of these stations are shown in Figs. 4.20, 4.21, and 4.22. Station 3 runs parallel to Challenger line 2 and Hill's layer velocities and depths are shown on the Challenger line 2 seismic section (Fig. 4.14). The results of all three stations are given in Fig. 4.23.

Refraction stations 1 and 2 are positioned over a magnetic low, and by implication a basement low. Station 3 runs across the highly magnetic shallow igneous ridge crossed by Challenger line 2. This explains the difference in depth to layer 2 between stations 1 and 2 and station 3. Hill wrote, with reference to station 3:

> " ... the thickness of layer 1 is considerably less than at the other two stations ... this conclusion is geologically difficult to interpret ... the shots for stations 1 and 3 are practically collinear, with the buoy positions of the latter station being close to the more distant shots of the former. No evidence for a component of dip in the direction of the line was obtained from either station 1 or station 2".

It can now be seen that the northeastern end of station 1 and the southwestern end of station 3 are practically coincident with the edge of the igneous ridge.

The experiment gives a velocity of 4.94 km/s for the igneous bodies, and 6.3 km/s for the crust.

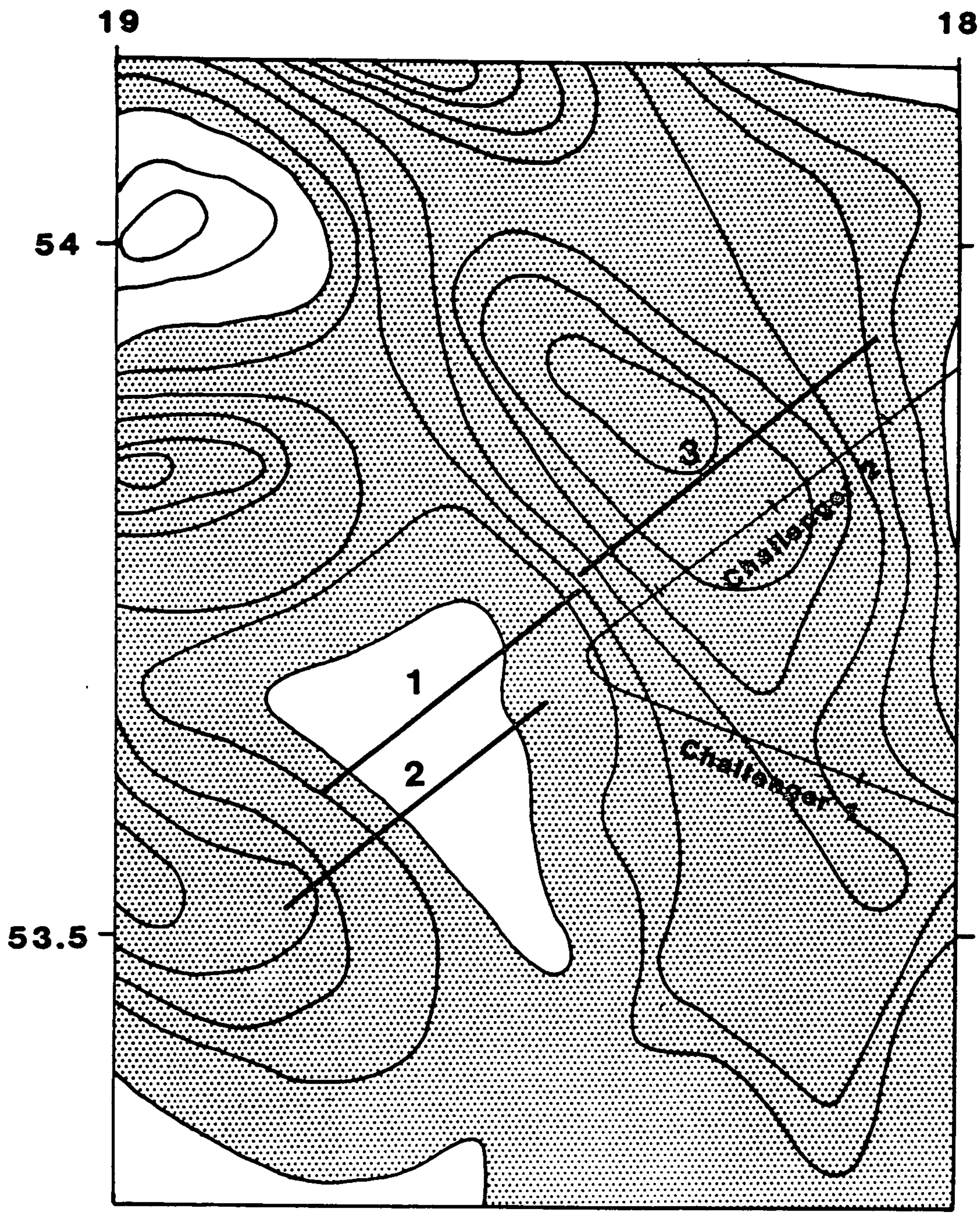


Fig. 4.22 Positions of three seismic refraction stations in southern Rockall Trough (Hill, 1952). This area is shown as area B in Fig. 4.9. Magnetic anomaly contours shown: contour interval 100 gamma

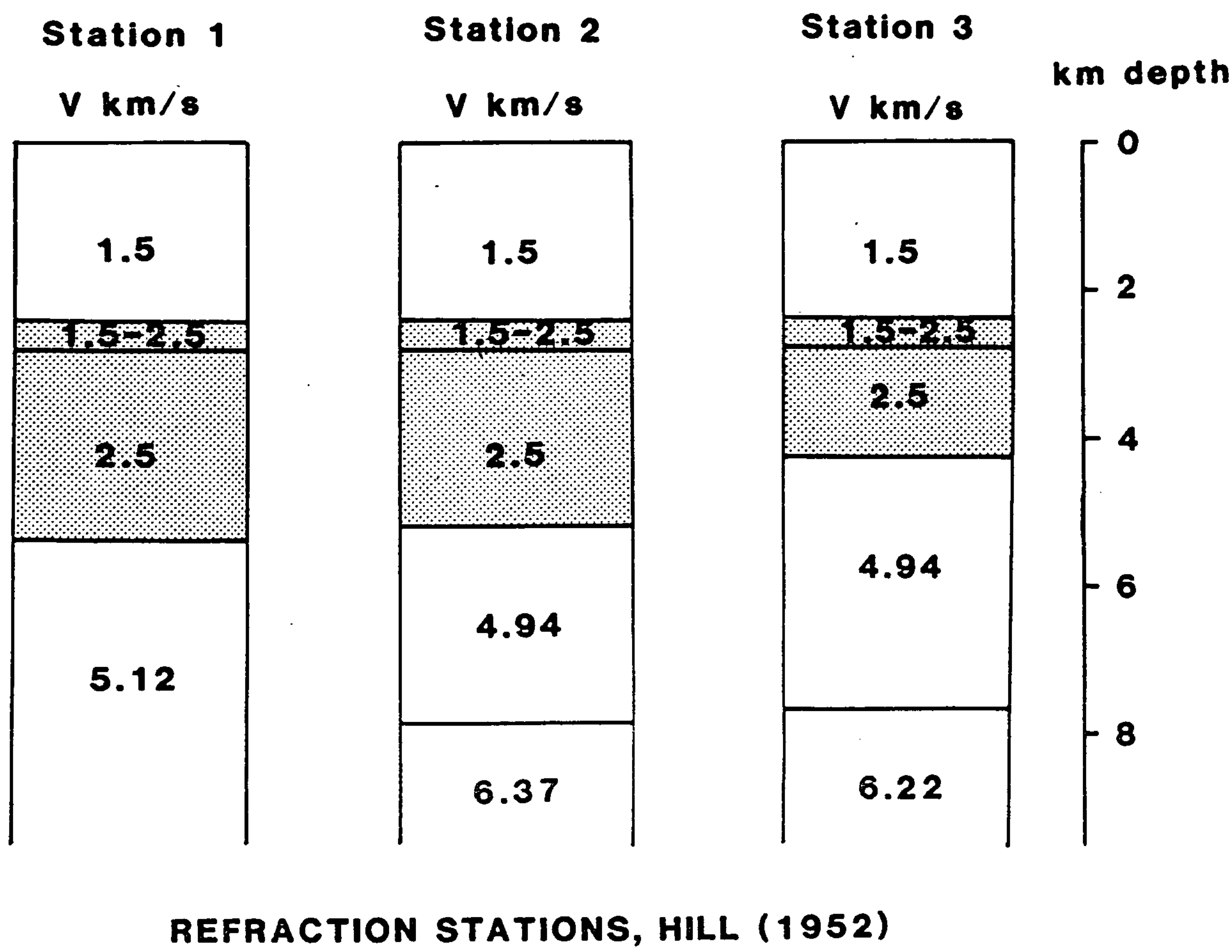


Fig. 4.23   Refer to Fig. 4.22 for position

4.2.5 Conclusions

It is postulated that the igneous intrusions in southern Rockall Trough are associated with the initial formation of the Charles Gibbs Fracture Zone and first oceanic crust to the south at anomaly 32, 32-33 time.

The history of southern Rockall Trough is concluded to be as follows:

i) ?Pre Jurassic basin(s) on Caledonian basement

ii) Jurassic rifting and basin formation associated with Atlantic formation to the south; cf Greenland, North Sea, Newfoundland etc.

iii) Further Cretaceous crustal stretching as the lithospheric crack propogates northwards and spreading starts in Biscay and off Goban Spur

iv) In the Upper Cretaceous, massive igneous intrusions (asthenospheric breakthrough) and fracturing associated with the formation of oceanic crust in Porcupine Abyssal Plain and immediately to the west of southern Rockall Trough

v) Subsidence (?) followed by Palaeocene uplift, and possibly further intrusive activity, due to the splitting of Greenland from Rockall Plateau

vi) the formation of Feni Ridge

vii) subsidence throughout the Tertiary

CHAPTER 5

THE CLARE LINEAMENT AND THE CHARLIE GIBBS FRACTURE ZONE

The Charlie Gibbs Fracture Zone is the most significant Northern Atlantic fracture zone. At its eastern extremity it separates the central magnetic noisy zone of Rockall Trough from the earliest formed oceanic crust of Porcupine Abyssal Plain. In this chapter the nature and early evolution of the eastern Fracture Zone are described.

The Clare Lineament is offset to the south and east from the Charlie Gibbs Fracture Zone. It lies between the eastern magnetic quiet zone of Rockall Trough, and the thinned continental crust west of southern Porcupine Bank. The Lineament appears to be contiguous with the Clare Trend (Chapter 3) which separates basement of different types on Porcupine Bank.

5.1 THE CLARE LINEAMENT

5.1.1 Seismic Character

Four Edinburgh lines cross the Clare Lineament; Shackleton lines 8, 9, and 13, and Challenger line 14 (Fig. 5.1). Challenger line 14 is shown in Fig. 5.2.

To the south of the Clare Lineament, and associated with the gravity low, basement is 7.5-8 seconds deep. The Clare Lineament itself is a WNW-ESE trending basement ridge at a depth of 6 seconds on Challenger line 14 with further basement highs and lows to the north. The difference in depth between the highs and lows to the north is 1.5 seconds or more. These features are not precisely correlatable from line to line and are either discontinuous features or en echelon ridges. The Lineament is a bathymetric feature in addition. The seafloor to the north is shallower partly due to sediment drap over basement at shallower depth, and partly due to a sedimentary section which is thicker overall. This feature is named the Clare Terrace (Appendix B).

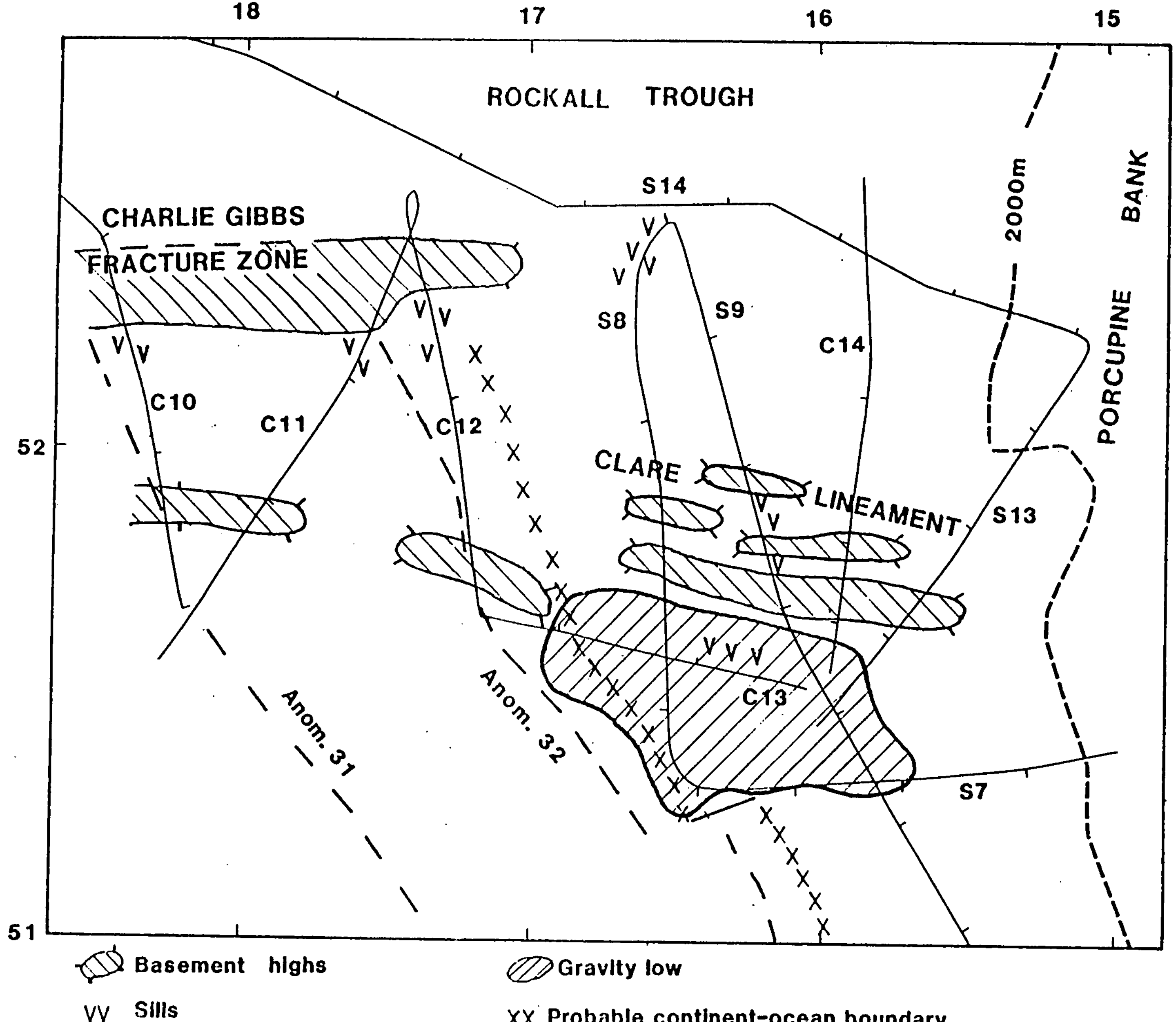


Fig. 5.1 Schematic structural map of the Charlie Gibbs Fracture Zone and the Clare Lineament. Edinburgh lines added.

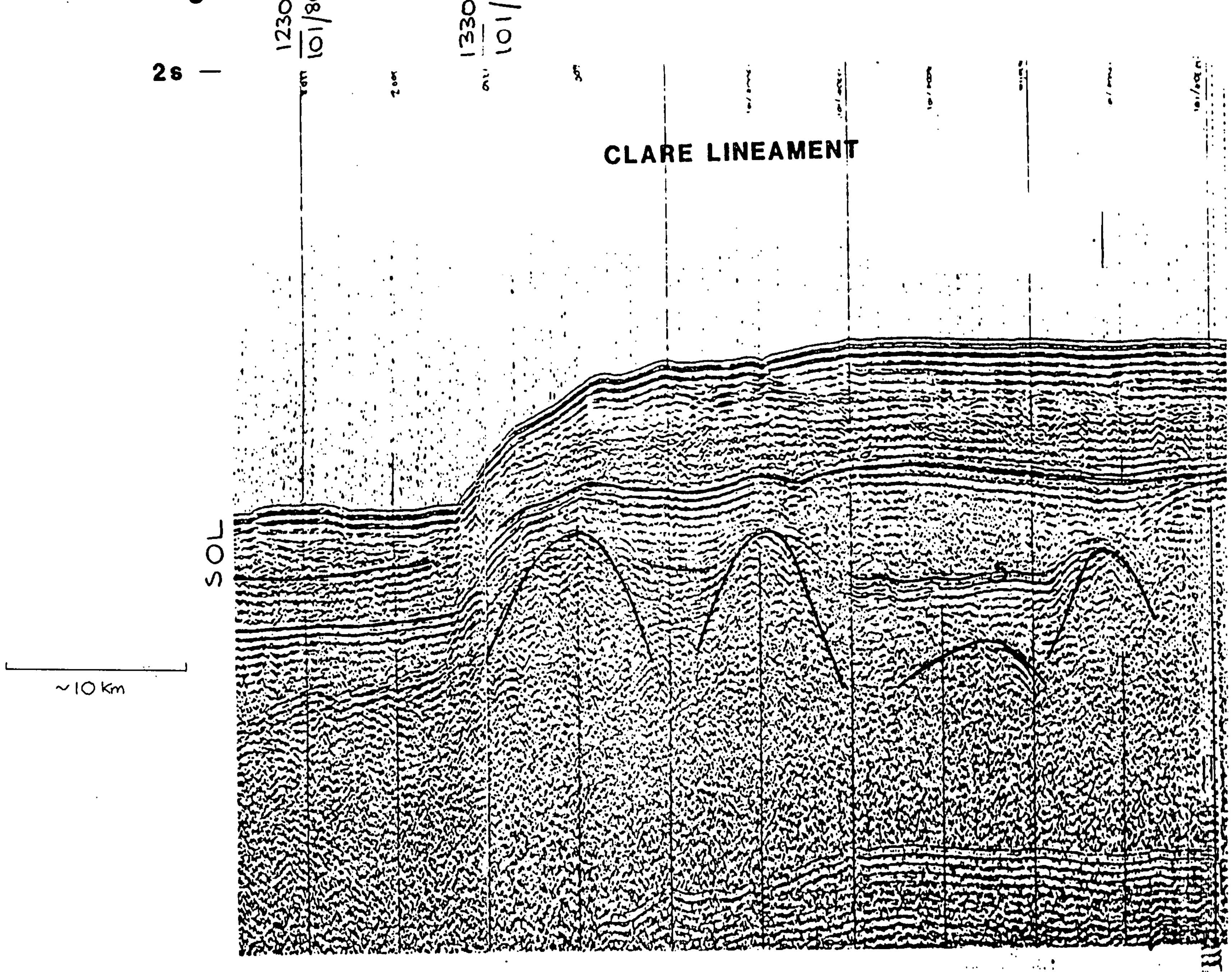


Fig. 5.2 Interpretation of Challenger line 14 across the Clare Lineament. Vertical scale is 2 seconds to 10 seconds (TWT)

To the west, the difference in elevation between the basement of the Clare Lineament high and the basement to the south decreases to 0.5 seconds or less. The highs and lows to the north deepen similarly.

5.1.2 Magnetic Character

A magnetic chart over the Clare Lineament and Charlie Gibbs Fracture Zone is given in Fig. 5.3. There are no anomalies of the same trend associated with the WNW-ESE trending major basement features of the Clare Lineament. This implies that the basement has a low magnetisation.

Clare Lineament itself and the quiet zone to the north are characterised by negative magnetic anomalies. Immediately to the south the magnetic field is highly complex (refer to magnetic profile of Shackleton line 7, Chapter 3), and predominantly positive.

This agrees with the hypothesis that the Clare Lineament and Clare Trend mark the boundary between two different geological provinces.

5.1.3 Gravity and Gravity Modelling

Figure 5.4 is the free air gravity anomaly map of the area.

A +10 mgal gravity high overlies the Clare Lineament, and a -30 mgal gravity low is associated with the deep basement to the south. The steepest gradient between the high and the low approximately coincides with the basement high of the Clare Lineament or with the southern boundary of the basement high.

Gravity models were constructed for Challenger line 14 (Figs. 5.2 and 5.5) and for Shackleton line 9 (Fig. 5.6).

Both show that the area of deep basement to the south of the Clare Lineament is out of isostatic equilibrium. The observed gravity low can be modelled by thickening the crust by approximately 2 km.

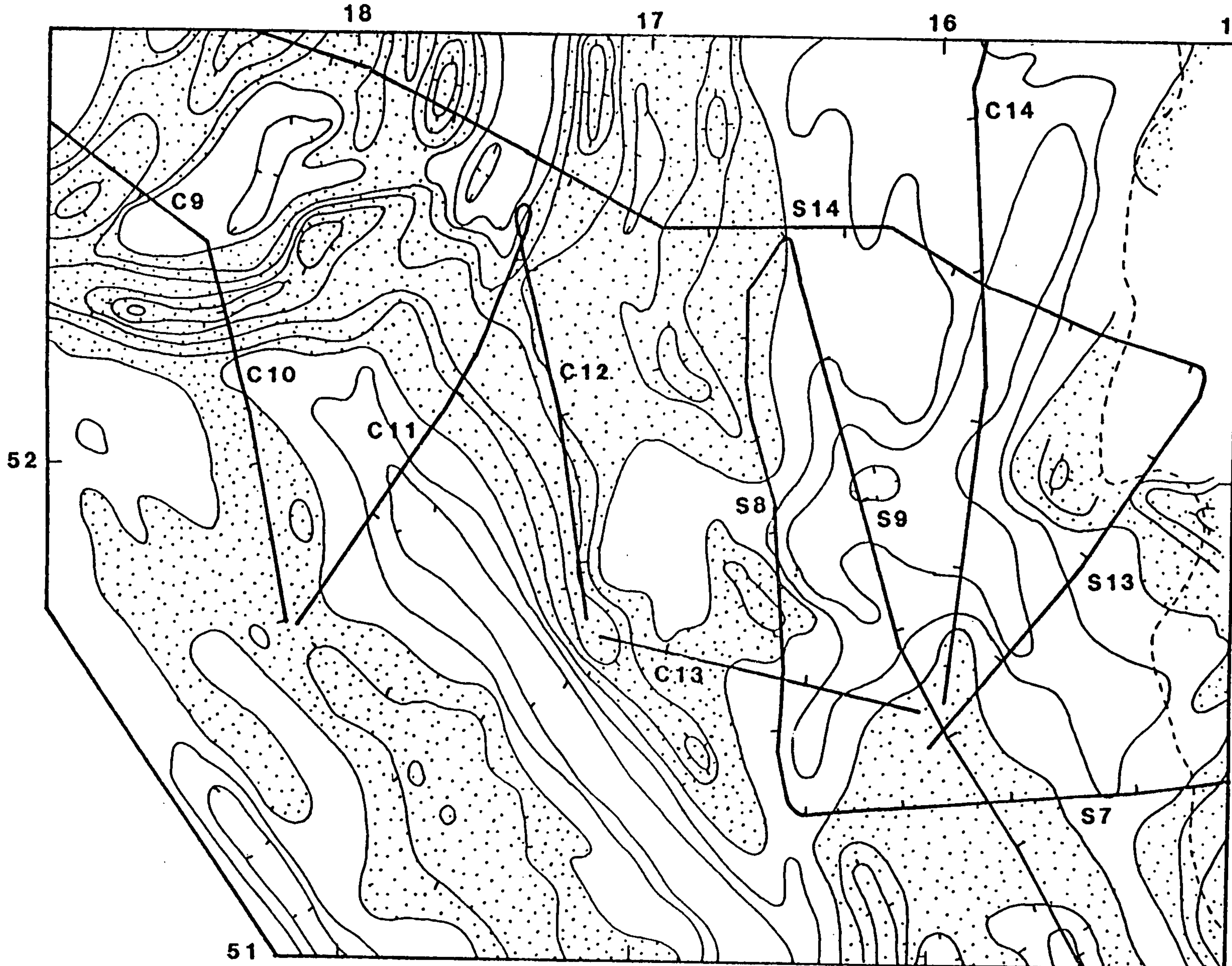


Fig. 5.3 Magnetic anomaly map of the Clare Lineament - Charlie Gibbs Fracture Zone area. Contour interval 100 gamma. Stippled area: magnetic anomaly >100 gamma

Fig. 5.4 Free air gravity anomaly map of the Clare Lineament - Charlie Gibbs Fracture Zone area. Contour interval 10 mgal. Stippled area : free air gravity anomaly >0 mgal

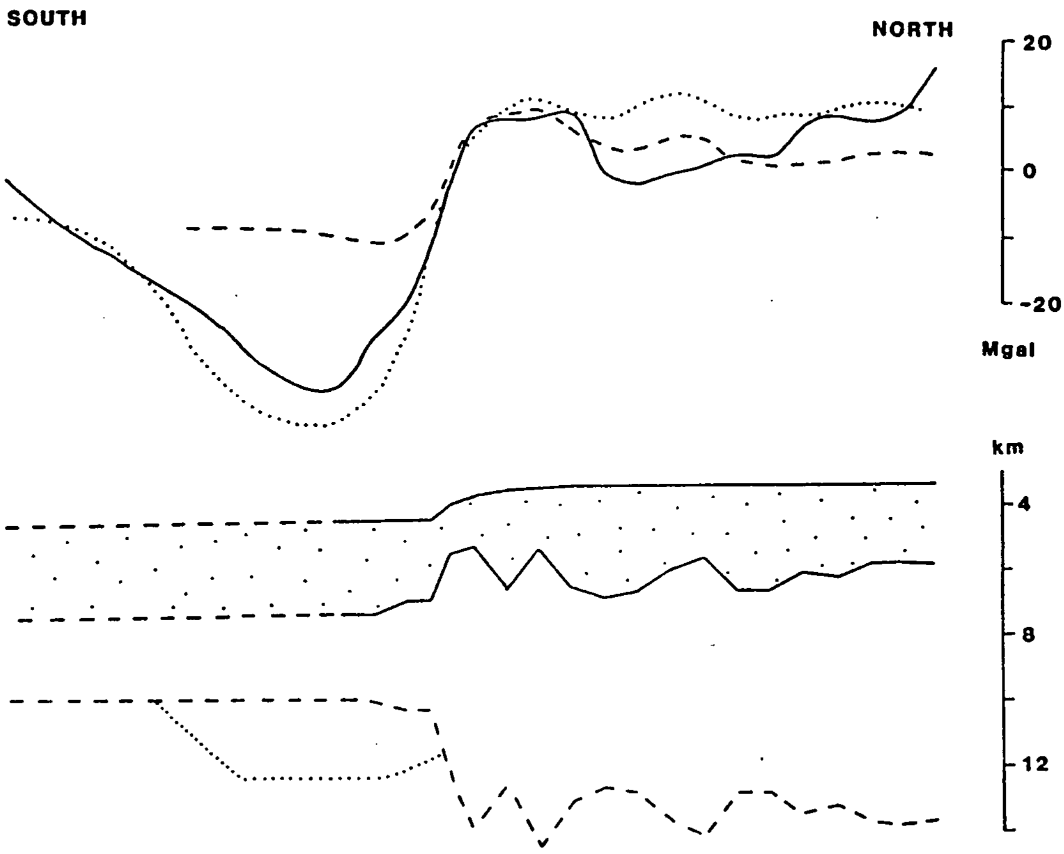


Fig. 5.5 Gravity model of Challenger line 14 over the Clare Lineament.

Solid line - observed anomaly

Dashed line - predicted anomaly from isostatic crustal model (dashed line, crustal section)

Dotted line - predicted anomaly from adjusted crustal model

Stippled area - sediments

Densities are 2.2 g/cm³ for sediments, 2.76 g/cm³ for crust, and 3.38 g/cm³ for mantle.

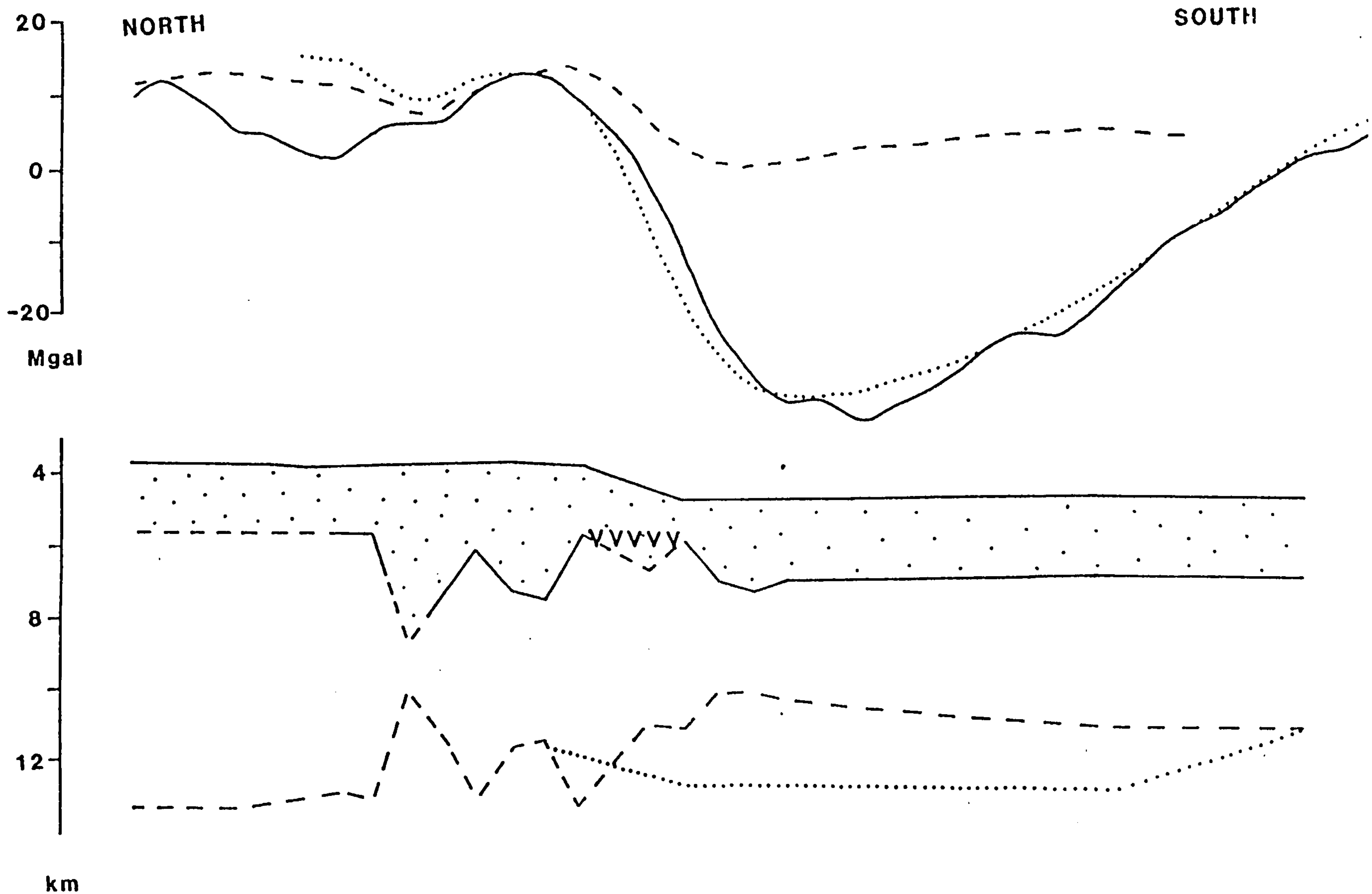


Fig. 5.6 Gravity model of Shackleton line 9 over the Clare Lineament
Key as for Fig. 5.5

5.1.4 The Nature of the Clare Lineament

It is postulated that the Clare Lineament marks the boundary between Caledonian basement (to the north) and Avalonian basement (to the south). It is expected that the Lineament formed as a result of different responses to stretching of the Caledonian and Avalonian basement. The low basement to the south of the Lineament may be a consequence of shear stresses associated with early stretching causing a segment of unsupported crust to founder.

5.2 THE CHARLIE GIBBS FRACTURE ZONE

Data over the Charlie Gibbs Fracture Zone is limited, as only Challenger line 10 crosses the Fracture Zone proper. Challenger lines 11 and 12 traverse the area where the continent ocean boundary is thought to intersect the Fracture Zone. The magnetic map also suggests that this is an anomalous area, and therefore Challenger line 10 is the only line modelled.

5.2.1 Seismic Character

There is little topographic expression of the Fracture Zone on Challenger line 10 (Fig. 5.7).

Close investigation of the seismic line reveals a basement high from 0330 to 0400z/100, with basalts overlying deeper basement immediately to the south. At 0700z/100 there is an emergent ridge of oceanic basement.

Challenger lines 11 and 12, to the east of the Charlie Gibbs Fracture Zone, cross a pronounced basement ridge. Figure 5.8 shows part of Challenger line 12. Immediately to the north of the basement ridge is a lens of acoustically near transparent sediments. The ridge has obviously influenced current flow and deposition.

5.2.2 Magnetic Character and Magnetic Modelling.

The Charlie Gibbs Fracture Zone south of Rockall Trough, and to the west as far as anomaly 24 in Porcupine Abyssal Plain, is characterised by an east-west trending positive magnetic anomaly. In the area crossed by Edinburgh profiles, this magnetic anomaly is up to 500 gamma in magnitude.

To the west of anomaly 24, the Fracture Zone is characterised by its very flat, near-zero, magnetic signature.

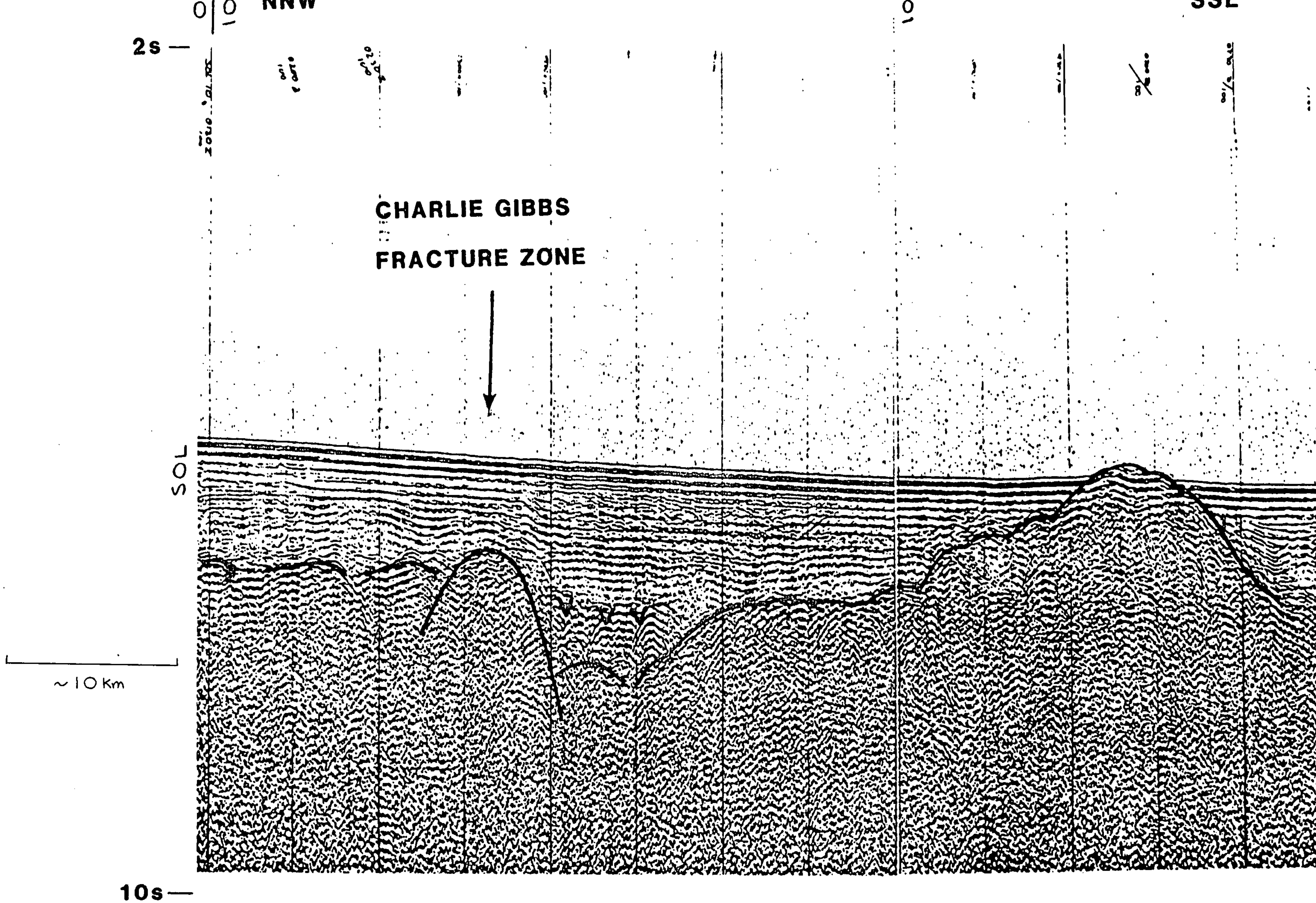


Fig. 5.7   Interpretation of Challenger line 10

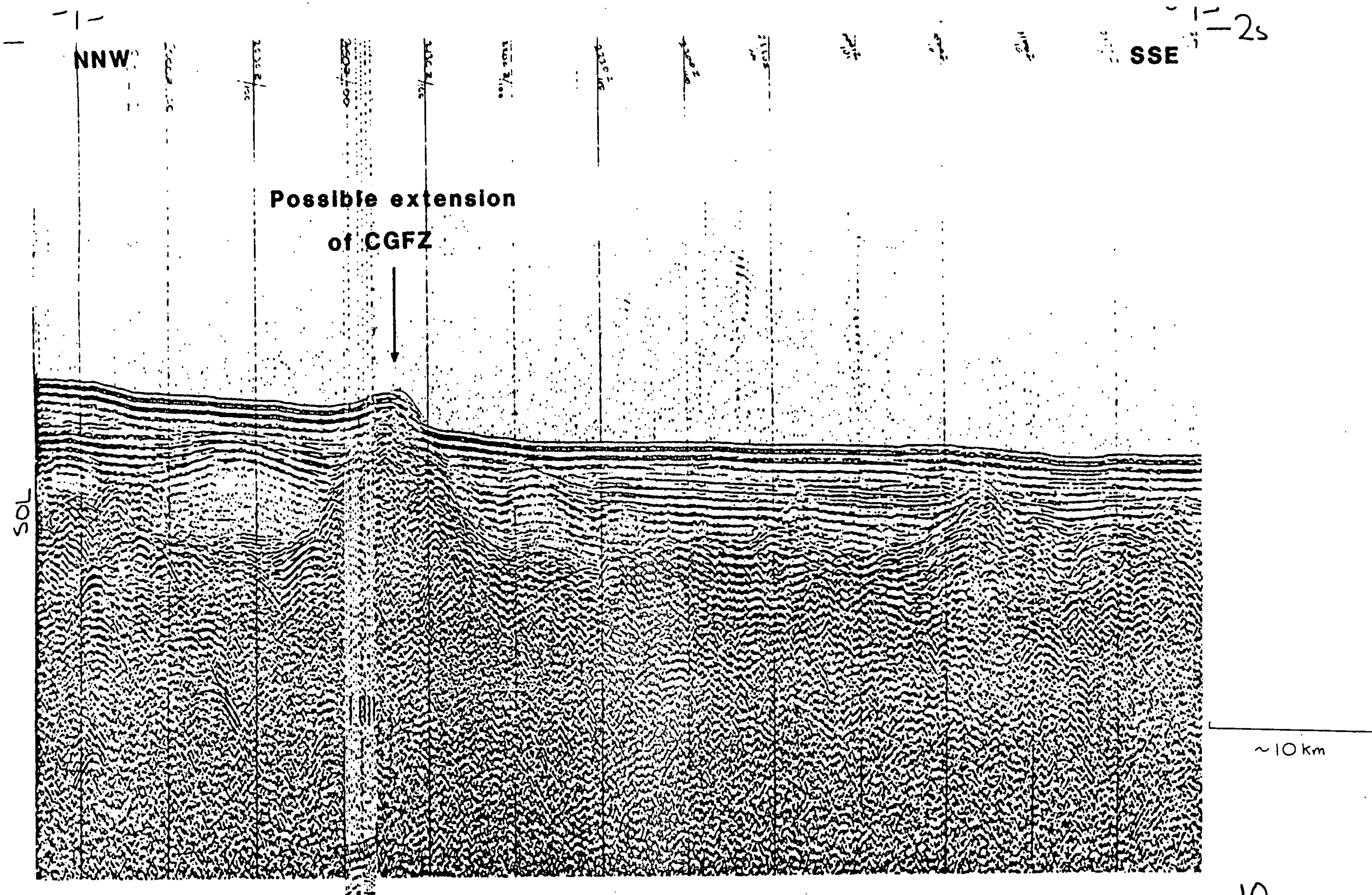


Fig. 5.8 Interpretation of Challenger line 12

A magnetic model was constructed of Challenger line 10 (Fig. 5.9). The basement high at the Fracture Zone was assigned a high magnetisation of 500 nT (= 550 gamma) and a reasonable fit was produced. The oceanic crust to the south was assigned a low magnetisation of 158 gamma as this line runs across, or near, the boundary between reversely magnetised ocean crust (anomaly 31-32) and normally magnetised ocean crust (anomaly 32).

5.2.3 Gravity and Gravity Modelling

A 10 mgal low is associated with the Charlie Gibbs Fracture Zone. Modelling of Challenger line 10 (Fig. 5.9) demonstrated that the features of the Fracture Zone are out of isostatic equilibrium. The gravity mismatch could be explained either by a region of less dense crust at the Fracture Zone, or a thickened crust, as shown in Fig. 5.9. The ridge to the south of Challenger line 10 is uncompensated, as is normal for oceanic crust.

5.2.4 Anomaly 24, and the Charlie Gibbs Fracture Zone as a Leaky Transform Fault

As mentioned earlier, the portion of the Fracture Zone between anomalies 24-32 is unusual in that it produces a strong positive magnetic anomaly. The modelling of Challenger line 10 shows that this can be modelled as a highly magnetic body, associated with a basement ridge.

Anomaly 24 is the oldest seafloor spreading anomaly between eastern Greenland and Rockall Plateau, and between western Greenland and Canada. Anomaly 24 in Porcupine Abyssal Plain is due south of anomaly 32 in the spreading sector southwest of Rockall Plateau. It was at anomaly 24 time, therefore, that the Charlie Gibbs Fracture Zone changed from a transform fault in continental crust, linking two separate spreading sectors, to a fracture zone within oceanic crust. (This point is illustrated in Chapter 6).

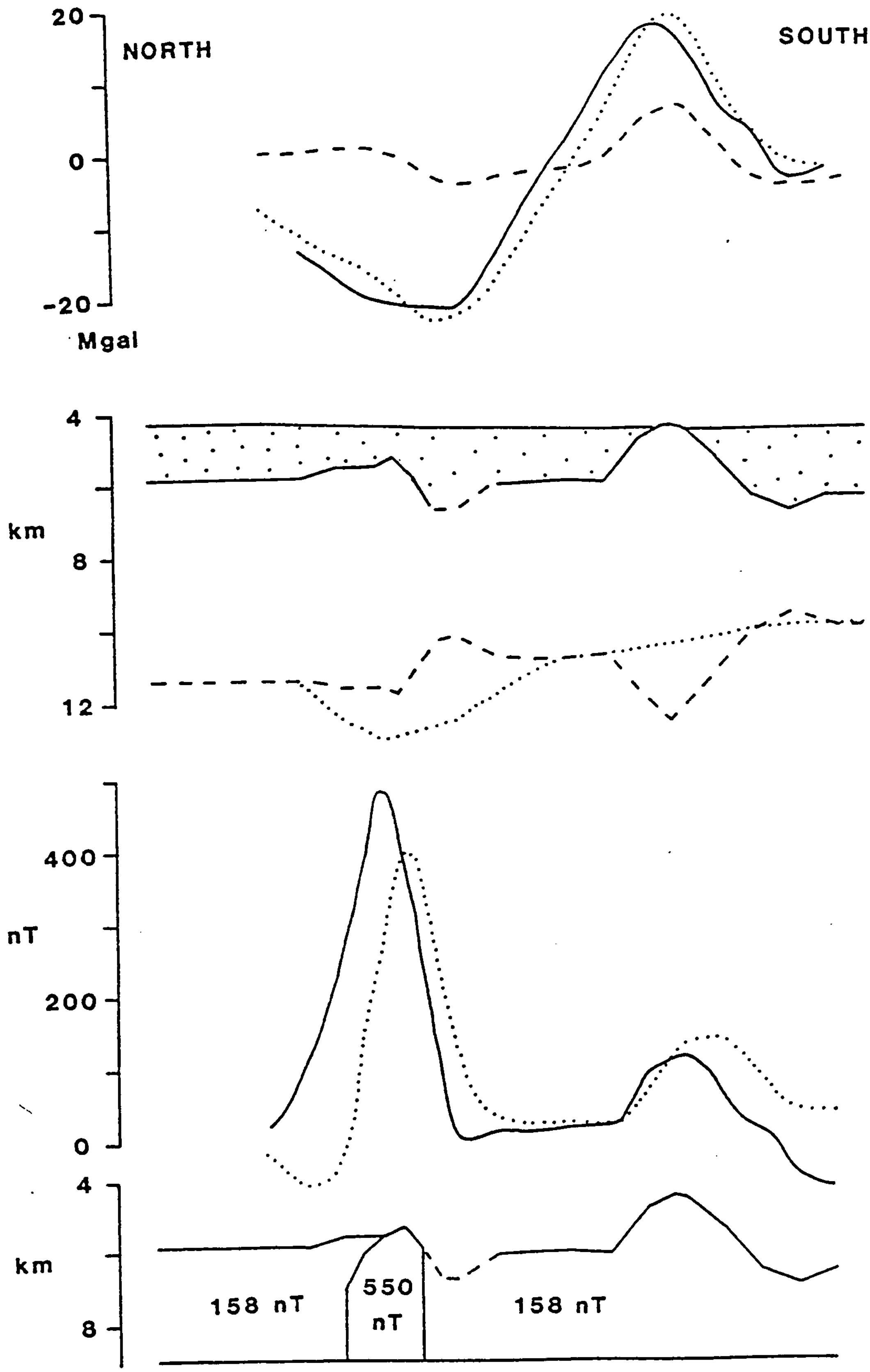


Fig. 5.9 Gravity and magnetic models across the Charlie Gibbs Fracture Zone, Challenger line 10

Gravity model : Key as Fig. 5.5

Magnetic model : solid line - observed anomaly

dotted line - calculated anomaly

It is proposed that the Charlie Gibbs Fracture Zone was initially a leaky transform fault, as would be suggested by the non-orthogonal relationship of the Charlie Gibbs Fracture Zone and anomalies 24-32. This produced intrusive activity along the transform fault, and large magnetic anomalies.

At anomaly 24 time, the transform fault became a fracture zone, and the eastern and western sides of the Atlantic were 'unlocked'. Spreading was then able to propogate northeastwards and northwestwards.

CHAPTER 6

CONCLUSIONS

This final chapter is in two parts. The first section is a general treatment of the evolution of the area through time. The second is a discussion of stretching models and the extent to which these quantitative models can be applied to the thinned continental crust of Rockall Trough and Porcupine Bank. A maturation calculation for Rockall Trough is presented in this section, as a very rough guideline to hydrocarbon prospectivity. Figure 6.1 is a present day structural map of the area.

6.1 EVOLUTION OF THE AREA

6.1.1 Palaeozoic

a) The Iapetus Suture and Caledonian trends

It is considered that the Clare Lineament and Clare Trend represent the boundary between the platforms to the north and south of the former Iapetus Ocean (Chapter 3). They form the link between the Iapetus Suture mapped in Ireland and the Dover Fault of Canada. The crust to the south of the Clare Lineament and Clare Trend has distinct affinities with the Avalon Zone of Newfoundland. This crust may also underlie the western part of southern Porcupine Seabight.

The northern part of Porcupine Bank shows typical NE-SW Caledonian trends. The crust underlying Rockall Trough probably shows similar trends which may have influenced later basin formation.

b) Devonian and Carboniferous Basin Formation

The Devonian Munster Basin in southern Ireland may have been formed by post-Caledonian orogeny wrench faulting. It is suggested that a similar Devonian basin underlies southern

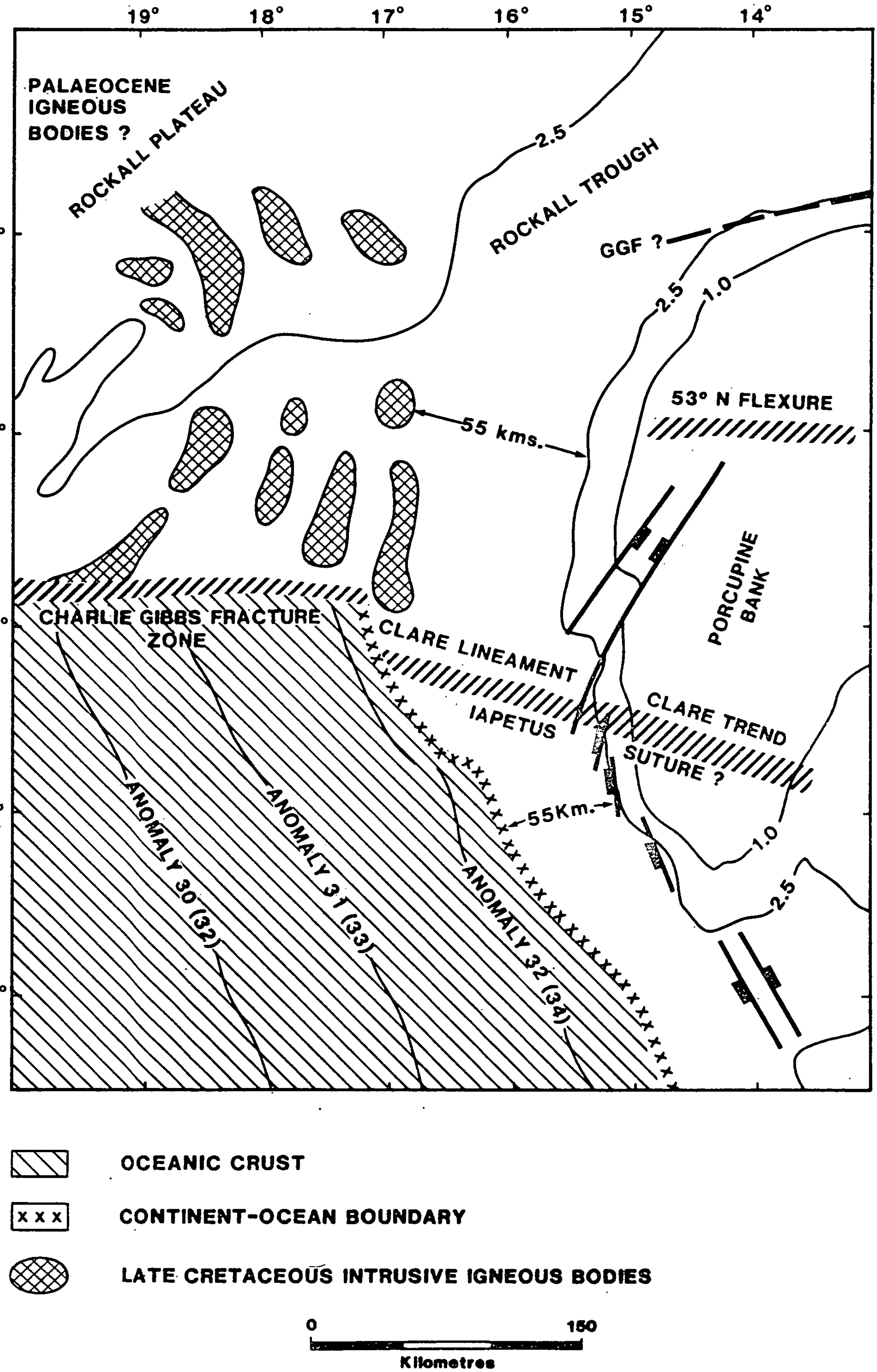


Fig. 6.1 Schematic structural map

Porcupine Seabight (P. Bennett, pers. comm.).

To the north, the Orcadian Basin formed as a result of post-orogenic wrench faulting along the Great Glen Fault. It is expected that a similar Old Red Sandstone Devonian basin may be preserved within Rockall Trough at the position where it is crossed by the Great Glen Fault. It is possible that the eastward offset of the Rockall Trough north of 54°N is a pointer to the location of the Great Glen Fault.

The Hercynian Front probably passes through southern Porcupine Seabight (Chapter 3). A Carboniferous basin may exist to the south of the Front. Gas may have migrated from the Carboniferous to reservoirs within southern Porcupine Seabight.

c) Permo-Triassic Basin Formation

During the Permian and Triassic, rifting was widespread in the area which is now the British Isles, Ireland, and their continental shelves. The North Sea, Western Approaches and Celtic Sea Basins, Bay of Biscay, Manx, Fair Isle, Hebridean Basins, and others, are sites of Permo-Triassic subsidence. It is felt likely that southern and northern Porcupine Seabight were the sites of Permo-Triassic graben formation. The crust between Porcupine Bank and Orphan Knoll may have rifted in the Triassic as a response to the rifting at the site of future formation of oceanic crust between the eastern USA and Africa.

It is probable that Rockall Trough is the site of two or more distinct Permo-Triassic basins which formed on Caledonian trends.

6.1.2 The Jurassic

The Jurassic was also a period of widespread crustal tension, as can be seen by the propogation of Atlantic spreading to the north and south. Rifting and subsidence occurred locally in the area from East Greenland and Newfoundland to the North Sea. Major Jurassic subsidence presumably occurred within Rockall Trough, between Porcupine Bank and Orphan Knoll, between Rockall Plateau, Greenland, and Canada, and within Porcupine Seabight.

Upper and Middle Jurassic shales are the source of hydrocarbons in the North Sea oil and condensate fields, in the Clair field (west of Shetland) and Porcupine Seabight.

Possible hydrocarbon sources in the Rockall rift area are:

a) Middle Jurassic shale cf the oil shales of the Hebridean Basin and

b) Upper Jurassic 'hot shales'. If, as is suggested by their relative proximity to Atlantic seafloor spreading, the Rockall-Porcupine Bank-Orphan Knoll rift system formed a large connected basin at this period, circulation would have been too unrestricted for the development of anoxic conditions. 'Hot shales' would not be found in this area.

6.1.3 Early Seafloor Spreading

a) Pre-anomaly 32 (34) crustal stretching

Figure 6.2 shows the extent of pre-anomaly 32 (34) oceanic crust and the area where appreciable crustal stretching is thought to have taken place. The Clare Lineament, the NE-SW graben of northern Porcupine Bank and the en echelon bounding faults of southern Porcupine Bank probably formed at this time.

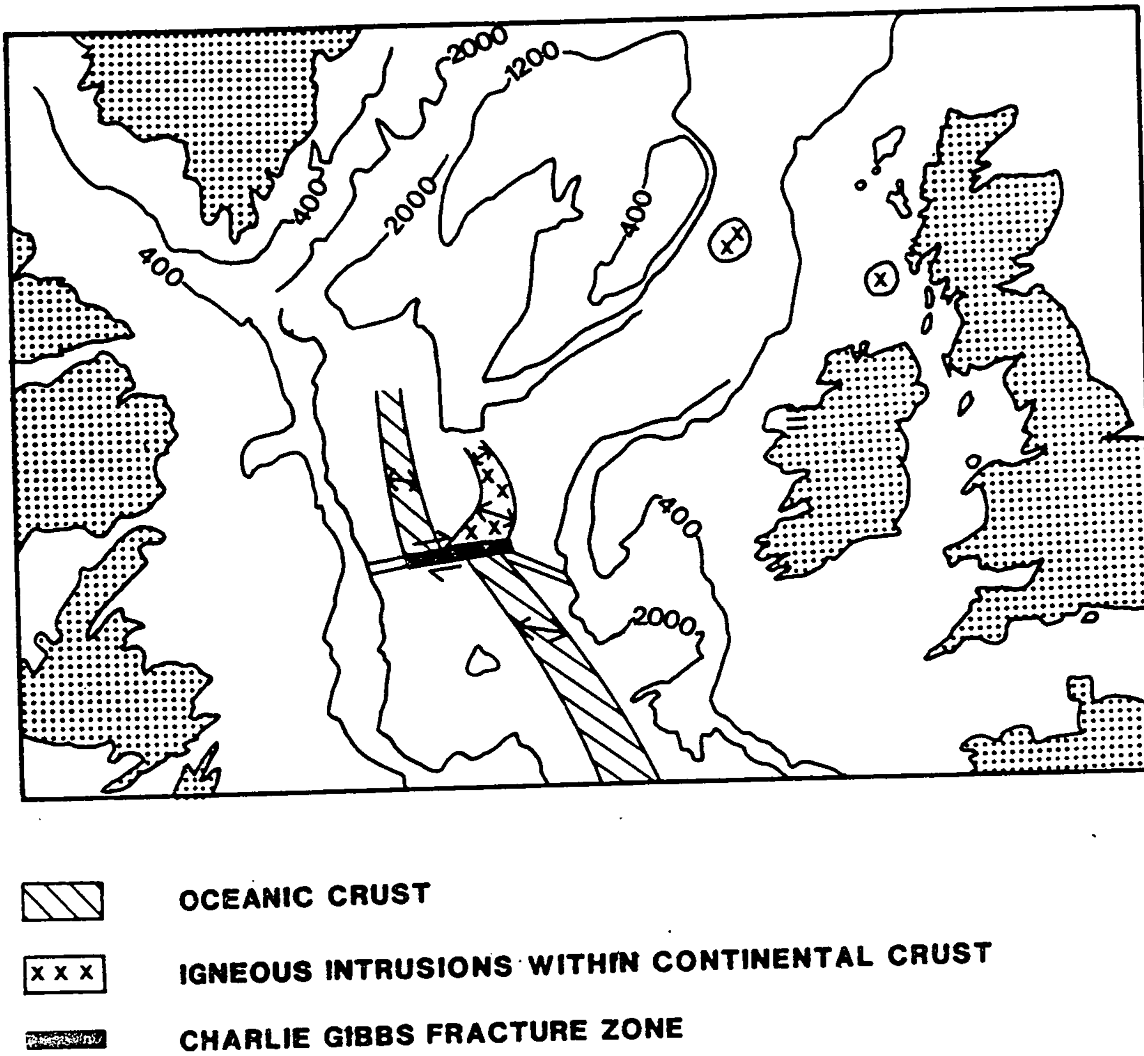


Fig. 6.2 North Atlantic reconstruction at pre-anomaly 32 (34) time. The reconstruction, which is post-stretching, is made by matching the eastern and western ends of the Charlie Gibbs Fracture Zone.

b) Anomaly 32 (34)

The lithosphere thinned sufficiently to allow asthenospheric breakthrough (Le Pichon et al, 1982) at a slightly older date than anomaly 32 (34) time, i.e. in the Campanian or Santonian. At this stage, continental crustal stretching (except by intrusion) ceased and thermal subsidence became effective.

The future Charlie Gibbs Fracture Zone began to form as a transform fault, offsetting two spreading centres. The fracture zone was initiated at the position where the Iapetus Suture intersected the propogating Atlantic crack (Fig. 6.3).

As spreading was not orthogonal to the transform fault, limited 'spreading' occurred across the transform fault, in the form of linear igneous intrusions. Further stresses at the eastern end of the transform fault resulted in up to 70 km (width) of igneous bodies being intruded within the thin crust of southern Rockall Trough. Igneous bodies were also intruded in northern Rockall Trough i.e. Anton Dohrn Knoll, ?Rosemary Bank. Sills within the sedimentary section of the Faeroe-Shetland Channel (Ridd. in press) mat be evidence of further igneous activity at this time.

c) Anomaly 24

At anomaly 24 time (Fig. 6.4) the Charlie Gibbs Fracture Zone became a Fracture Zone solely within oceanic crust. This may have allowed the 'unlocking' of the European and American plates, and resulted in spreading in the Labrador Sea and between East Greenland and Rockall Plateau. The Palaeocene was a time of intensive igneous activity in northern Rockall Trough, Rockall Plateau, west of Scotland and possibly in southern Rockall Trough.

The post-anomaly 24 history of Rockall Trough was dominated by subsidence and the build up of Feni Ridge by contour currents in the ?Eocene.

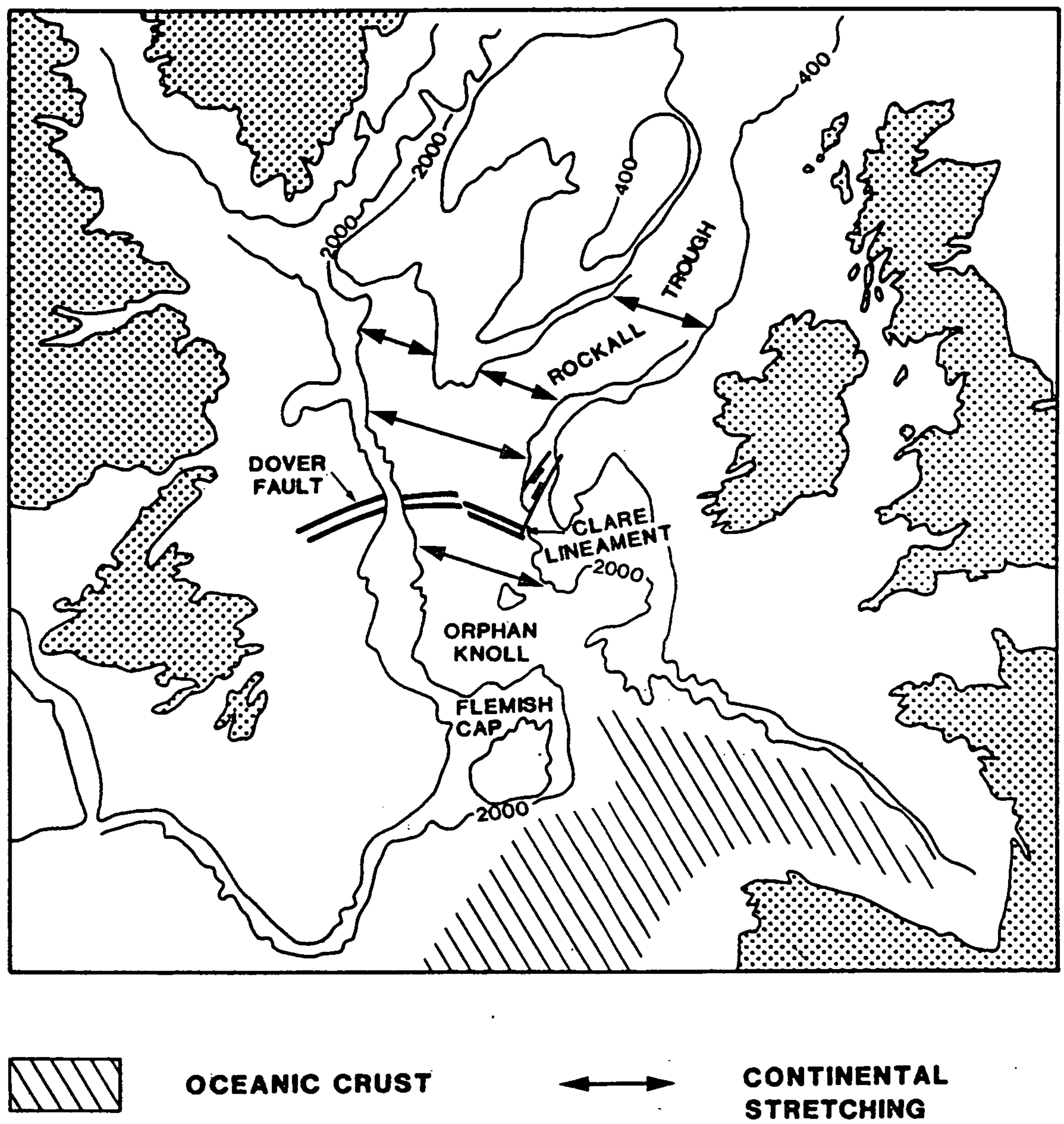
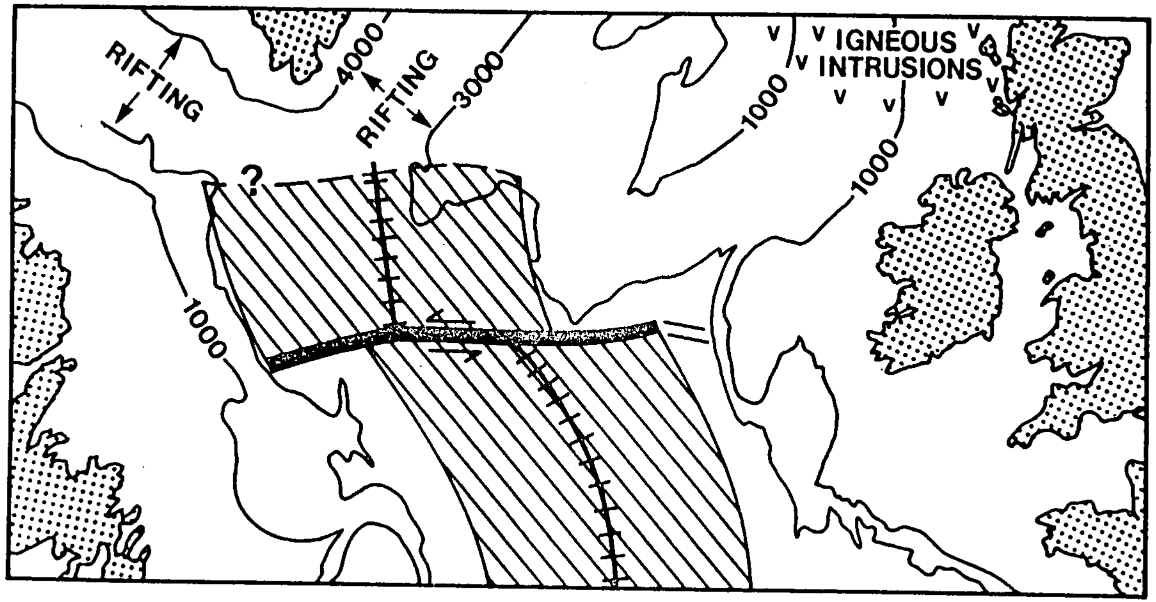


Fig. 6.3 North Atlantic reconstruction at anomaly 32 (34) time



OCEANIC CRUST ANOM. 24-32(34)

SPREADING RIDGE

CHARLIE GIBBS FRACTURE ZONE

Fig. 6.4 North Atlantic reconstruction at pre-anomaly 24 time

6.2 STRETCHING AND SUBSIDENCE

6.2.1 β Values

The β value is the measure of the extent of thinning of continental crust (McKenzie, 1978).

a) Thinned continental crust of Porcupine Abyssal Plain

The crustal model of Shackleton line 4 (Fig. 3.22) gives an average crustal thickness of approximately 6.5 km for thinned continental crust, and 26 km for the crust of southern Porcupine Bank. This give a β = 4, or 400% <u>cumulative</u> stretching. The thinning is probably a result of several distinct rifting episodes.

As the distance from Porcupine Bank to the continent-ocean boundary is at present 55 km, the pre-stretching crustal width of this crust was about 14 km.

b) Southern Rockall Trough

This area is complicated as a result of extension caused by igneous intrusion. The width of intrusions varies: it is 70 km along Shackleton line 14, to the south; 25 km along Challenger line 1; and 55 km along Challenger line 2, to the north.

The gravity high, situated where igneous intrusions are at their minimum width, represents extension by extreme crustal thinning by asthenospheric breakthrough within the lower crust, as opposed to extension by igneous intrusive activity.

The present width of southern Rockall Trough is ca 300 km. This gives a pre-igneous intrusion (i.e. pre-anomaly 32 (34)) width of ca 230 km. The crustal model of Challenger line 1 (Fig. 4.6) gives a non-igneous crustal thickness of ca 7km, and therefore a β value of 4.

The pre-(cumulative) stretching width of Rockall Trough was approximately 60 km.

6.2.2 Discussion of Stretching and Subsidence Models

Two quantitative models have been developed to predict features of extensional basins.

a) McKenzie's (1978) stretching model. The lithosphere is thinned by a factor β, producing initial thermal subsidence followed by exponential thermal subsidence.

b) Royden et al's (1980) model invokes crustal stretching purely by igneous intrusion. Subsidence is caused by thermal decay and by crustal density increase.

For both models, heat flow and subsidence through time can be predicted through time if β is known. The subsidence curve through time is very similar for both mechanisms.

Many basins stretch by a combination of lithospheric thinning and igneous intrusion. Examples are the Faeroe-Shetland Basin, southern Rockall Trough, and the North Sea.

The main use of such models is in predicting whether a basin is of commercial interest at an early stage. If the age of stretching and the crustal thinning mechanism and factor are known, the thermal history of the basin is also known and the maturity of sediments at various horizons and localities can be calculated.

These models cannot be used for Rockall Trough and for the thinned continental crust of Porcupine Abyssal Plain. The cumulative crustal stretching factor only is known, not the amount of crustal stretching at each rifting episode.

6.2.3 Maturation Calculation for Rockall Trough

The results of this calculation are necessarily very approximate. The calculation was made using Lopatin's method, as described by Waples (1980). For the formation of interest a burial depth vs age profile and an estimate of the geothermal gradient through time are required. A cumulative 'Time Temperature Index' is arrived at by calculating the length of time the formation spends in each 10°C temperature interval and multiplying by the appropriate temperature factor.

The correlation of the 'Time Temperature Index' with several important stages of oil generation and presentation is given below:

Stage	TTI
Onset of oil generation	15
Peak oil generation	75
End of oil generation	160
Upper limit for occurrence of wet gas	1,500
Last known occurrence of dry gas	65,000

The area chosen for the calculation was a graben in the eastern quiet zone of southern Rockall Trough. The exact position is at 0845 z/096 on Challenger line 1 (Fig. 4.5). Sediment thickness here is about 3 seconds, TWT. The sediment column is over 4 seconds thick under the Feni Ridge, and deep sediments here will be more mature than those on the eastern side of Rockall Trough. However, the Shackleton horizon has only been picked in eastern Rockall Trough, and the age of the Shackleton horizon is roughly known. A more accurate calculation is therefore possible. A schematic geological interpretation of Challenger line 1 is given in Fig. 6.5

The maturity of a horizon at 7.3 seconds was calculated. It is thought that this may be the 'J', or Base Cretaceous horizon, and be close to the position of potential source beds.

The following interval velocities were assumed (with possible large errors):

Seabed -	Challenger	(38 my)	4,000 ft/s
Challenger -	Shackleton	(38 - 55 my)	6,500 ft/s
Shackleton -	?Top Jurassic	(55 - 144 my)	8,000 ft/s

This results in the burial depth vs age profile shown in Fig. 6.6.

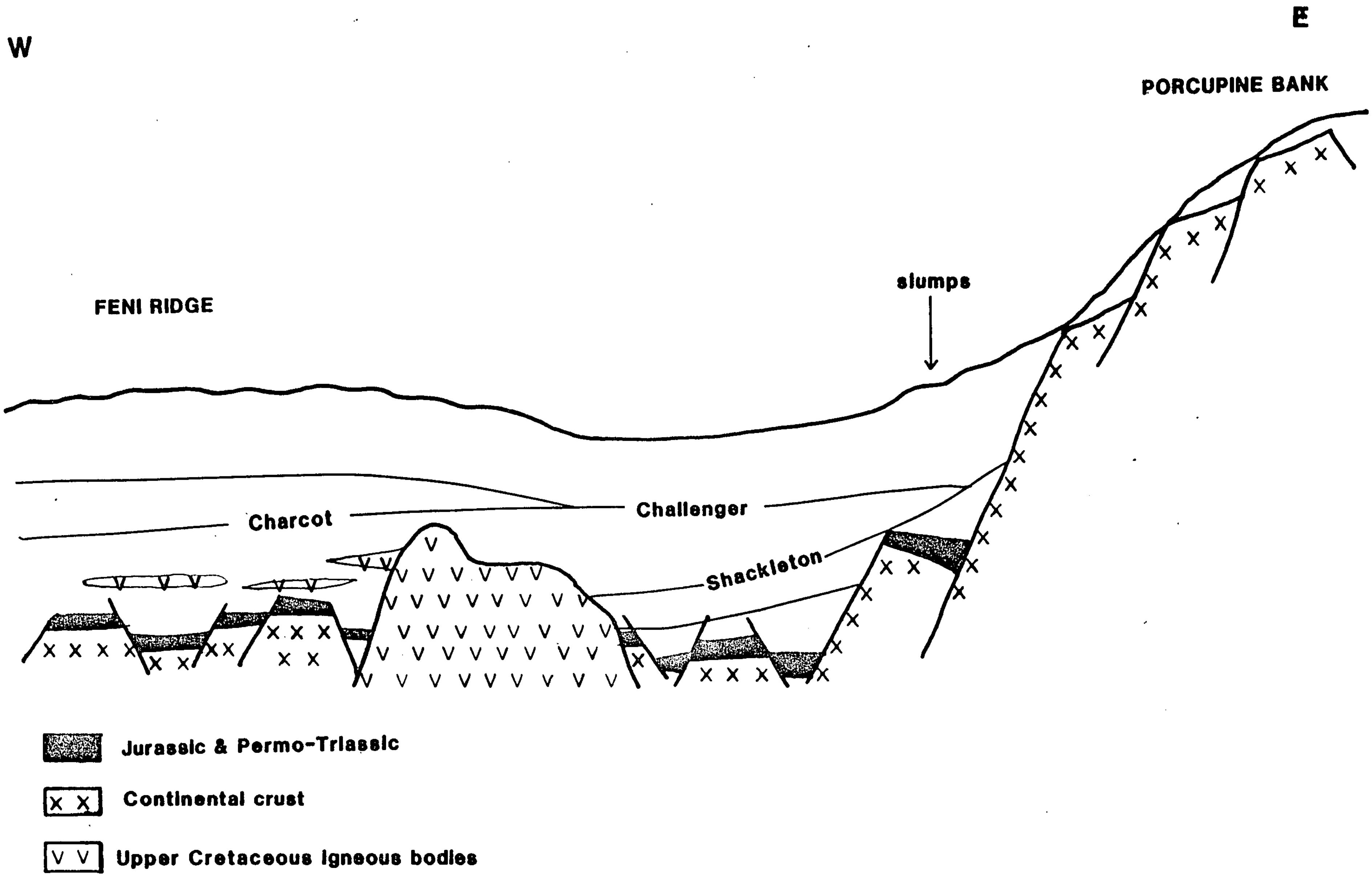


Fig. 6.5 Geological interpretation of Challenger line 1.

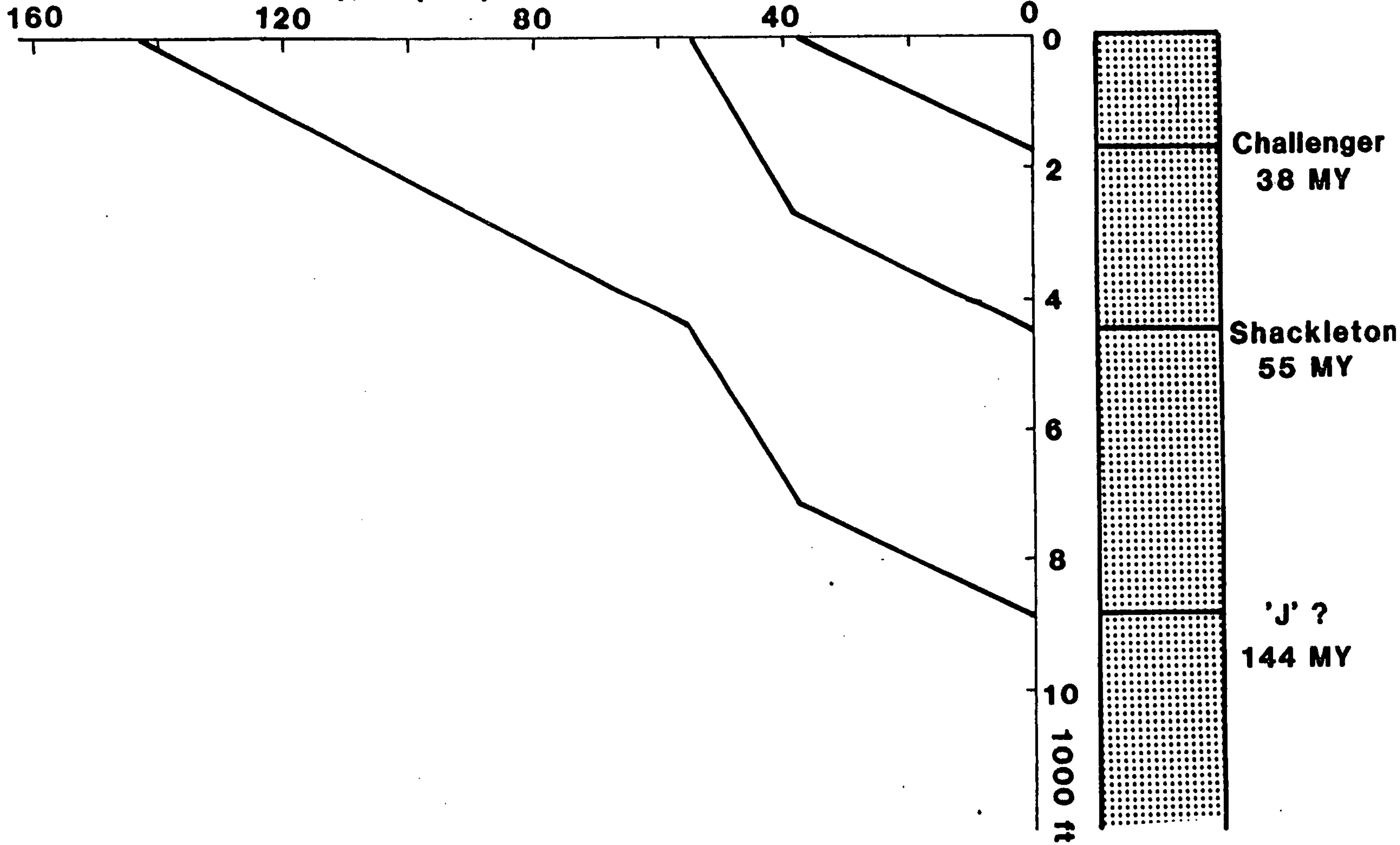


Fig. 6.6 Burial depth versus age profile for eastern Rockall Trough, at 0845 z/096, Challenger line 1

The geothermal gradient chosen is shown in Fig. 6.7. The gradient was assumed to decrease following a Jurassic stretching event until the Late Cretaceous and Palaeocene, when intrusive activity took place. Temperature gradients chosen are highly hypothetical. Temperature is crucial and the largest errors are introduced by assumptions made at this stage.

The resulting cumulative TTI for the ?top Jurassic is 19. This is within the oil generating window. Oil generation would have started 14 my ago.

Therefore if the assumptions made about temperature gradient are approximately correct, it is expected that deeply-buried sediments within eastern Rockall Trough are marginally mature, and that those within western Rockall Trough are well within the oil generating window.

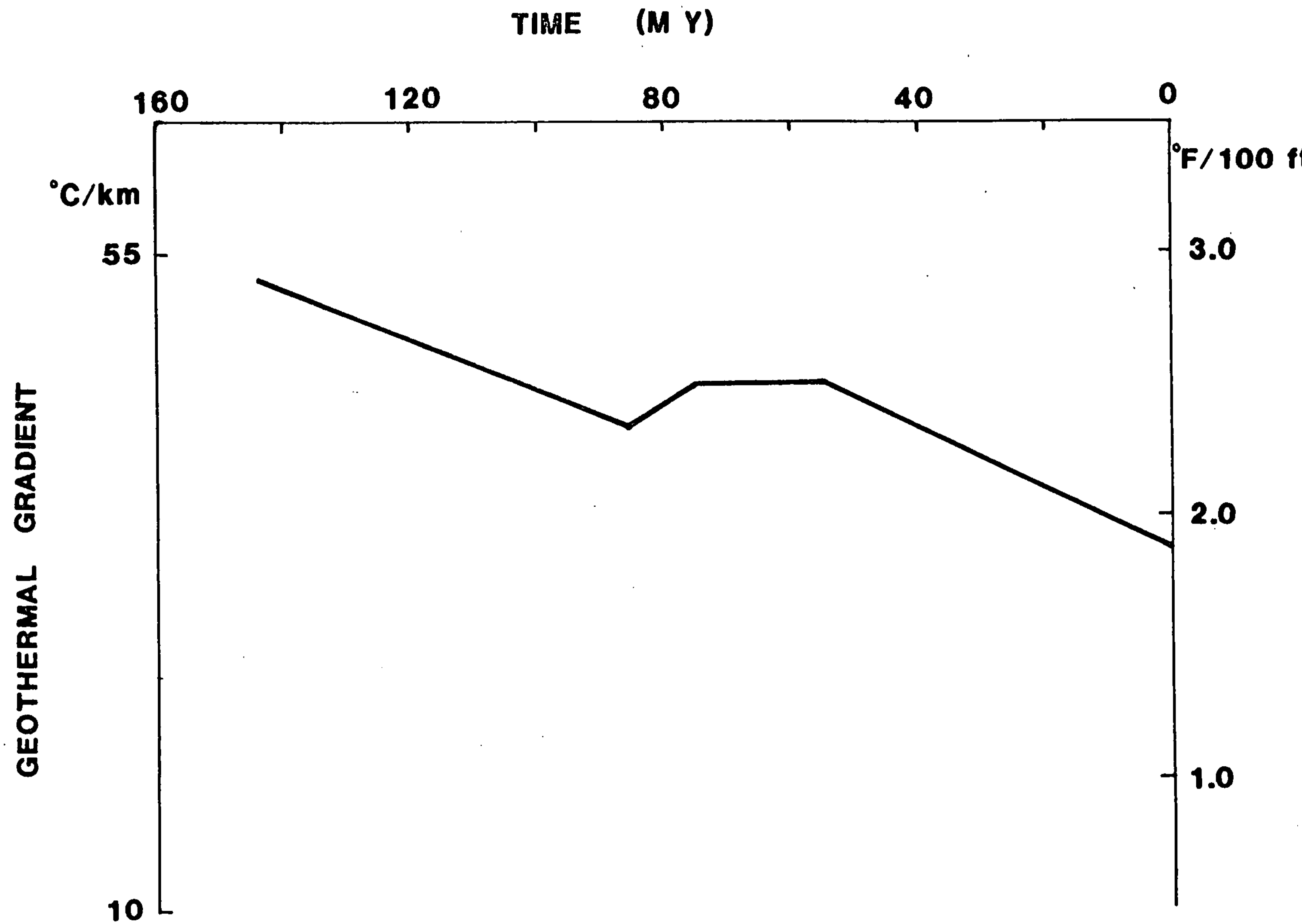


Fig. 6.7 Hypothetical geothermal gradient through time, southern Rockall Trough

APPENDIX A

DISTINGUISHING CONTINENTAL FROM OCEANIC CRUST AT PASSIVE CONTINENTAL MARGINS

When symmetrical magnetic reversals in the sea-floor were described (Vine and Matthews, 1963) and the theory of ocean-floor formation by generation of new crust at oceanic ridges developed, the arguments of the 'fixists' in favour of oceanisation seemed to be dispelled. However the rigid plate theory has had to suffer some modification; it has been known to geologists for a long time that the continental lithosphere does not act as a rigid plate during orogeny, but deforms irreversibly, and recent research has demonstrated that the continental lithosphere deforms at some passive margins by thinning, sometimes to an extreme degree (Montadert et al, 1977). Models of ancient passive continental margins also incorporate deep and block faulted continental crust bordering oceanic crust.

To some authors it is clear that the discovery of continental crust at oceanic depths, and of oceanic crustal thickness, has decreased the ideological distance between 'fixists' and 'mobilists', and also that it may be difficult or impossible to distinguish geophysically thinned continental and ocean crusts. The following is an extract from Van der Linden's 'How much continent under the ocean?' (1977):

> "Most of the data obtained thus far are from indirect geophysical observations and fundamental geological proof for either alternative is conspicuously absent [the alternatives being the existence of broad zones of attenuated continental crust at trailing margins, and a clean-cut transition between continental crust of near-'normal' thickness and ocean crust]. As a matter of fact, it is not very clear how to conduct an experiment apart from deep drilling, that will establish valid criteria for the distinction of either model. As argued, the magnetic anomaly pattern can be interpreted either way; seismic reflections from deep basement under thick sediment-cover do not distinguish between layer 2 or block faulted, magmatically intruded, and altered continental crust; seismic velocities can be interpreted as representative of either continental or oceanic lithologies and gravity data never provide a unique solution."

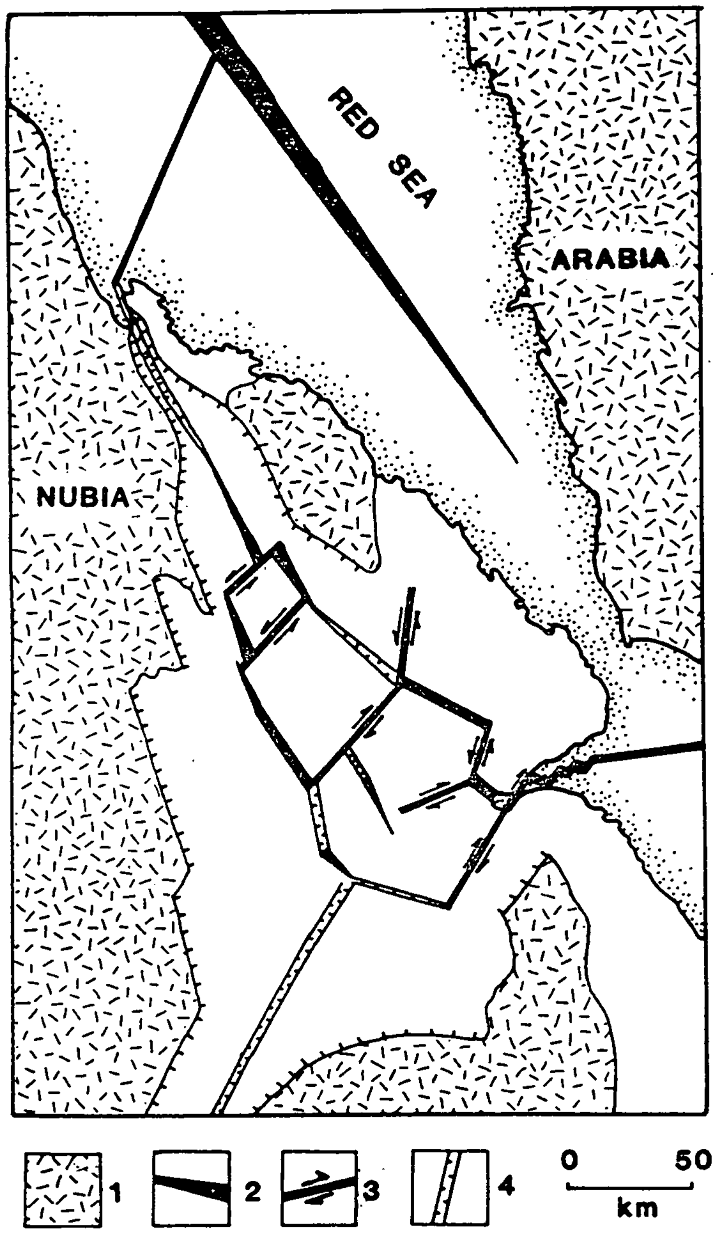


Fig. A1 Schematic structural map of Afar, after Barberi and Varet, 1977.

1. Outcropping continental basement
2. Axes of spreading (axial ranges in Afar)
3. Relative motion along transform faults
4. Manifestations of extensional tectonics (large grabens without axial volcanic ranges)

The purpose of this section is to discuss the extent to which this is true, and whether there are in fact any geophysical criteria, considered individually or collectively, which can distinguish continental from oceanic crust, and/or enable the continent-ocean boundary to be mapped.[1]

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[1]Leaving aside temporarily the problem of the geophysical detection of the continent-ocean boundary, it is interesting to consider whether there is any general and clear-cut geological distinction between oceanic and continental crusts, or whether the two form opposite ends of a continuum within which it is impossible to say exactly where 'true' oceanic crust is generated. (The analogy could be used of the contiuum between basalts of tholeiitic 'oceanic' composition, and basalts of alkali 'continental rift' composition).

Afar is an anomolous area (Schilling 1973) in which the gradations between crustal types are displayed, and where disagreements arise over crustal definitions even though the upper crust can be directly sampled. For most continental margins the distinction would be considerably easier due to less intense volcanic activity. Afar could be described as a complex of microplates, with spreading ridges, transform faults, and leaky fracture zones (Barberi and Varet, 1977), (Fig. A1). Most of the floor of Afar is composed of the Stratoid Series flood basalts that were emplaced in the last 4 m.y. and is attenuated and block tilted. All refraction experiments show that this is underlain by low velocity mantle, but some authors interpret the data for the crust as oceanic (Berkheimer et al, 1975; Makris et al, 1975) overlain by thick basalts. In the last million years, crustal extension became localised at the Axial Ranges, which are described as spreading ridges by some authors. However, according to Barberi and Varet (1977) there is no or very little crustal separation under the Axial Ranges, and their volcanic evolution is characterised by two stages which are not seen on mid-ocean ridges. It may be that the types of igneous activity associated with the Axial Ranges are typical of the initial stages of ocean formation, which are those areas usually now buried by sediments at passive margins.

It is a matter of opinion as to how much crustal separation constitute an ocean, and whether the Axial Ranges can be defined as spreading ridges, but two criteria of oceanic crustal formation could be

a) crustal extension by dyke intrusion must be confined to one plane at any one time (i.e. disseminated dyke intrusion within continental crust is simply crustal extension, and not formation of new crust, however pervasive)

and

b) new dykes must in general intrude the most recent dykes, to give spreading symmetry.

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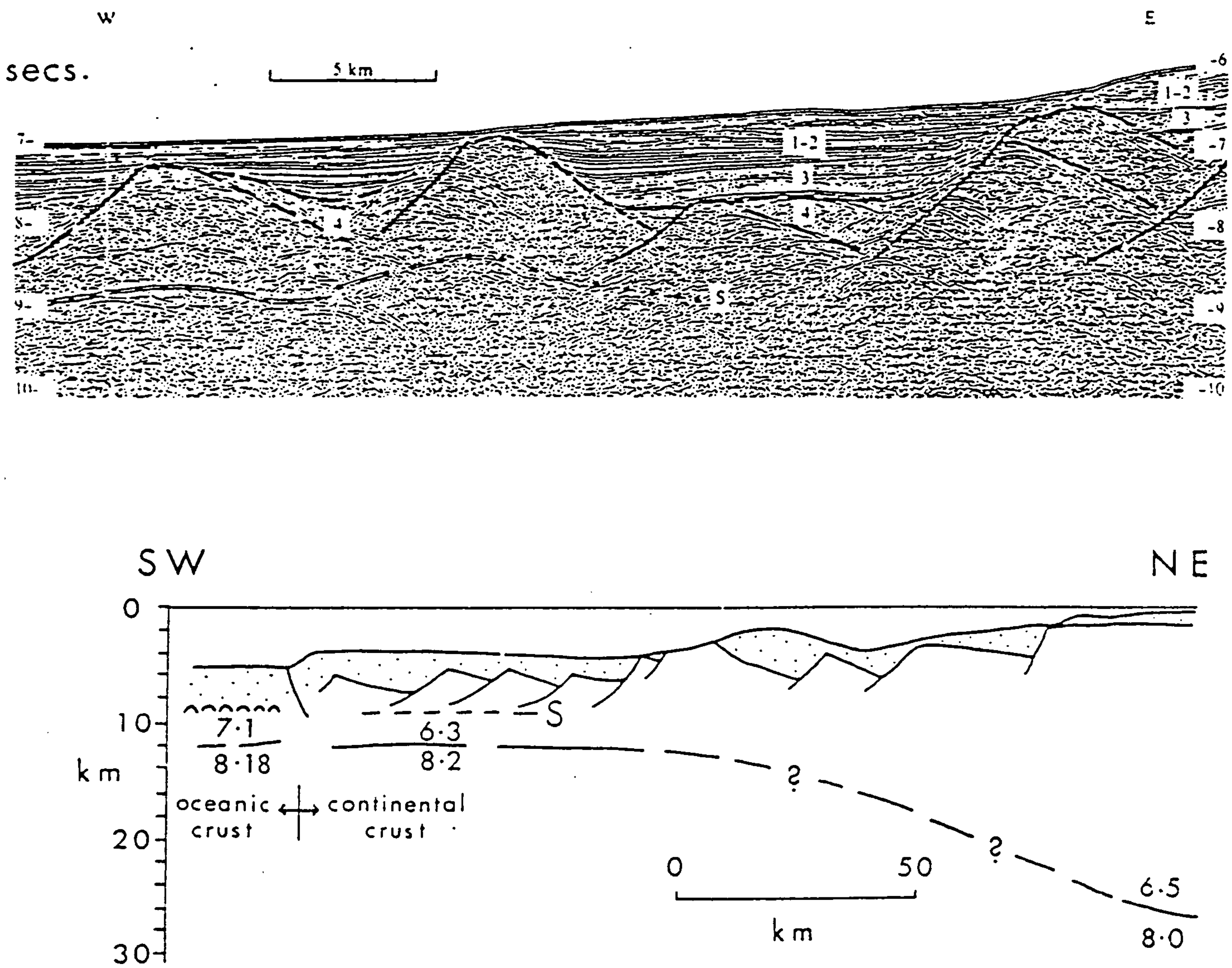

Fig. A2 Attenuated continental crust in oceanic water depths, Biscay (de Charpal et al, 1978)

Top: a multi-channel seismic reflection profile west of Galicia Bank showing tilted fault blocks and listric faults. The horizontal reflector S may mark the boundary between brittle upper crust and ductile lower crust.

Below: a schematic crustal section through the North Biscay margin

It has often been assumed that one criterion that can be used to define the presence of oceanic crust is crustal thickness, and that a large and discrete change in thickness of the crust occurs under the continental slope. For instance, Bullard et al's (1965) reconstruction of the Atlantic used the 500 fathom bathymetric contour as a continental edge. Keen and Barrett (1981), working on the East Canadian margin, consider that crustal thickness may still be an important indicator; "thicknesses of about 17 km appear to be near the minimum measured beneath subsided continental regions, although more observations are highly desirable." As a predictive tool, this seems untenable in the face of Montadert et al's (1977) conclusive recognition of continental crust 8 km or less thick in the Bay of Biscay (Fig. A2). Therefore this section will first discuss individually some other methods which may detect the continent-ocean boundary geophysically, although obviously in any survey as much information as possible will be used in conjunction. There are hazards in merely listing the geophysical features of previously mapped continent-ocean boundaries as this assumes that the identification was correct, and the argument becomes a circular one. A later sub-section will briefly review different interpretations of crustal type at various controverial passive continental margins.

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[1] continued

It is suggested here that the term 'transitional crust' is of little use as it usually used to define crust of a thickness between that of 'typical' oceanic crust and 'typical' continental crust, which if more closely investigated would be found to be thick oceanic crust abutting thinned, faulted, and possibly intruded continental crust. The term 'transitional crust', if used at all, should be a temporary term, and discarded as soon as possible.

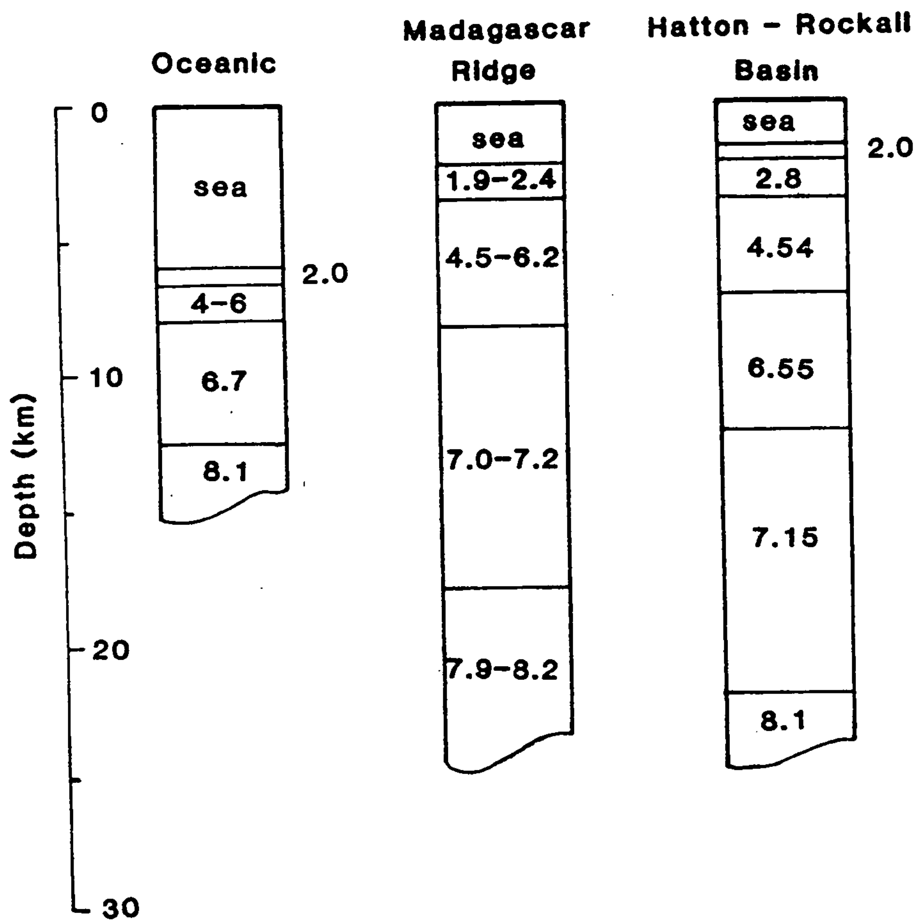

Fig. A3 Crustal structure models from simple first arrival plane-layer analyses. Sinha et al, 1981.

a) SEISMIC REFRACTION

Sinha et al (1981) summarise what they see as the differences in velocity structure between continental and oceanic crust; "Refraction results .....indicate that where thinning of continental crust has occurred, the ratios of the thicknesses of the two major layers do not change significantly. In particular, the 6.8 - 7.0 km/s lower crustal layer continues to make up at most only half of the crustal thickness." On this basis they concluded that the Madagascar Ridge was of oceanic origin (Fig. A3), as the upper crustal layers (Vp = 6.5 km/s or less) make up less than one-third of the total crustal volume.

Further they suggest that the deep crustal layer (Vp ≐ 7.5 km/s) found under Madagascar Ridge from synthetic seismogram analysis is another indicator of oceanic crust, being a layer never seen under continents, but observed as a second arrival over much of the Pacific (Sutton et al, 1971) and inferred to be present under the Mid-Atlantic Ridge (Fowler, 1976).

There are areas beneath passive margins for which the velocity structure, as obtained from seismic refraction results, seems neither typically oceanic or continental, e.g. the Voring Plateau (Talwani and Eldholm, 1976) and the Sao Paulo Plateau (Leyden, 1976), some of which will be discussed further later. Furthermore, the refraction data from a portion of the continental Orphan Basin (Keen and Barrett, 1981) suggests that in places the lower crustal layer (Vp = 7.4 km/s) is thicker than the upper crustal layer (Vp = 5.9 - 6.6 km/s). It is possible that these ambiguities would disappear if synthetic seismogram analysis techniques were used, as according to Sinha et al these techniques tend to show that the average crustal velocities are lower, and the high velocity (7 km/s) layer in the lower crust is thinner, than is indicated by first arrival analyses.

Van der Linden (1977) states that anomalously low velocity mantle (7.2 - 7.4 km/s) beneath passive margins may underlie attenuated continental crust. If this really does underlie continental crust (Van der Linden reports this anomalous mantle from parts of the Labrador margin, and the Voring Plateau) then regardless of whether it is crustal (which is possible) or anomalous mantle material, it could be confused with the 7.5 km/s layer which according to Sinha et al is never found under the continents. The lower crust under the Orphan Basin (see above) also has a velocity of 7.4 km/s. The process involved in generating low-velocity mantle under rifts (Van der Linden 1978) is one of intrusion of the lithosphere by the asthenosphere, and a resulting 'crust-mantle mix'. Many people would call this intruded lower crust, or a similar term, but the definition is unimportant in this context compared with the probable presence of a high-velocity layer under some continental margins. It is possible that the lack of a high velocity layer, after sophisticated analysis has been carried out, is indicative of continental crust, but that its presence would be inconclusive. On the other hand it is possible that sophisticated analysis could resolve the magmatic intrusions to some extent and show that the high velocity layer observed under continental crust at some passive margins is a product of higher-velocity intrusions and lower velocity continental crust. Van der Linden does suggest that a wide range of lower-crust or mantle velocities observed in the same area would point towards the existence of a 'mixed' layer, i.e. one which underlies continental crust. (Refer also to the later sub-sections on the Red Sea and South Atlantic).

b) SEISMIC REFLECTION

As has been mentioned already, under ideal circumstances (i.e. thin and acoustically non-opaque sediment cover over 'normal' continental and oceanic crusts), multichannel reflection seismic surveys can unambiguously distinguish continental from oceanic crusts (Montadert et al 1977). Continental crust is recognised in Biscay by its internal sedimentary layering within blocks tilted by listric faulting. A reflector called the 'S reflector' is sometimes present, which may seismically represent the boundary between the upper, brittle, crust and the lower, ductile, crust (Fig. A2).

Although extensive faulting by tilting along listric faults may not be typical of passive margins (and certainly is not typical of the sheared portions of these margins); tilted blocks have been recognised in Afar (Black, 1976), off the British Honduras (Dillon and Veder, 1973), in the Bay of Biscay (Montadert et al, 1979), off S.E. Baja California (Normark and Curray, 1968), and off Nova Scotia (Given, 1977), etc.

When an area of faulted continental crust is covered with thick sediments and/or relatively unsophisticated reflection seismic techniques are used, the diffraction produced by 'highs' of continental basement are difficult or impossible to distinguish from the diffraction patterns produced by uneven oceanic basement; alternatively the basement may be too deep to be visible. Also, as Sheridan (1979) pointed out, layers of salt and limestone which produce a high acoustic impedance 'barrier' tend to be especially developed at the continental slope and continental edge, where interest may centre. Salt in particular obscures large areas of the mid and South Atlantic, and the Mediterranean and Red Seas.

Another problem is that continental and oceanic crusts near the continent-ocean boundary may be 'anomalous'. Continental crust near a hot-spot will usually be heavily intruded and basalt-covered, e.g. the Afar floor (Black, 1976) and the Faeroes (Bott, 1978). Unusal initial spreading conditions (i.e. fast or slow spreading) probably produce oceanic crust with an uncharacteristic seismic signature. Eldholm et al 1979 identified an area of smooth basement (with sub-basement reflectors) off the Voring Plateau as oceanic crust. It is suggested that the smooth surface may be a result of an initial very high rate of basalt production, with flows and pyroclastic material interlayered with sediments, overlying slightly older oceanic basement. This smooth basement is common adjacent to continent-ocean boundaries in the northernmost Atlantic region.

c) MAGNETICS

The magnetic and gravity fields over passive margins are often used as the only, or a major, means of finding a continent-ocean boundary; the collection of potential field data is relatively cheap and straightforward. Disagreements over the significance of particular anomalies, however, are common (refer also to the later sub-sections on particular margins). One reason is that the variables used in calculating magnetic models can take one of a very wide range of values. For instance, if the modelling of the magnetic field of oceanic crust is considered, the measured susceptibilities from basaltic oceanic rocks vary by a factor of 20 (Vacquier, 1972); the Konigsberger ratio Q for oceanic basalts can be in the range of 1-100 (Vacquier op. cit.), and the extra variable introduced by the possibility of reversed magnetisation results in a large number of sets of values to choose from. This is without taking into account other rocktypes with different magnetic properties which are present within oceanic crust; Cochran (1973) and Rabinowitz et al (1976) modelled fracture zones as areas of low magnetisation, as although the ultrabasic rocks present in fracture zones have a high susceptibility, they have a very low Konigsberger ratio (Cochran, op. cit.).

If no other information were taken into account, it would be possible to produce a large number of models which satisfied the observed anomaly or set of anomalies.

Magnetic Quiet Zones

Most controversy centre around the origin of the magnetic quiet zones; 'quiet zones' being a term which covers a range from areas with a completely flat magnetic field, to areas with anomalies of a few hundred gamma. Bott (1976) and Cochran (1981) point out that quiet zones may have several different origins. Cande et al (1978) regard the Pacific Jurassic quiet zone, which shows a subdued anomaly pattern, as the result of a series of reversals in a period of low magnetic field intensity in the Jurassic. A different origin has been suggested for the origin of the quiet zone in the

Fig A4

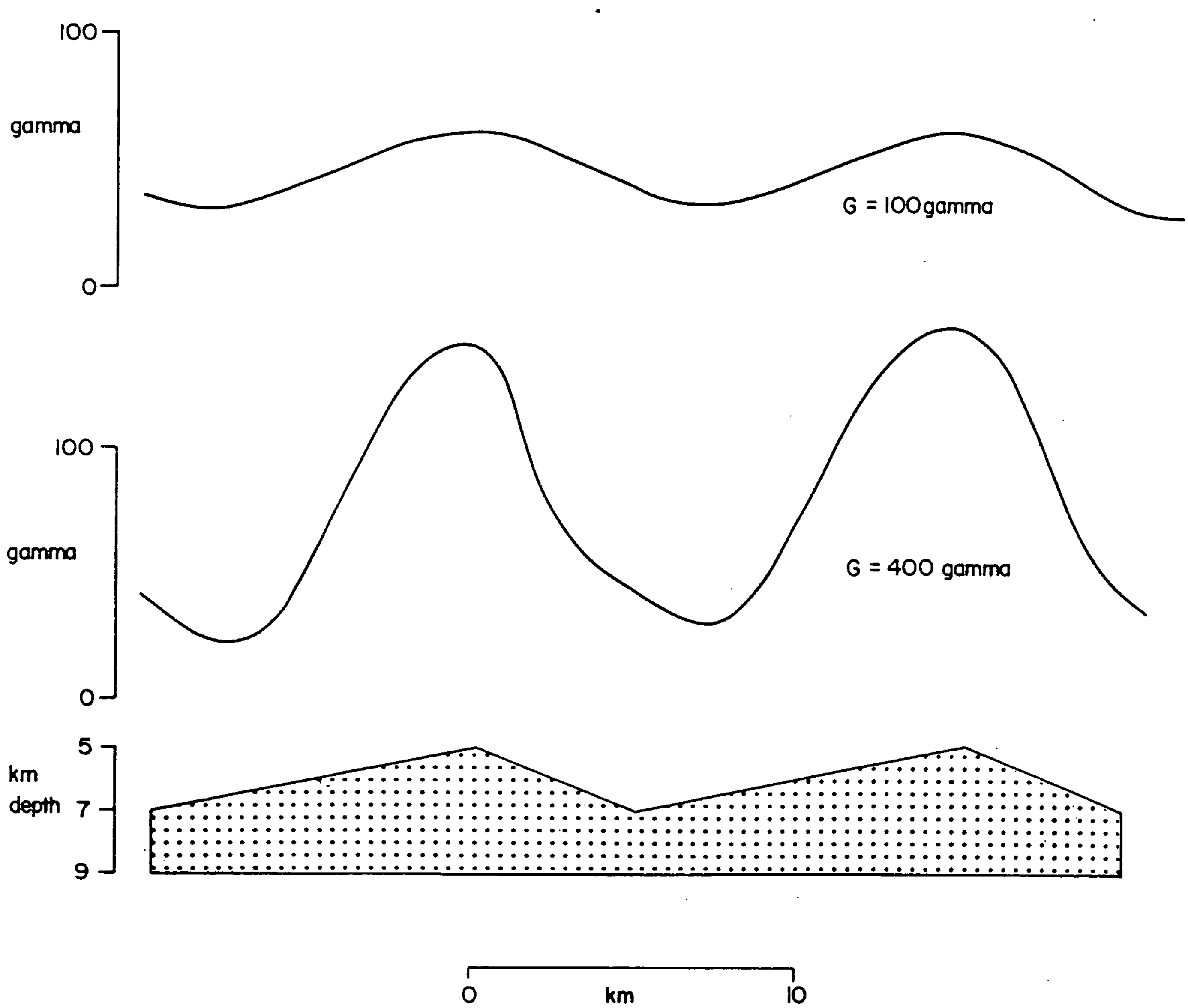

Fig. A4 Magnetic anomalies produced by deep crust tilted by listric faulting

Top G = 100 gamma (continental crust)

Below G = 400 gamma (oceanic crust or highly magnetic continental crust).

Gulf of California by Larson et al (1972). The mechanism is that of an extremely high sedimentation rate which has continually buried the ridge axis with the result that dikes fail to reach the surface and form sills, and also cool very slowly, thus failing to record magnetic reversals. Reversals are not however necessary for the production of an anomaly pattern over oceanic crust, as topography in oceanic basement (which could be regarded as a contrast of susceptibility between basement and sediment) will have the same effect. Twigt et al (1979) estimate that off the Azores ocean floor topography is responsible for magnetic anomalies of up to 400 gamma.

Van der Linden (1977) and Cochran (1981) emphasise the importance of igneous intrusions along sub-parallel fractures in continenal crust, in producing anomalies at passive margins. There is a spectrum of degree of igneous activity at passive continental margins, varying from little or none (North Biscay), through extensive (Sao Paolo Plateau), to pervasive (Afar, Faeroes Block). A set of magnetic anomaly profiles were calculated for idealised models of continental crust at the outer margin, using approximate basement topography taken from Le Pichon and Sibuet (1981).

The magnitudes of the anomalies could be greatly varied by different assumptions about layer thickness and magnetisations, and depth of burial, and a different topographic model. The models were calculated for a latitude of 52°N, N-S trending fault blocks, and normal magnetisations. Fig. A4 shows that the magnitude of anomaly over unintruded continental crust could be 50 gammas or more. Finely disseminated intrusions would produce an anomaly pattern of similar shape but greater magnitude. The lower profile is that calculated for a model of the same topography, but for oceanic crust instead of continental crust. Fig. A5 shows the modelled anomalies over continental crust with small vertical igneous intrusions. Interestingly, in the top profile magnetisation the values and topography chosen for the model result in <u>smaller</u> anomalies than for unintruded crust. When the intrusions alone are considered (lower profile), intrusions 2 km wide, 2 km thick, and at depths of 7 km produce anomalies of 50 gamma magnitude when

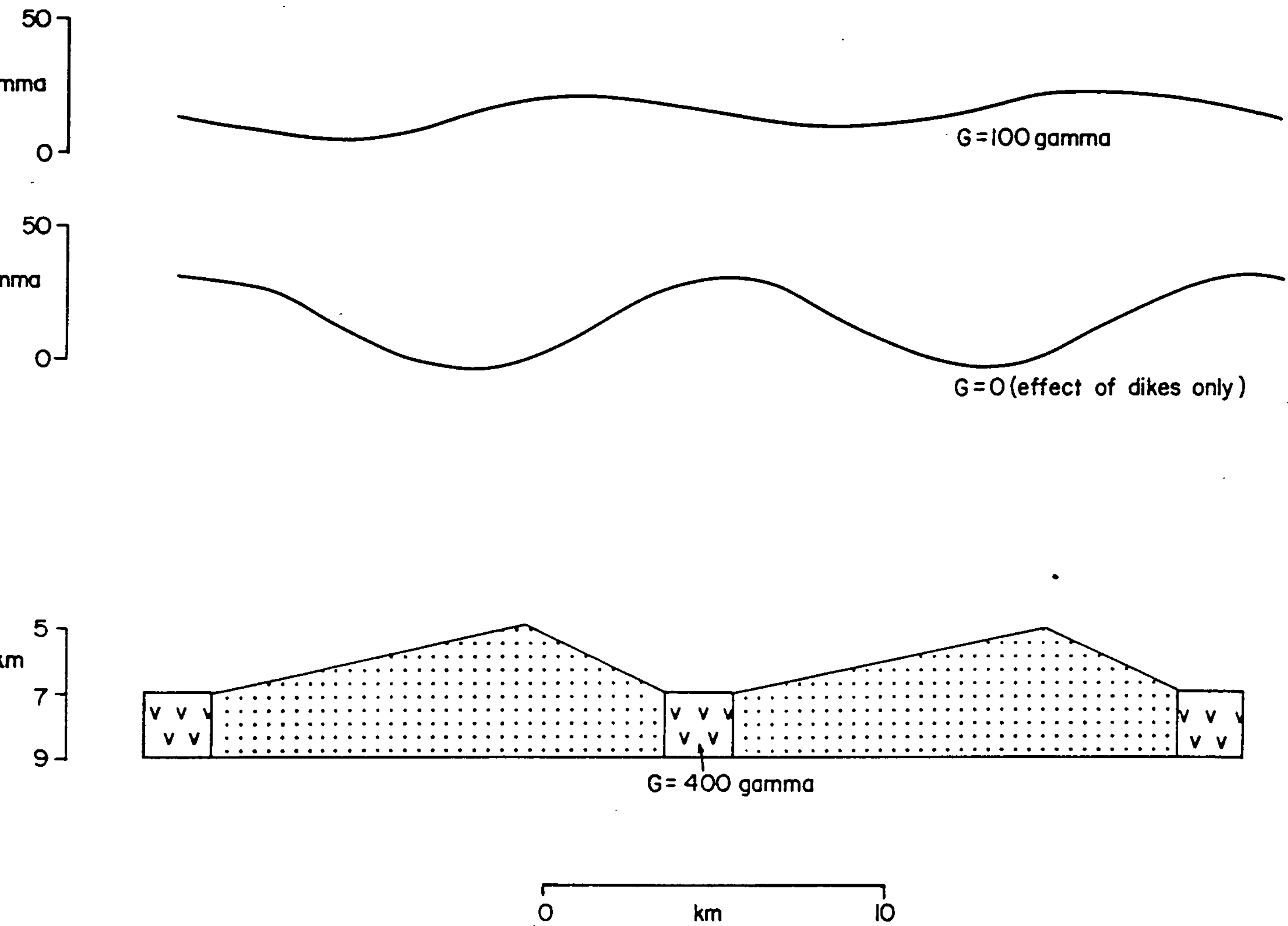

Fig. A5 Magnetic anomalies produced by dikes (G = 400 gamma) in crust of low magnetisation (G = zero gamma, top; G = 100 gamma, below).

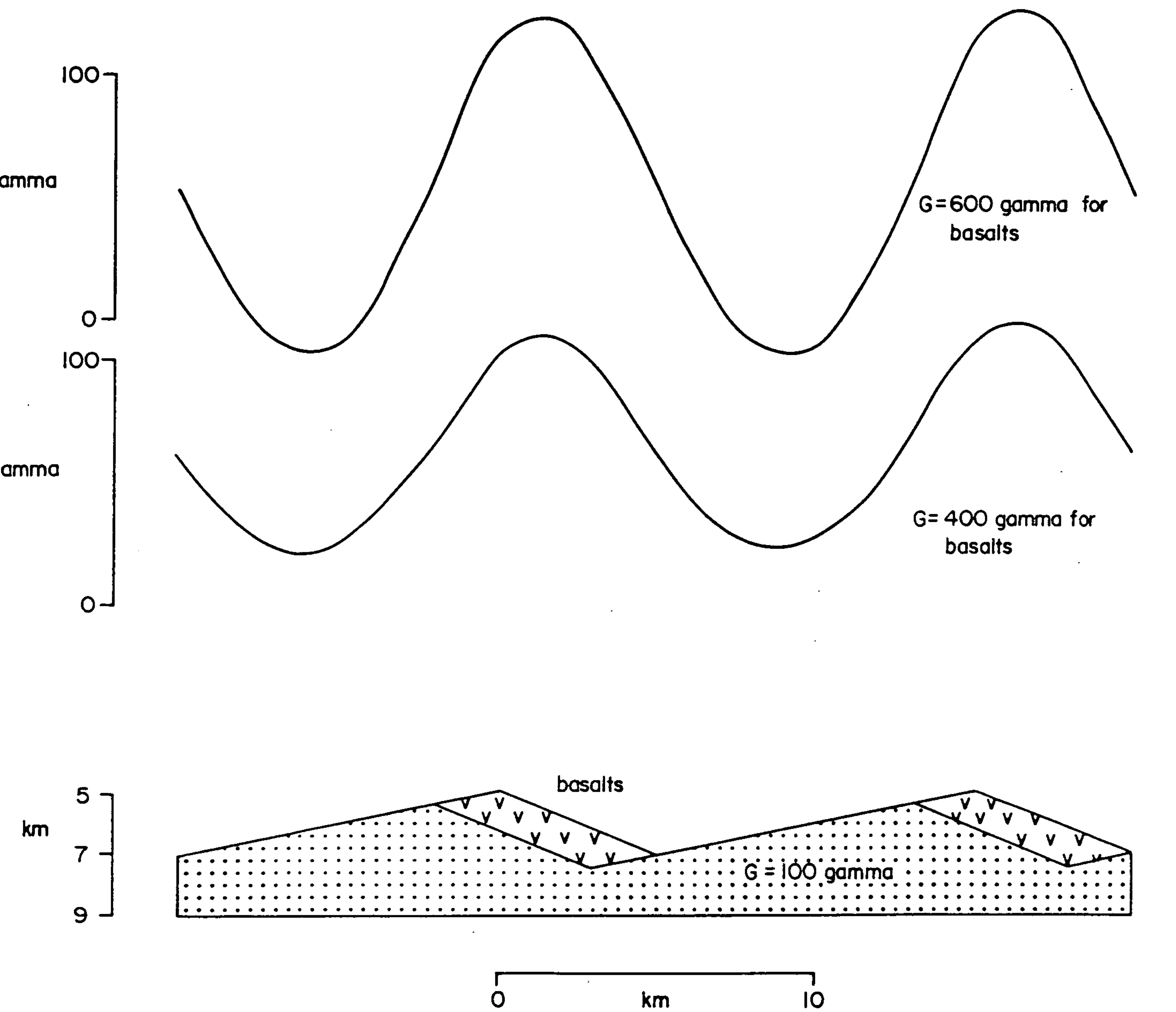

Fig. A6 Magnetic anomalies produced by basalts overlying tilted fault blocks of crust of low magnetisation.

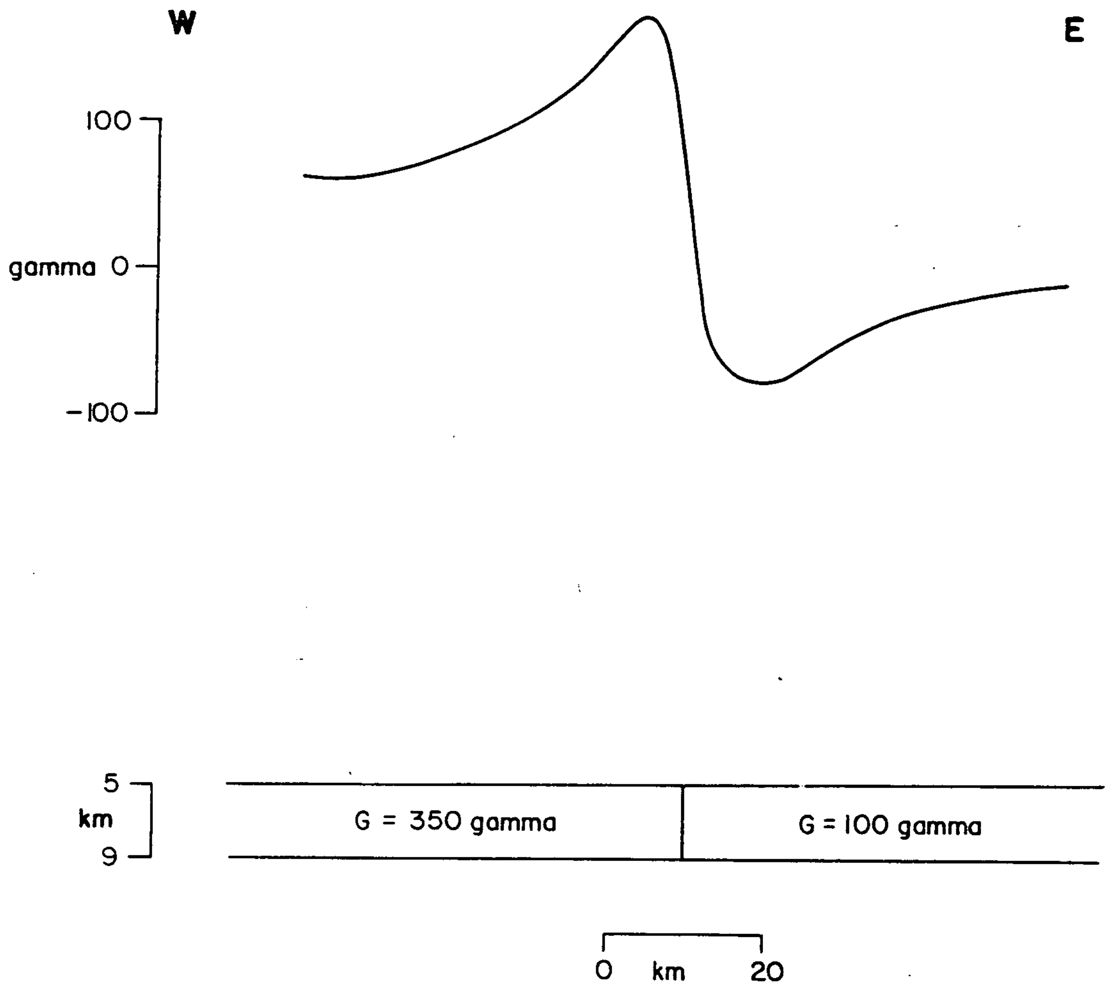

Fig. A7 Magnetic anomaly produced by a hypothetical continent ocean boundary. Oceanic crust is to the west, continental crust to the east.

their magnetisation is assumed to be 400 gamma. Fig. A6 shows anomalies of 100-150 gamma amplitude over a modelled continental crust which has basalts about 1.5 km thick on the tilted blocks.

Magnetic models were calculated for hypothetical continental-ocean boundaries at the same latitude; fig. A7 illustrates one. The modelled anomaly is only slightly larger than those over basalt-covered continental crust (Fig. A6).

Criteria for distinguishing between those anomalies produced by a continent-ocean boundary, and anomalies produced by intrusions within continental crust or ocean floor topography could be:

(i) Shape of anomaly

The continent-ocean boundary anomaly has more of an 'edge-effect' appearance than the other modelled anomalies, which are almost symmetrical. The symmetry or asymmetry of all these anomalies will however be modified by basement topography.

(ii) Continuity of anomaly

Where fracture zones are not closely spaced, it would seem reasonable to expect anomalies due to reversals in oceanic crust and the continent-ocean boundary to be more longitudinally continuous than anomalies due to intrusions in continental crust, at any one margin. However, continent-ocean boundaries can show frequent offsets, and intrusions in continental crust may show considerable continuity (refer to later discussion on the Red Sea).

Continental crust could to some extent be distinguished by its magnetic signature, without necessarily identifying the continent-ocean boundary. If acoustic basement is seen to have considerable topography but a very quiet magnetic field (50 gamma or less), the crust must be of continental origin. However, a somewhat noisier magnetic field (i.e. a few hundred gamma) would not be diagnostic, as it could be a product of topography in oceanic crust, or igneous activity in continental crust.

TABLE A1.

Magnetic anomalies observed over some continent-ocean boundaries at passive margins.

Location	Magnitude of COB anomaly	Type of margin	Comments
East Coast USA	400-700 gamma (East coast magnetic anomaly)	Rifted	There is disagreement over which anomaly represents the COB
"	<100 gamma (Rabinowitz, 1974)		
Red Sea	250 gamma (Cochran, 1981)	Rifted	
Agulhas Plateau	700 gamma (Rabinowitz et al 1976)	Sheared	
Falkland Plateau	400 gamma (Rabinowitz et al, 1976)	Sheared	
Goban Spur	500 gamma (Scrutton, 1979)	Rifted	
Galicia Bank	0 or 50 gamma (Boillot et al, 1980)	Rifted	

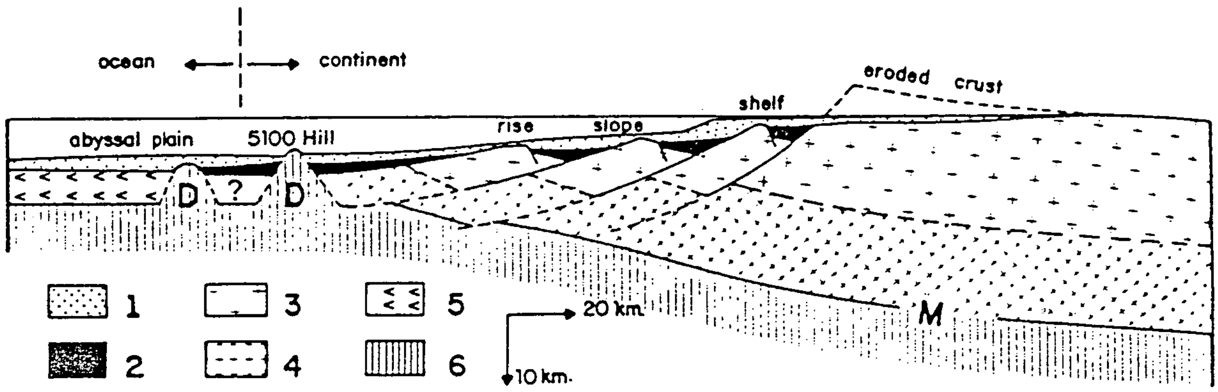

Fig. A8 Model of mechanism of crustal thinning beneath the Iberian margin, after Boillot et al, 1980. Listric faulting extends to the base of the crust.

1. Post-rift sediments
2. Syn-rift sediments
3. Upper continental crust
4. Lower continental crust
5. Oceanic crust
6. Upper mantle

D. Serpentinite diapir

M. Moho

There are magnetic anomalies of 0-50 gamma associated with the serpentinite diapirs.

A very noisy magnetic field (anomalies of 1,000 gamma or more) over deep basement is probably always a result of reversals in highly magnetic (i.e. oceanic) crust, as it is difficult in other cases to have the necessary contrasts in magnetisation. All these methods show the great importance of magnetically modelling quiet zones and continent-ocean boundaries with a knowledge of basement topography.

Table A1 is a compilation of magnetic anomalies observed over various continent-ocean boundaries. The magnitudes vary widely, and as yet there seems no 'representative' or 'typical' magnitude. An atypical continental or oceanic crust may mean that the continent-ocean boundary is totally unrecognisable by magnetic means. Boillot et al (1980) describe a serpentinite diapir at the continent-ocean boundary off Galicia Bank, over which there is no magnetic anomaly (Fig. A8).

GRAVITY

Most modellers assume that there is no systematic difference in density between typical oceanic and continental crust. Indeed most workers assume that the densities of oceanic and continental crust are the same. If there is in fact a distinctive difference between the seismic velocities typical of continental and oceanic crust (refer to earlier sub-section on seismic refraction) then there would also be a distinctive difference in crustal densities. The gravity method however is unable to distinguish between small differences in average crustal densities, and small differences in crustal thickness.

Isostatic gravity anomalies, of the order of 50 mgal magnitude, are commonly observed in the South Atlantic at the inferred position of the continent-ocean boundary. Rabinowitz and La Breque interpret this isostatic anomaly as the result of an uncompensated oceanic basement high at the continent-ocean boundary, and suggest that an isostatic anomaly can be used as a continent-ocean boundary indicator at other margins. The presence of an isostatic anomaly at the continent-ocean boundary however is dependent on the presence of an uncompensated oceanic basement high, which may be a product of early spreading processes peculiar to the South Atlantic. Indeed, Roots et al (1979) working on the West Australian margin, concluded that the initial oceanic crust at rifted margins is of exactly the same thickness as the adjacent continental crust, and is isostatically compensated. Roots et al use different modelling techniques; a depth of compensation of 100 km, raher than the 32 km depth of compensation used by Rabinowitz and La Brecque. Differences of opinion here can also be introduced by the fact of different initial processes at different margins. Simple isostatic anomalies are caused by the change in depth of crust beneath the sediment surface, which may or may not be related to the change from continental to oceanic crust.

Grow (1979) found no systematic isostatic gravity anomaly associated with the East Coast Magnetic Anoamly which is interpreted by most workers as the continent-ocean boundary off the USA.

On the whole, the gravity method alone can be of little use as seismic refraction experiments are much better at detecting unequivocal changes in density and crustal thickness; the gravity method will only be useful in mapping the continent-ocean boundary where a pronounced crustal thickness change (detectable from the FAA or isostatic anomaly) can be proved to be correlated with the continent-ocean boundary - as occurs at sheared margins.

OTHER GEOPHYSICAL METHODS

In the North Atlantic, oceanic crust beneath the Iceland-Faeroe Ridge is 30 km thick, and there has been difficulty in deciding whether the crust beneath the adjacent basalt-covered Faeroe Block is oceanic or continental. In Bott et al 1975, use was made of a converted seismic phase at a crustal contact to provide evidence for a continental origin for the Faeroe Block. As far as is known by the author, there are no other areas where these converted waves have been detected.

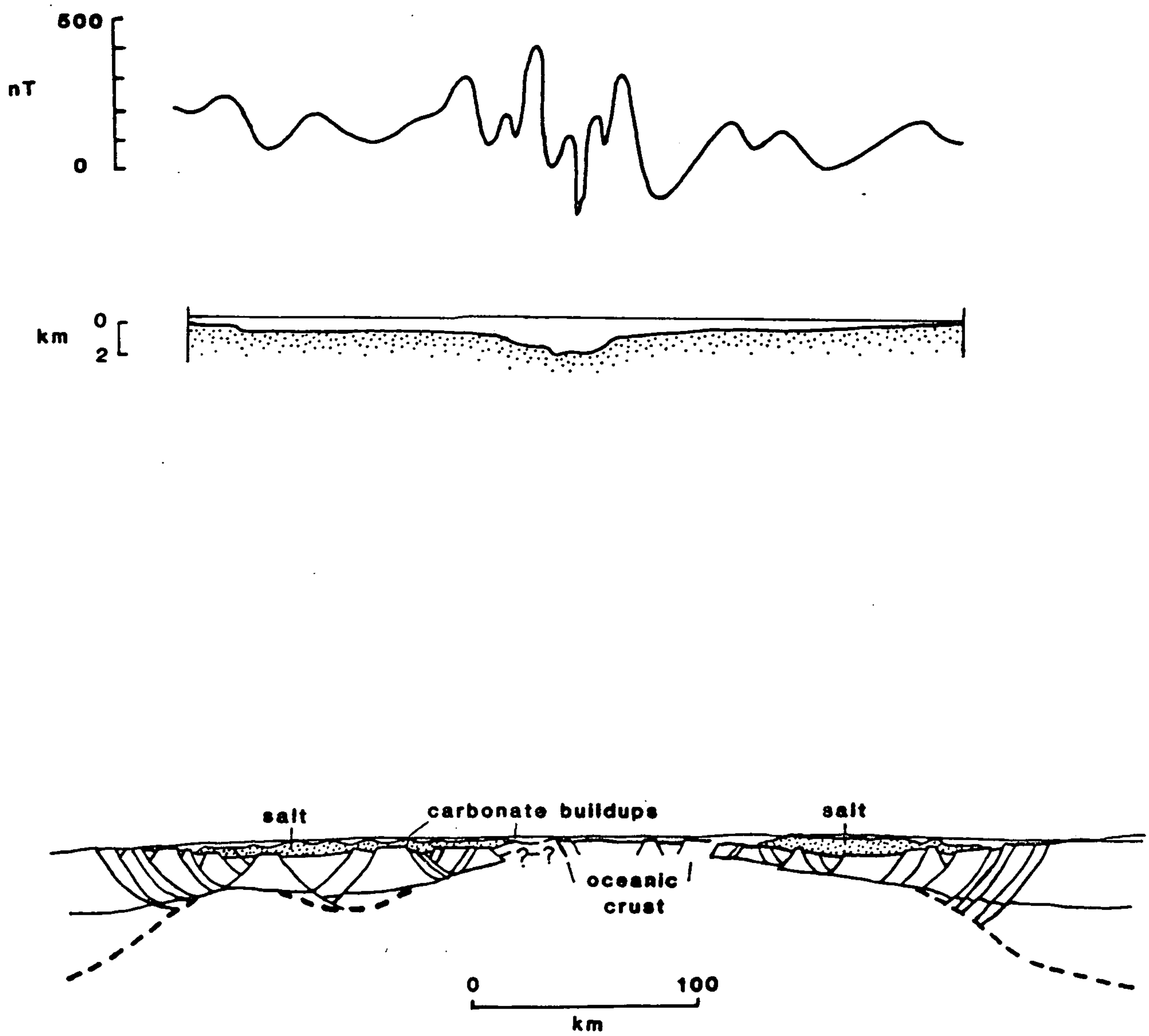

Fig. A9 Above: Magnetic profile and bathymetric profile across the Red Sea, from Girdler and Styles, 1974.

Below: Schematic structural cross-section across the Red Sea, from Lowell and Genik, 1972.

The preceding sub-section discussed the use of individual geophysical methods in detecting the continent-ocean boundary; the following sub-sections will discuss the use made of geophysical methods at continental margins for which there is disagreement on the position of the continent-ocean boundary. This will hopefully illustrate the inherent problems in detecting the continent-ocean boundary, especially in areas obscured by salt or reefs. In many cases personal preferences for a more 'fixist' or 'mobilist' viewpoint determines in part the interpretation of data, and the weight given to different evidence.

THE RED SEA AND GULF OF ADEN

It is generally agreed that the Gulf of Aden and the Red Sea are opening at present by sea-floor spreading, and that Afar is the area into which the spreading centres have, in the last million years, been propogating as lithospheric cracks (Abdallah, 1979; Barberi and Varet, 1977; Black, 1976; Courtillot et al, 1980; and refer to Fig. A1 (Barberi and Varet)).

There are varying views on the amounts of the Red Sea and Gulf of Aden underlain by oceanic crust. Both seas are characterised by a narrow central zone of high-amplitude magnetic anomalies (approximately 500-1,000 gamma), flanked by wider zones of subdued magnetic anomalies (Fig. A9). In the Red Sea these large central anomalies have been correlated with the sea-floor spreading time-scale from the present to 4 or 5 m.y. B.P., in the Eastern Gulf of Aden to 10 m.y. B.P. (Girdler and Styles, 1978; Laughton et al, 1970; Cochran, 1981).

Some workers interpret the flanking quiet zones as oceanic crust (Girdler and Styles, 1974; Girdler and Styles, 1978) and others as continental crust (Cochran, 1981; Lowell and Genik, 1972).

The Red Sea, apart from the axial rift, is largely covered with salt which masks the basement structures, but there is more gravity and refraction data from the Red Sea than the Gulf of Aden. Refraction data from the quiet zones of the Red Sea has been explained in different ways; Drake and Girdler (1964) and Girdler (1969) interpreted their results as indicative of sialic crust (V_p = 5.84 km/s) overlain by evaporites and sediments (V_p = 4.07 km/s), while Davies and Tramontini interpreted their results as oceanic crust (V_p = 6.31 km/s) overlain by evaporites and sediments (V_p = 4.3 km/s sec.). Ross and Schlee (1973) assumed that the wide range of seismic velocities found is more likely to be typical of continental than oceanic crust (Black, 1976). Girdler and Styles (1974) then modelled the 'quiet' zone as ocean crust which evolved from 41-34 m.y. B.P. Some of the evidence was that there must be a much greater separation of the Red Sea than the width of the axial (oceanic) trough, and the other evidence was magnetic modelling of the quiet zone anomalies which are over 200 gamma in magnitude. Girdler and Styles assume that the contrast of large axial anomalies and smaller flanking anomalies suggests two phases of spreading, and also the fact that the magnetic anomalies can be linearly correlated over a distance of 230 km suggests to them that these anomalies must be sea-floor spreading anomalies. However, if these margins were continental, it does not seem impossible that igneous intrusions should be sub-continuous for 200 km. There is no geological evidence for this early spreading in Afar (Black, 1976). Furthermore, the histories of the Red Sea and Gulf of Aden are intimately related, and there is no evidence for 41-34 m.y. old oceanic crust in the Gulf of Aden. In fact Girdler and Styles (1978) interpreted magnetic profiles in the Gulf of Aden as the product of two phases of sea-floor spreading; the first from 30-15 m.y. B.P. and the second from 5 m.y. B.P. to the present.

Cochran's (1981) analysis of the Gulf of Aden and the Red Sea discussed Girdler and Styles (1978) magnetic modelling and points out that the discontinuous spreading model (30-15 m.y. and 5-0 m.y. B.P.) gives a remarkably similar anomaly sequence to the continuous spreading model (10-0 m.y. B.P.). Cochran states that the good fit of the discontinuous model is not due to the section of time scale chosen but to special assumptions made that would imply both an unusual and very complicated history for Sheba Ridge and a great number of differences in the history recorded in two profiles that are only about 40 km apart.

Laughton et al, (1970) also suggested that the quiet zone of the Gulf of Aden was generated by sea-floor spreading, prior to 10 m.y. B.P. Cochran, on the other hand, interprets the quiet zone as continental crust. This assumption is based on the fact that although the boundary between the sea-floor spreading anomalies and the quiet zone contains anomalies of up to 450 gamma, these are not correlatable from one profile to the next. Cochran concludes that 80-160 km of opening occurred in the Gulf of Aden and Red Sea prior to the initiation of sea-floor spreading. Lowell and Genik (1972) also interpreted the flanks of the Red Sea as being underlain by stretched continental crust, and assumed the thinning would be produced by listric faulting; an interpretation similar to the Biscay margin (see Fig. A9).

This illustrates the different interpretations that can be made of the same magnetic field; and presumably that magnetic modelling <u>without</u> seismic reflection information, (especially topography and changes in structure and basement type) can be of little value. If magnetic modelling is not supported by geological observations and other geophysical data the conclusions may prove nothing as almost any chosen model can be made to fit if the right initial assumptions are made.

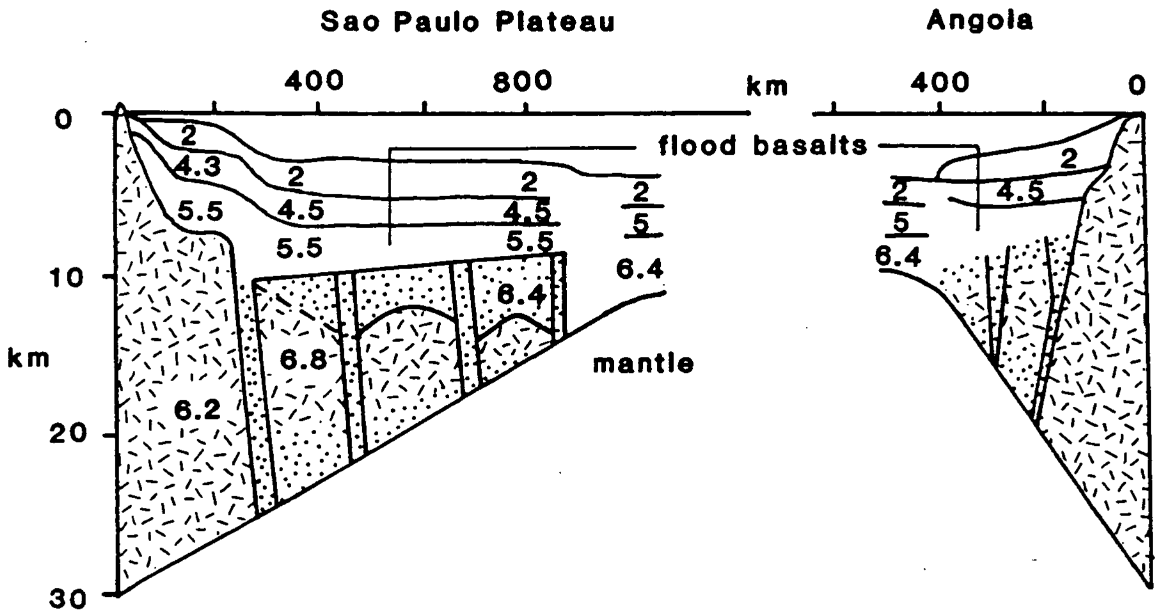

Fig. A10 The structure of the Sao Paulo Plateau and the Angolan continental margin, from Leyden, 1975. Oceanic velocities beneath the Sao Paulo Plateau are explained by massive intrusions of magma through rifted continental crust.

THE SOUTH ATLANTIC

The major salt province in the South Atlantic is the Sao Paulo Plateau (off Brazil), and on its conjugate margin, off Angola. There have been different interpretations as to the nature of the crust beneath the salt; Beck and Lehner (1974) claimed that the basement is pre-Cambrian, which would place the continent-ocean boundary 100-200 km from the shelf edge on the Brazilian side, but no basement has actually been drilled (Leyden, 1976). On the evidence of seismic refraction profiles across the Sao Paulo Plateau (Leyden et al; 1971), the crust was interpreted as oceanic. Interestingly, Leyden himself (1976) later decided that these seismic refraction velocities could also be attributed to continental crust which has been thinned and intruded (see Fig. A10).

The 5.5 km/s velocity would represent extrusive floor basalts interbedded with clastics, and the 6.7 km/s velocity mafic intrusives and metamorphosed continent. There are large magnetic anomalies on the Sao Paulo Plateau, and Leyden points out that there is no evidence to suggest that these are generated by sea-floor spreading - indeed the largest flood basalt deposits in the world are nearby in onshore Brazil.

There may be problems with Leyden's (1976) model, when pre-drift of the continents is attempted (Leyden. op cit.), but it illustrates differing interpretations of the same velocity data.

THE CONTINENT-OCEAN BOUNDARY OFF THE EAST COAST OF THE USA

The most pronounced magnetic features off the eastern USA coast are the east coast magnetic anomaly, a high-amplitude anomaly which is located in places as far seawards as the continental rise, and as far landward as the coastline (Taylor et al; 1968), and a seaward magnetic quiet zone (Heirtzler and Haynes, 1976). Sediment cover is very thick: up to 10 km or more, which makes it difficult to distinguish crustal types by means of seismic reflection methods.

Two different anomalies have been modelled as the continent-ocean boundary edge-effect magnetic anomaly. Keen (1969) modelled the east coast magnetic anomaly as the anomaly resulting from the juxtaposition of a thick magnetic continental crust and a thin magnetic oceanic crust and nonmagnetic mantle. Rabinowitz (1974) on the other hand modelled an anomly within the quiet zone, of amplitude 200 gamma or less, as the continent-ocean boundary. This anomaly ('anomaly E') separates the inner smoother quiet zone, from the outer and noisier quiet zone.

More recent interpretations have usually placed the continent-ocean boundary beneath the East Coast Magnetic Anomaly (e.g. Grow 1979) although basement here is very deeply buried and the transition of basement types not observable by reflection seismic methods.

SIGNIFICANCE OF IDENTIFYING CRUSTAL TYPE

From the economic point of view, the recognition of crustal type will become very important as oil exploration moves into deeper waters.

Oceanic crust has a relatively young sedimentary cover, which is relatively unaffected by faulting at passive continental margins. There may be a potential source, if part of the section is mature. Trapping mechanisms would be dominated by stratigraphic traps in deep sea fans or sandy contourites.

Thinned and subsided continental crust displays a wider variety of possible hydrocarbon sources and trapping structures. Pre and syn-rift sediments could provide good reservoir and source rocks. Pervasive faulting creates tilted fault-block and horst-block traps. The break-up unconformity can provide an important seal as mudstones drape structural highs (cf the North Sea Kimmerian unconformity). At continental margins dyke intrusion and heating due to crustal stretching will shorten maturation times. Source rocks which are immature in mid-continental graben systems may be mature at continental margins where thinning is greater.

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ACKNOWLEDGEMENTS

I gratefully acknowledge the advice and help of my supervisor, Roger Scrutton, and the support of a NERC research grant. I would like to thank the Masters, Officers, and Crews of RRS Shackleton and RRS Challenger, and the staff of NERC Research Vessel Services at Barry for their help and expertise in data collection and processing. Richard Dingle, Richard Haworth, Pam Edwards, Annie Slevin, Tom Furey, and Susannah Betts also provided valuable assistance with data collection. I am grateful for useful discussions with John Waldron, John Ridley, Paul Bennett, David Smythe, and Richard Dingle.

Finally, this thesis would not have been completed without the help of Marjorie Bruce, Richard Hodgkinson, and Colin Walden.

APPENDIX B

ACOUSTIC STRATIGRAPHY OF THE SEDIMENTARY SUCCESSION WEST OF PORCUPINE BANK, NE ATLANTIC OCEAN: A PRELIMINARY ACCOUNT

R.V. Dingle, J.B. Megson, and R.A. Scrutton.

Marine Geology, 47 (1982) 17—35
Elsevier Scientific Publishing Company, Amsterdam — Printed in The Netherlands

ACOUSTIC STRATIGRAPHY OF THE SEDIMENTARY SUCCESSION WEST OF PORCUPINE BANK, NE ATLANTIC OCEAN: A PRELIMINARY ACCOUNT

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(Received May 8, 1981; revised and accepted August 8, 1981)

ABSTRACT

Dingle, R.V., Megson, J.B. and Scrutton, R.A., 1982. Acoustic stratigraphy of the sedimentary succession west of Porcupine Bank, NE Atlantic Ocean: a preliminary account. Mar. Geol., 46:17—35.

Acoustic basement lies at an average of between 6.0 and 6.5 sec two-way time below sea level in the southern Rockall Trough and northern Porcupine Abyssal Plain. The overlying sedimentary succession reaches maximum thicknesses of at least 4.0 sec, and can be divided by 3 regionally-developed seismic reflecting horizons, which are used as a framework to establish an acoustic stratigraphy for the area by selecting three "type" seismic sections. These reflectors are named, in ascending order, Shackleton, Charcot and Challenger. The area is crossed by E—W basement high structures, the Clare Lineament (which *may* be an easterly extension of the Charlie Gibbs Fracture Zone), that separates the Porcupine Abyssal Plain from the eastern part of southern Rockall Trough. Under the latter, the post-Shackleton acoustic sequence is thickened, as if dammed to the north of the Clare Lineament, whilst a further thickening, above reflector Charcot, occurs along a NE—SW line somewhat farther north into the southern Rockall Trough. This can also be related to shallow-lying acoustic basement features. Pre-Shackleton sediments overlie a very irregular basement topography.

The acoustic characters of the various sediment packages are described and it is speculated that major changes in the sedimentary environments took place across reflectors Shackleton and Challenger, the latter probably establishing the modern bottom current circulation patterns. No ages can be unequivocally assigned to the main reflectors, but previously published data suggest a late Eocene—Oligocene age for Challenger. Possible lavas or sills are identified in the succession between reflectors Shackleton and Charcot.

INTRODUCTION

The continental margin of NW Europe, to the west of the British Isles, contains numerous plateau and basin-like features whose tectonic and sedimentary history have given rise to considerable controversy and speculation (e.g. Russell and Smythe, 1978; Roberts et al., 1981). Continuing crustal studies (by R.A. Scrutton and J.B. Megson) to the west of Porcupine Bank and on Goban Spur (Fig.1), have produced large amounts of high-quality single and double-channel seismic reflection data which allow, for the first

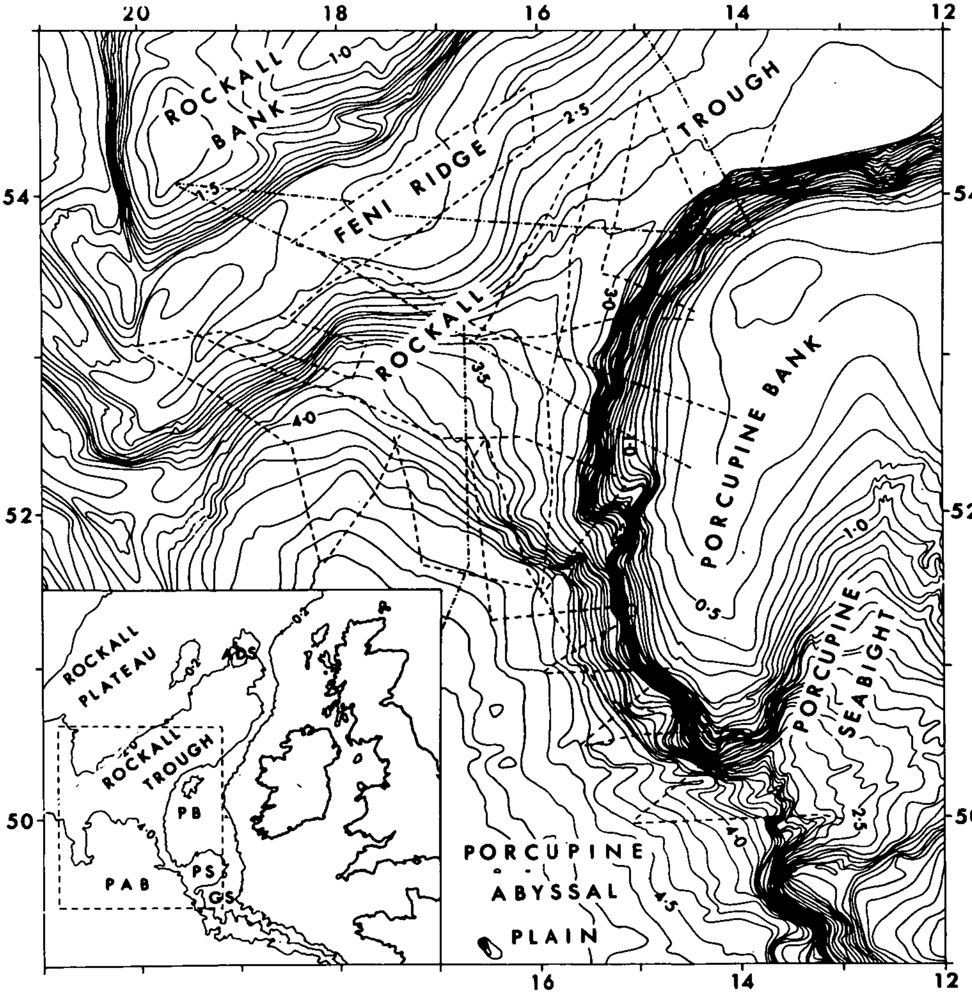


Fig.1. Bathymetric chart of the continental margin west of Ireland (see inset for location), redrawn from Roberts et al. (1979). Isobaths are at 100 m intervals. *ADS* = Anton Dohrn Seamount; *GS* = Goban Spur.
Dashed lines are the tracks of RRS "Shackleton" and RRS "Challenger". Chained lines are GSCOB (1971) and USNOO lines used in this study.

time, a well-controlled acoustic stratigraphy to be established for the southern Rockall Trough—Porcupine Abyssal Plain.

Both Rockall Plateau and Porcupine Bank are shallow-lying continental blocks (Scrutton, 1971), with variable, but generally thin (ca. 1.0 sec two-way time*) sediment cover. In contrast, the southern Rockall Trough and Porcupine Abyssal Plain have thicker sedimentary fills (typically greater than 2.0 sec) which overlie acoustic basement that locally occurs at depths exceeding 6 sec beneath sea level. The junctions between deep-sea basins and

*All reflection times quoted in this paper are two-way times.

adjacent high blocks are topographically and structurally abrupt, with the result that there is generally a major discontinuity in the sedimentary succession across them. In the absence of borehole data, this effectively precludes confident correlations on seismic records between the basins and the banks.

The southern Rockall Trough was included by Roberts (1975) in his survey of the structure and stratigraphy of the area between the mid-Atlantic Ridge (Reykjanes Ridge) and NW Europe. However, its acoustic stratigraphy was accorded only a brief mention because of the relatively few seismic traverses available, and the difficulties in extending the stratigraphy from the better-known areas west of the Rockall Bank. In the latter, Roberts (1975) established a 5-fold sedimentary succession, divided by 4 reflecting horizons which he designated and dated as follows: R4 (Upper Eocene—Oligocene); X (ca. 60 m.y.); Y (ca. 76 m.y.); and Z (ca. 100 m.y.). Recently, a firmer, younger age (52—55 m.y.) has been published for reflector Y (Roberts et al., 1981) which implies that reflector X cannot be ca. 60 m.y. as originally thought, but is also much younger, between ca. 37 and 52 m.y. Inspection of the seismic profiles in Roberts (1975) leads us to believe that the tentative correlation of X, Y and Z into, and within the southern Rockall Trough continues to present difficulties. For this reason, we erect a local acoustic stratigraphy that does not rely upon specific ages and long-range correlations, and which will establish a seismic stratigraphic framework for future detailed studies.

Earlier work by Le Pichon et al. (1970), Jones et al. (1970), GSCOB (1971) and Cherkis et al. (1973) delimited some of the structural elements of the area, but made no consistent attempt to identify and map individual reflectors within the sedimentary sequence. DSDP and IPOD drilling (Laughton et al., 1972; Montadert et al., 1979) on the Rockall Plateau and Armorican margin provide valuable data on regional palaeo-sedimentary environments in the NE Atlantic and have allowed the accurate dating of Roberts' (1975) reflector R4. Shallow-penetration seismic and sediment studies by Flood et al. (1979), Roberts and Kidd (1979) and Lonsdale and Hollister (1979) have established regional late Neogene sedimentation patterns, particularly in the western Rockall Trough area.

The data presented here were collected on cruises RRS "Shackleton" 3/1979 and RRS "Challenger" 6/1980 using 160 and 300 in^3 Bolt airguns and a Geomechanique hydrophone array.

BATHYMETRY

Rockall Trough lies between the Rockall Plateau and the continental margin of western Ireland and Porcupine Bank (Fig.1). It is outlined by the 2 km-isobath, and at its southern end passes into the Porcupine Abyssal Plain where the 4 km-isobath can be conveniently used to delimit the two basinal regions. The southern Rockall Trough is asymmetric, with the 3.0 km-isobath at the foot of the Porcupine Bank in the east, and having depths of only 2.3—2.5 km along its western margin, where the Feni Ridge forms a relatively shallow feature. Sea floor gradients decrease rapidly into the Porcupine Abyssal Plain where depths exceed 4.8 km.

Porcupine Bank is a prominent marginal plateau with a crest depth of less than 200 m. It has a steep and locally precipitous western flank on which there is a major offset and gradient change of about 52° N. This position coincides with a well-developed ESE—WNW lineament in the isobaths of the adjacent deep sea basin.

STRATIGRAPHY

Acoustic basement

In the southern Rockall Trough and the Porcupine Abyssal Plain sediments overlie an irregular acoustic basement. Because a full discussion of the nature and physiography of this basement will be presented elsewhere (Megson, in prep.) we will confine this account to mentioning only the main features that have controlled the shape and size of the receptacle into which sediments have been deposited. Figure 2 shows an outline of the structural framework of the area in terms of basement ridges and the 6.5 sec depth-to-basement (from sea surface) contour. The western edge of Porcupine Bank is clearly delimited by numerous large, mostly westward throwing faults, via which acoustic basement rapidly descends from less than 1.5 sec under the bank to depths below 6 sec (locally >6.5 sec) within the southern Rockall Trough and Porcupine Abyssal Plain. In these deeper areas some clear basement trends can be seen (Figs.2, 3 and 4).

South of 51.7°N, the western edge of Porcupine Bank runs NNW—SSE and this is parallel to a series of large, en echelon ridges in the adjacent deep basin which lie just above and below the 6.5 sec contour. This trend continues southeastwards across the mouth of the Porcupine Seabight and onto the Goban Spur (Dingle and Scrutton, 1977, 1979; Scrutton, 1979). It is abruptly truncated at about 51.7°N by a series of ESE—WNW acoustic basement ridges that locally rise above 6.5 sec. We will call these ridges the Clare Lineament (nearest Irish county is County Clare), and whilst noting that they *may* be related to an easterly extension of the Gibbs Fracture Zone (Olivet et al., 1974; Fig.2), we prefer to leave the question open at the present time. Where the Clare Lineament abuts the Porcupine Bank, the latter has a graben-like indentation (Figs.1 and 2) and changes its trend to approx. N—S. The Clare Lineament separates structurally the Porcupine Abyssal Plain and Rockall Trough. To the north of it, acoustic basement ridges strike approx. N—S or NNE—SSW (i.e., parallel to the Porcupine Bank) into the southern Rockall Trough. In the eastern part of this region, depths to acoustic basement are >6.5 sec, whereas they are <6.5 sec west of about 16.5°W. This asymmetry and the associated acoustic basement discontinuity were noted by Le Pichon et al. (1970) who considered them to mark a major fault line, the Jean Charcot Fault (Figs.2 and 4). At about 53°N the trend of the edge of Porcupine Bank swings round to NE—SW and then at 54°N to

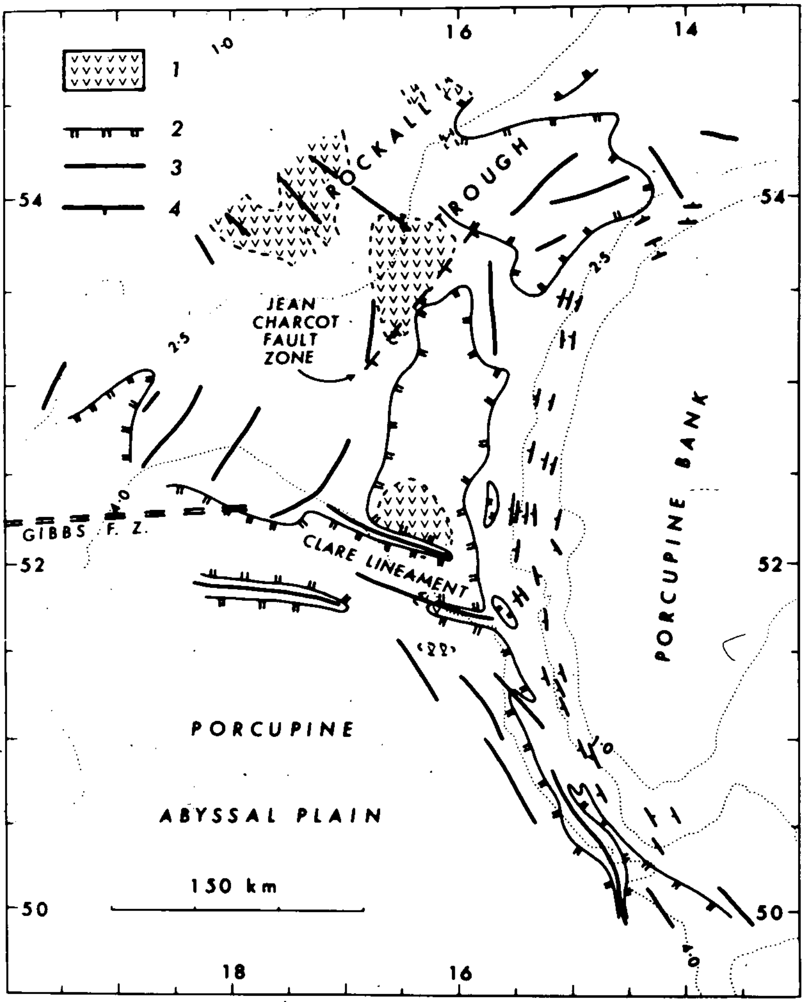


Fig.2. Outline structural framework of the study area. *1* = The known extent of lavas or sills within the sedimentary succession; *2* = 6.5 sec isochron on acoustic basement, ticks on deeper side, *3* = Acoustic basement ridge crests; *4* = Normal faults beneath the steep slopes around Porcupine Bank. The positions of the Gibbs Fracture Zone and Jean Charcot fault zone are after Olivet et al. (1974). The 1.0, 2.5 and 4.0 km isobaths are also shown.

E—W, directions which are followed by acoustic basement ridges in the adjacent Rockall Trough. The last direction is the same as that described by Roberts (1975) as his Porcupine Fault Zone. We do not have sufficient data to corroborate his extension of it west of 16°W.

The northernmost sector of our Fig.2 is separated from the central region by a NW—SE cross-high, which abuts the Porcupine Bank where it alters strike. We suspect that this high extends from a zone of NW—SE acoustic basement ridges in the northwestern part of our area. Roberts et al. (1981) postulate the existence of a fracture zone in this position.

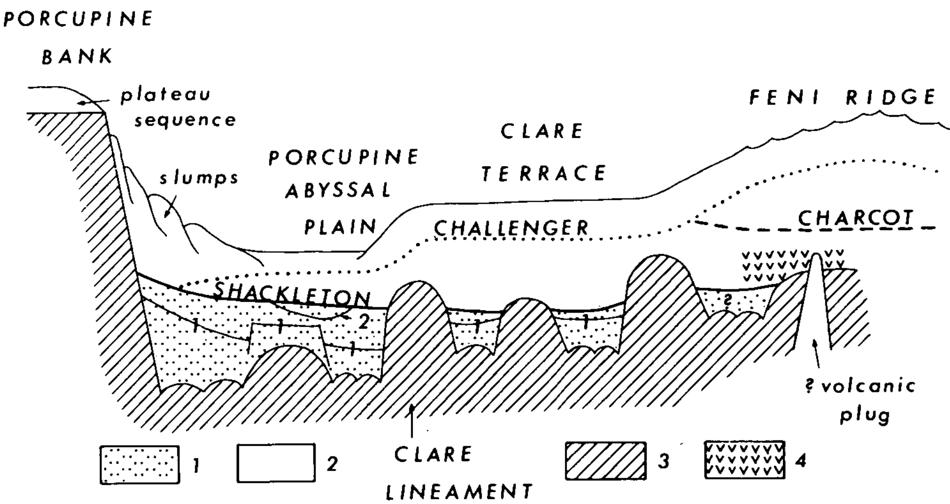


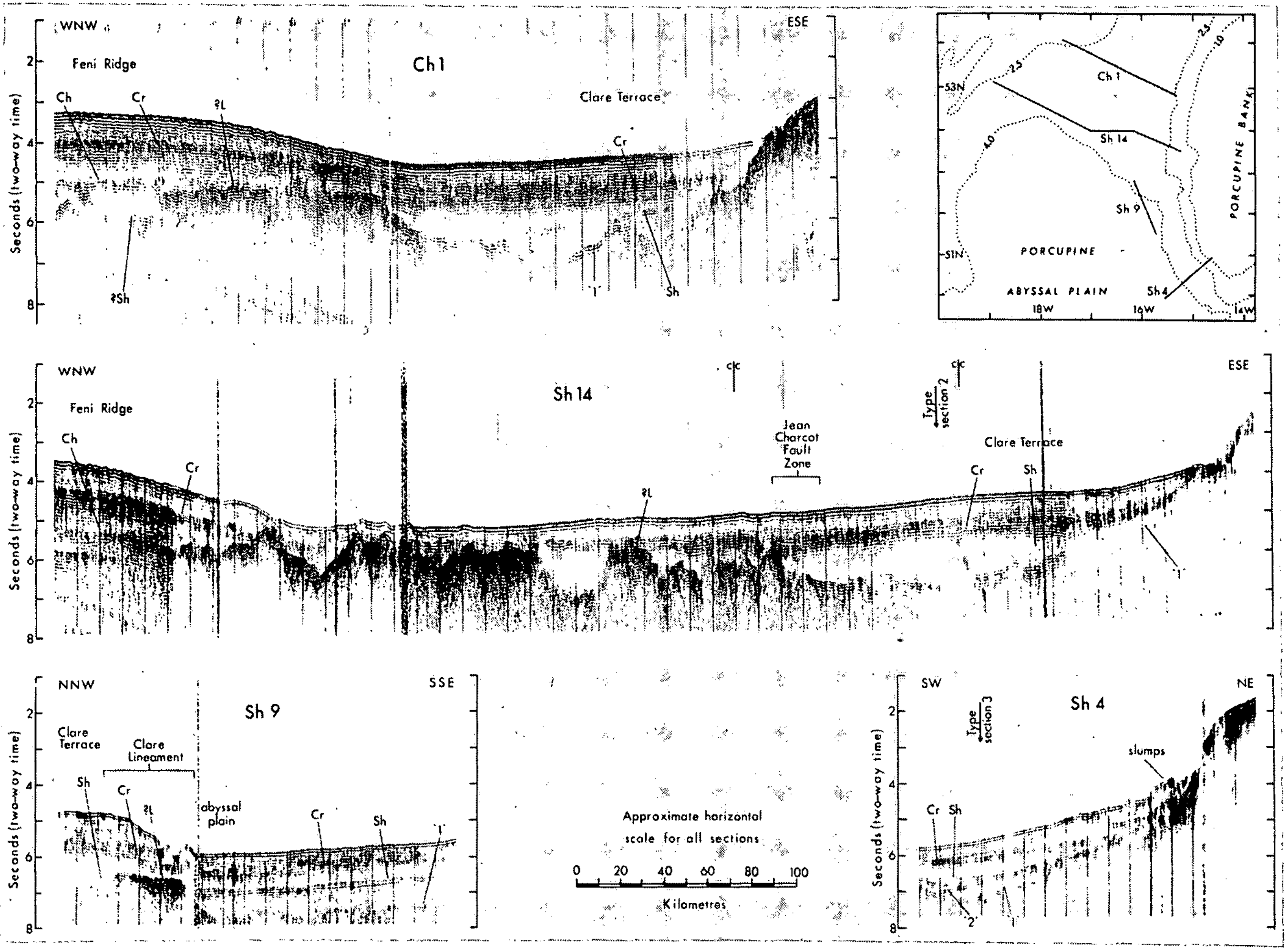
Fig.3. Schematic correlation chart showing the relationships between the seismic stratigraphic subdivisions used in this study. The chart extends westwards from Porcupine Bank, then turns northwards over the Porcupine Abyssal Plain towards the Feni Ridge, keeping east of 17°W. *1* = Pre-Shackleton sequence containing three packages defined by reflectors 1 and 2; *2* = Post-Shackleton sequence; *3* = Acoustic basement; *4* = Lavas(?) or sills. Not to scale.

Finally, it should be noted that in the central region (52.7°—54°N, 16.5°—18°W) acoustic basement may locally coincide with the top of extensive lava sequences or sills that possibly overlie older sediments (Roberts et al., 1981, fig.6).

Sediments

The thickness and internal structure of the sedimentary fill of the southern Rockall Trough and northeastern Porcupine Abyssal Plain vary considerably from north to south, but comparison of seismic profiles (e.g., Figs.3 and 4) shows that these variations are apparently mainly confined to the upper part of the succession, and that there is a regionally-developed reflecting horizon below which such changes are less well-developed. This horizon generally coincides with the upper level of a sediment sequence that smooths out the irregular acoustic basement surfaces, and we will refer to it as Shackleton (after RRS "Shackleton" from which we first mapped it). Such a subdivision allows us to describe the acoustic stratigraphy of the area in terms of pre- and

Fig.4. Photographs of parts of seismic sections RRS "Shackleton" lines 4, 9 and 14 (Sh 4, Sh 9 and Sh 14) and RRS "Challenger" line 1 (Ch. 1). Section locations and vertical and approximate horizontal scales are given. Noteworthy features are labelled. Arrows point to top of reflectors. *Cr* = Challenger; *Ch* = Charcot; *Sh* = Shackleton; *1*, *2* = reflectors in pre-Shackleton sequence; *L* = lavas or sills.



WNW
ESE
Ch1
Feni Ridge
Clare Terrace
Ch
Cr
?L
?Sh
Sh
Seconds (two-way time)
53N
51N
Ch 1
Sh 14
Sh 9
Sh4
PORCUPINE
ABYSSAL PLAIN
PORCUPINE BANK
18W
16W
14W
Sh 14
Jean Charcot Fault Zone
Type section 2
Sh 9
NNW
SSE
Clare Terrace
Clare Lineament
abyssal plain
Approximate horizontal scale for all sections
0 20 40 60 80 100
Kilometres
Sh 4
SW
NE
Type section 3
slumps

post-Shackleton acoustic sequences (Figs.3 and 4). It is in the latter that the major N—S variations referred to above occur, and these can be directly related to sedimentary environments that were established adjacent to the few major acoustic basement structures that remained unburied at the end of the pre-Shackleton sedimentary episode. Although most of these basement structures were subsequently covered, the sedimentary features that can be related to them were perpetuated during post-Shackleton sedimentation and have been inherited in the modern depositional patterns. Figure 3 schematically shows the relationships of the various acoustic stratigraphic subdivisions that we have identified, whilst Fig.4 shows seismic profiles across the northern and southern parts of the area.

Table I sets out "type sections" for the acoustic stratigraphic scheme proposed here — one section for the area south of the Clare Lineament, one

TABLE I

Type sections (depths and thicknesses in sec)

"Challenger" Line 15, 1030 hr, 53° 46.8′N, 15° 34.0′W

Reflector	Depth	Thickness: Package	Thickness: Sequence
sea floor	0		
Challenger	0.6	0.6	
Charcot	1.05	0.45	2.12
Shackleton	2.12	1.07	
basement	?3.0		0.88

"Shackleton" Line 4, 0330 hr, 50° 34.7′N, 15° 23.5′W (see Fig.4)

Reflector	Depth	Thickness: Package	Thickness: Sequence
sea floor	0		
Challenger	0.38	0.38	1.02
Shackleton	1.02	0.64	
pre-Shackleton 2	1.23	0.21	
pre-Shackleton 1	1.47	0.24	0.88
basement	1.90	0.43	

"Shackleton" Line 14, 2030 hr, 52° 29.6′N, 16° 15.0′W (see Fig.4)

Reflector	Depth	Thickness: Package	Thickness: Sequence
sea floor	0		
Challenger	0.75	0.75	1.70
Shackleton	1.70	0.95	
pre-Shackleton 1	2.2	0.50	0.80
basement	2.5	0.30	

section for the area north of the Clare Lineament on the eastern side of the southern Rockall Trough, and one section from the northwestern area under the Feni Ridge. All have been confidently linked by cross-over seismic profiles, and will enable other workers in this area to correlate their results to ours.

Pre-Shackleton sequence

We have recognized Shackleton on all our seismic profiles adjacent to Porcupine Bank, but have been unable to trace it west of about 17°W, either because of lack of data (as in the Porcupine Abyssal Plain), or because it abuts acoustic basement (southern Rockall Trough). Pre-Shackleton sediments fill in a very irregular basement topography, and consequently vary greatly in thickness, from maxima of about 2.5 sec in basement hollows, to nil over highs (Fig.5). Close to Porcupine Bank, Shackleton lies as much as 0.5 sec above the crests of large basement highs, but westwards it steadily descends, so large peaks frequently protrude through it. Because their distribution is closely related to acoustic basement topography, isopachs for these sediments mirror the main basement trends already described from Fig.2.

Shackleton is usually a prominent and continuous seismic event which undulates over buried basement features or drapes, saddle-like between those that it does not cover. Although some of these undulations on it may be related to minor faulting, few if any large faults can be unequivocally identified cutting it. The pre-Shackleton sequence exhibits considerable lateral variation in thickness and seismic character, and can be subdivided into three acoustic packages by two reflecting horizons (1 and 2 on Figs.3 and 4).

The lower package (between acoustic basement and reflector 1) is everywhere represented by a lower, densely reflective zone, locally up to 1 sec thick (but usually between 0.2 and 0.4 sec) that consists of numerous broken and small hyperbolic events. It is draped over acoustic basement and is often highly irregular in attitude. The upper part of the lower package consists of an acoustically more transparent layer with numerous short, apparently well-bedded units, but it is frequently absent over basement highs. It is separated from a middle package by reflector 1 which is a prominent, continuous event up to 0.4 sec thick that smooths out much of the basement topography, but which is frequently cut by large faults across the flanks of structural highs.

The middle package is acoustically similar to the upper part of the lower package, and is locally transparent with numerous broken reflectors. It is unconformably cut out westward either by the upper package, or by Shackleton. The upper package is found only under the Porcupine Abyssal Plain (Figs.4 and 5), where it dips relatively steeply westwards to rest directly on acoustic basement on the western ends of our profiles. Acoustically, the upper package consists of strong, well-spaced, continuous reflectors.

Reflector Shackleton abuts acoustic basement ridges in the Clare Lineament, and is found at approximately the same level on either side of it,

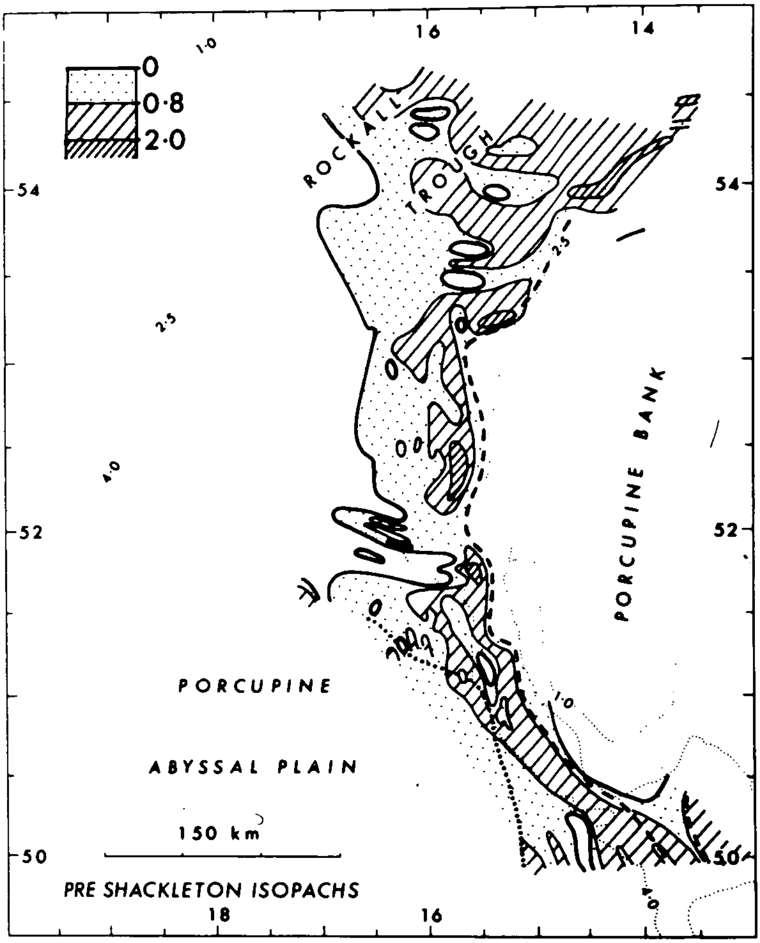


Fig.5. Isopachs of the pre-Shackleton sequence given in sec two-way time. The heavy dashed line is the eastern limit of reflector Shackleton, and the heavy dotted line the eastern limit of the upper package of sediments lying between reflectors 2 and Shackleton (see Fig.3).

suggesting little or no differential crustal movement across this lineament since early in the sedimentary history of the area. At several localities under the southern Rockall Trough, Shackleton abuts reflectors which form a dense, subhorizontal layer with numerous, small hyperbolic echoes. Roberts (1975) and Roberts et al. (1981) have described similar features farther north in the Rockall Trough, and suggest that they are lavas within the sediment column. High magnetic anomaly values over these structures in the southern Rockall Trough recorded during our surveys support this suggestion, but except in rare and equivocal instances, we do not receive recognizable, laterally coherent seismic returns from beneath these layers. In some cases, definite reflectors are seen well below the top of these dense horizons through small windows, and in other areas there is evidence that they are resting on top of the more "normal", mound-like, acoustic basement. At this stage of our researches, we only tentatively identify these events as lavas or sills, which in some areas rest on or near older acoustic basement, and are at the same horizon as or younger than Shackleton.

At its eastern limit, Shackleton rises towards the major basement discontinuity along the edge of the Porcupine Bank, but in no instances are our records clear enough to enable us to trace it onto the bank. It is in this region, immediately west of the basement discontinuity that the thickest pre-Shackleton sequences are found (>2.5 sec).

Post-Shackleton sequence

With the exception of minor basement outcrops on the Clare Lineament, and on the steep western scarp of the Porcupine Bank, the post-Shackleton sequence is found over the whole of the area covered by our investigation. There are, however, important north—south variations in both its thickness and layering characteristics (Figs.3, 4 and 6). These variations can be directly related to underlying basement structures and give rise to three regions that have their own typical macro-topography, sea floor micro-relief, and acoustic stratigraphy: south of the Clare Lineament (Porcupine Abyssal Plain); above and north of the Clare Lineament (Clare Terrace); the whole of the north and northwestern part of Fig.6 (Feni Ridge Complex). Regionally, the post-Shackleton sequence is subdivided by a prominent reflector which we call Challenger (after RRS "Challenger" which was used for most of the survey in the northern area). It occurs over the whole of the area on Fig.6, and allows us to recognize pre-Challenger and post-Challenger acoustic packages. Under the Feni Ridge Complex, an additional acoustic package intervenes beneath Challenger which can be demarcated by a farther strong, regionally-developed reflector that we call Charcot (after the French research vessel "Jean Charcot", which surveyed in this area in 1969) (Fig.3).

To the south of the Clare Lineament, the post-Shackleton sequence reaches maximum thicknesses of between 1 and 1.2 sec in a broad swath 50—100 km wide adjacent to the Porcupine Bank (Fig.6). Westwards, it thins to <1.0 sec. In this area, Challenger is a prominent event composed of discontinuous, medium-strong reflections. The pre-Challenger package is usually acoustically transparent, with only short, weak, broken horizontal reflectors. Within a few kilometers (<10 km) of the foot of the Porcupine Bank scarp it is overstepped by the post-Challenger package, a generally strongly-laminated unit composed of continuous, sub-parallel reflections. Close to the Porcupine Bank, however, the post-Challenger package comprises short broken, often jumbled reflections.

The sea floor rises by between 750 m and 1000 m across the Clare Lineament. This is caused by the rapid thickening of the post-Shackleton sequence above and immediately north of the ESE—WNW basement ridges, producing a flat, or locally convex upwards terrace between about 52°N and 53½°N

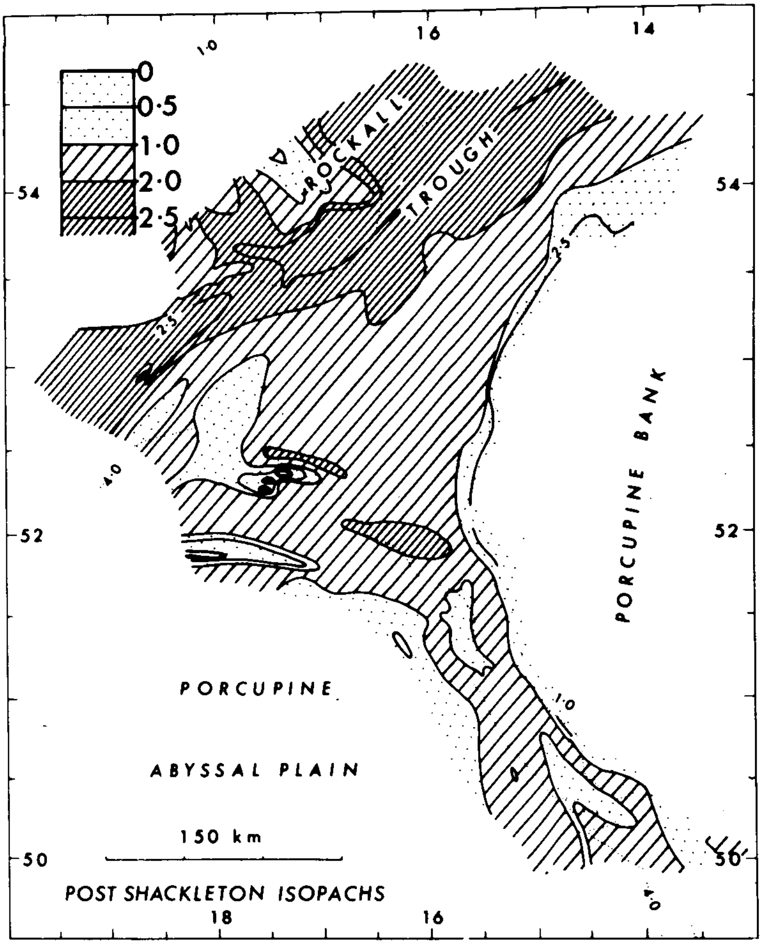


Fig.6. Isopachs of the post-Shackleton sequence given in sec two-way time.

that extends from the base of the Porcupine Bank scarp to about 17½°W. We call this the Clare Terrace (Figs.4 and 7). Beneath it, the post-Shackleton sequence is between 1.0 and 1.8 sec thick, but immediately behind and on top of the Clare Lineament it reaches 2.1 sec. In these areas, both the pre- and post-Challenger packages are thicker than their counterparts under the Porcupine Abyssal Plain, but whereas the latter maintains a relatively uniform thickness (0.8—1.0 sec), the pre-Challenger package varies from 0.2 sec to 1.2 sec. Challenger abuts Shackleton (Fig.4) immediately west of the Porcupine Bank, in a similar fashion to the relationship south of the Clare Lineament, and here also the pre-Challenger package is largely acoustically transparent, in contrast to the densely laminated, post-Challenger package.

Farther west (see Sh 14, Fig.4), acoustic basement rises by about 1.0 sec across the Jean Charcot Fault line, and the pre-Challenger package thins and rests directly on basement. The post-Challenger package thins steadily

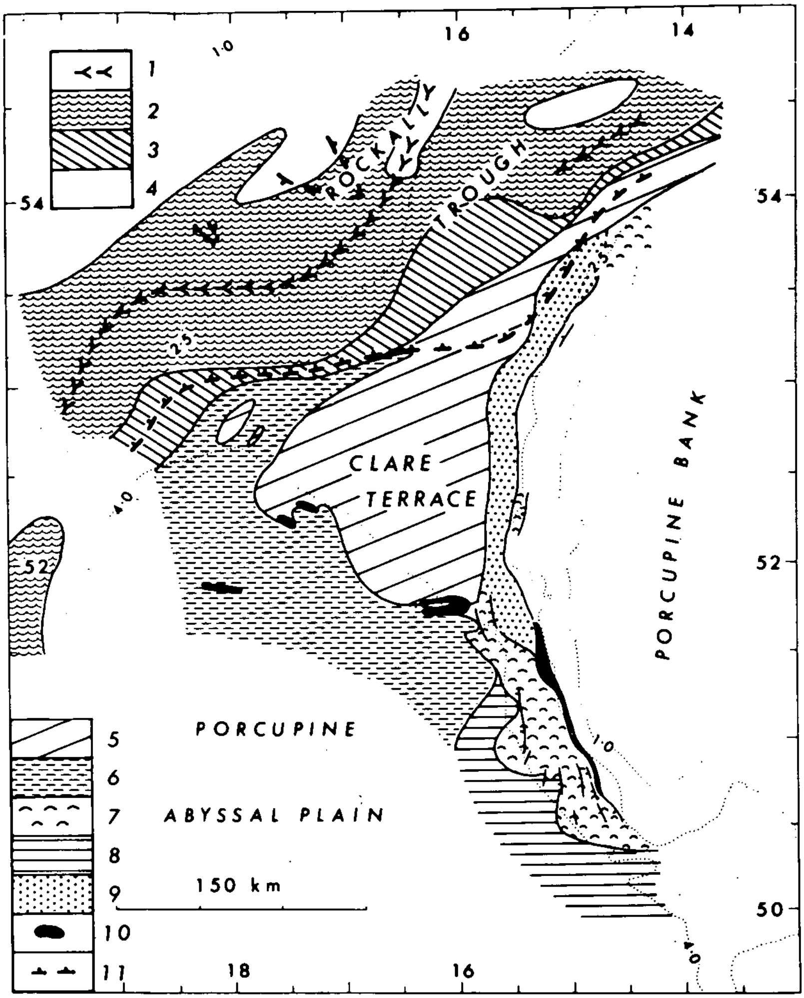


Fig.7. Speculative lithofacies and sedimentary environments of the sequence between reflector Challenger and the sea floor. *1* = Sediment ridge crest (main feature in NW is Feni Ridge); *2* = Rippled sea floor; *3* and *4* = Un-rippled areas over Feni Ridge complex, 3 = Being the steep SE flank; *5* = Convex up, smooth topped Clare Terrace structure, *6* = Horizontal, densely laminated abyssal plain turbidites; *7* = Slumps, with main glide plane scars having ticks on downward movement side; *8* = Partially buried slumps; *9* = Convex up structures, reworked slumps (?); *10* = Acoustic basement outcrops; *11* = Extent of reflector Charcot (present on ticked side of line).

westwards from the zone adjacent to the Porcupine Bank, with the result that the whole post-Shackleton sequence under the Clare Terrace becomes attenuated westwards, accompanied by a progressive increase in water depth. These phenomena ultimately lead to the degeneration of the post-Shackleton Clare Terrace structure into a relatively thin (<1.0 sec) abyssal plain succession that laps around shallow-buried basement peaks in the vicinity of 52½°N 18°W (Figs.4, 6 and 7).

The third major region that can be related to the post-Shackleton sequence occupies the whole of the north western part of our area. It runs approx. NE—SW across Fig.6 and roughly coincides with the 2.0 sec isopach. It too is related to major acoustic basement structures, and on all crossings of our survey lines, lies to the north of large, high-standing ridges that remained bare of sediment at the time Shackleton was formed. The main difference between this area and that under the Clare Terrace is the presence of an additional acoustic package between reflectors Shackleton and Challenger that is underlain by reflector Charcot (Figs.3 and 4). Charcot is essentially horizontal, but above it, Challenger is convex upwards resulting in a plano-convex lens of sediment (Charcot—Challenger package). Where thickest, this is acoustically transparent in its lower part, but it has an upper zone of dense, short, often crenulated reflectors. It reaches a maximum thickness of about 1.2 sec.

The Challenger-sea floor acoustic package drapes over the Charcot—Challenger lens, and typically consists of crenulated, short, medium to weak events, and over most of the area has a strongly crenulate (rippled) upper surface (sea floor) which rises to a low sinuous crest (Feni Ridge Crest) that can be traced across the northern and western parts of the southern Rockall Trough. The south and east facing slope of the Feni Ridge drapes over the feather edge of the Charcot-Challenger package and is non-rippled (Figs.4 and 7).

The zone of maximum thicknesses of the post-Shackleton sequence coincides approximately with the zone of the Feni Ridge crest (up to 3.0 sec) and forms a broad NE—SW swath across Fig.6. It should be emphasized, however, that the bulk of the thickness increase across the eastern edge of the Feni Ridge Complex is contained in the Charcot—Challenger package, and that the post-Challenger sediments do not vary significantly in thickness from those over the Clare Terrace or even parts of the Porcupine Abyssal Plain.

In the extreme northwest of Fig.6, post-Shackleton sediments are relatively thin (<1.0 sec) and reflector Charcot is locally absent where acoustic basement highs occur (Fig.7). Here, several large NW—SE acoustic basement ridges, accompanied by strong magnetic and gravity anomalies have been mapped, and although we do not as yet know their significance (Roberts et al., 1981, interpret them as fracture zones), they clearly were sediment-free features until immediately prior to the local formation of reflector Challenger, which passes over them with only minor upward deflections. Numerous strong, short and locally hyperbolic acoustic events also occur in this area and are particularly associated with these basement highs. By comparison with the features mentioned earlier, we tentatively map these as lavas, which are all pre-Charcot in age (see Fig.2).

Finally, the acoustic stratigraphy along the northern edge of the Porcupine Bank shows only minor differences to that farther south. Here, there is a narrow zone, comparable to that seen under the eastern Clare Terrace, north of which is an expanded post-Shackleton sequence containing a Charcot—Challenger package. The latter forms a thickened Feni Ridge-like structure with a rippled upper surface. The 2.0 and 2.5 sec post-Shackleton isopachs run NE—SW across this northern sector and approach to within 60 km of the edge of the Porcupine Bank.

Total sediments

Figure 8 shows isopachs of total sediment above acoustic basement in the southern Rockall Trough and Porcupine Abyssal Plain. South of the Clare Lineament, maximum thicknesses occur in a relatively narrow zone parallel and close to the western edge of Porcupine Bank (2.0—2.4 sec), but westwards under the abyssal plain, the succession thins to between 1.0 and 1.5 sec.

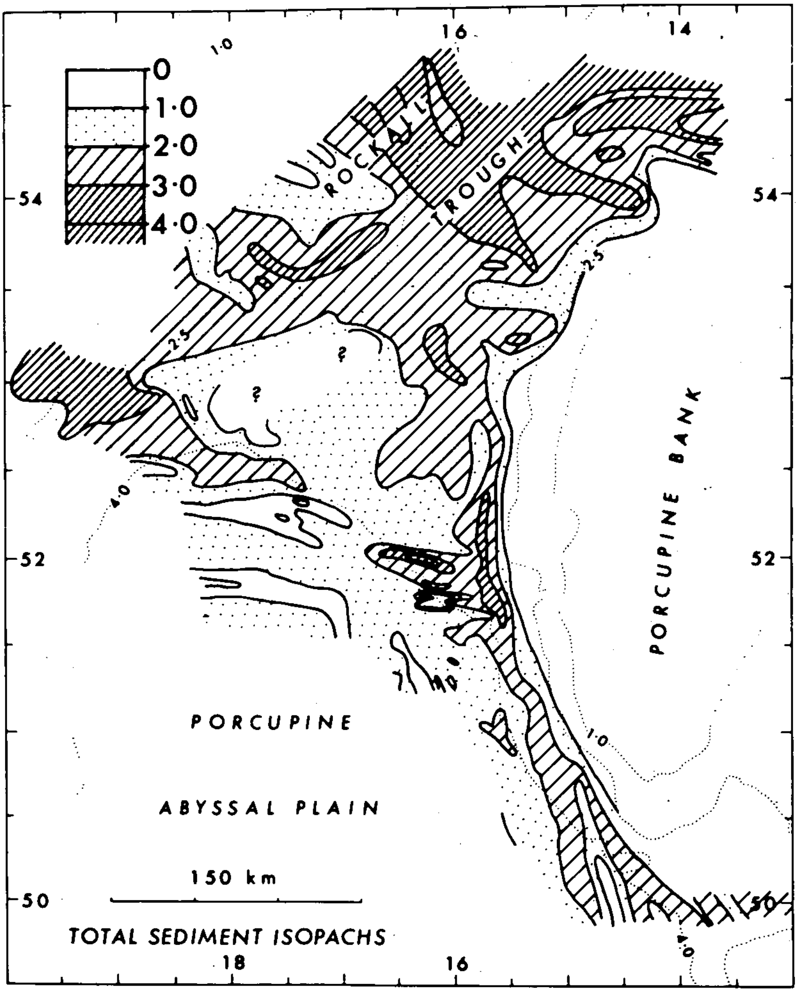


Fig.8. Isopachs of total sediment thickness above acoustic basement given in sec two-way time.

Thin cover over the acoustic basement ridges of the Clare Lineament is clearly brought out, but so too is the small thick elongate pocket (>4.0 sec) immediately to the north of it, which lies adjacent to the thickened wedge at the foot of the Porcupine Bank scarp (maximum 3.4 sec). The Clare Terrace is underlain by a relatively thick succession (>1.5 sec) which has several small elongate accumulations that reach >2.5 sec. Westwards across the Jean Charcot Fault, the succession thins rapidly to <1.5 sec.

Isopachs in the north and west of Fig.8 for the most part show a NW—SE elongation, which is a reflection of basement topography that masks the influence of the NE—SW Charcot—Challenger package isopachs under the Feni Ridge. This is most strongly brought out when a comparison is made between Figs.6 and 8 for the area at the northern end of the Porcupine Bank. Sediments >3.0 sec thick occur over large areas of the north and extreme west of Fig.8, illustrating the disparity in general thicknesses between most of the southern Rockall Trough (>2.5 sec) and the areas immediately west of the Clare Terrace and over and south of the Clare Lineament (<2.5 sec).

DISCUSSION

In the absence of deep drilling, it is not possible to establish either the age or lithofacies of the bulk of the sediments that we have mapped, and it must be stressed that in the following discussion attempts to delineate sedimentary palaeoenvironments are largely speculative.

Firstly, it is not known if the major reflectors are diachronous horizons, or whether they are synchronous. In this connection, any attempt to extend acoustic correlations into the southern Rockall Trough and Porcupine Abyssal Plain from elsewhere must be treated with caution. The only reflector identified by Roberts (1975) in southern Rockall Trough that we can confidently correlate with our sequence is reflector R4, which probably correlates with Challenger. Although there is no substantive proof at this stage that R4 in the southern Rockall Trough is equivalent in age to the Upper Eocene—Oligocene R4 in the vicinity of DSDP and IPOD sites in Biscay and Rockall Plateau, Challenger is probably of about mid-Palaeogene age. Assigning a lithology to reflector Challenger is equally difficult, but there is some evidence that it may be chert beds (Roberts, 1975).

The major change in sedimentary environments occurred across reflector Shackleton. Pre-Shackleton sediments filled in a highly irregular basement topography, and their deposition was accompanied by large-scale faulting. We find no significant faulting younger than that just mentioned, and all the features attributed to faulting of R4 by Roberts (1975), in our opinion, could be drapes over uneven surfaces. After the formation of Shackleton, sedimentation patterns altered and were dominated by the construction of huge features whose geometry and distribution were, at least in Neogene and Quaternary times, controlled by sea floor current paths influenced by topography. At the formation of Shackleton the topography included narrow elongated basement peaks and ridges in the east and south, and large irregular structures west of the Jean Charcot Fault line.

Adjacent to Porcupine Bank, especially immediately north of the Clare Lineament, a lens-shaped body was constructed between Shackleton and Challenger, suggesting sediment dispersion by bottom currents flowing from the north parallel to the flank of Porcupine Bank. Development of the Charcot—Challenger package in the north and west can be viewed in two possible ways. If reflector Challenger is synchronous, then the implication is that the deposition of the Charcot—Challenger package was contemporaneous with the upper Shackleton—Challenger package farther southeast, indicating the establishment of an additional bottom current-controlled sediment dispersion route to the northwest of the original pathway. If on the other hand, Challenger is diachronous, then the Charcot—Challenger package may be younger than the upper Shackleton—Challenger package under the Clare Terrace, and would have been formed because of a northwest shift of the earlier sediment dispersion routes. Either way, it appears that bottom currents were active in this area before the formation of Challenger.

Post-Challenger deposition has been under sea floor current conditions which entailed a significant re-organization of those operating earlier. This change to a circulation pattern dominated by cold Norwegian sea overflow water is well documented (see Roberts, 1975, pp.499—502). Construction of the Feni Ridge over a core of the Charcot—Challenger package belongs to this period, although post-Challenger thicknesses vary little: 0.8 sec under Feni Ridge; 0.7 sec on the central Clare Terrace; 0.5 sec under Porcupine Abyssal Plain; and up to 0.9 sec adjacent to Porcupine Bank. If Challenger is synchronous, these figures indicate that the most recent accumulation rates have been remarkably similar over the whole area. Conversely, they could be used to make a case for Challenger being diachronous.

Modern circulation probably became established with the Challenger event, and Fig.7 shows a first speculative attempt at relating the seismic fabric of the post-Challenger package to lithofacies and sedimentary environments. Large slumped masses are evident against the steep flanks of Porcupine Bank in the south. Westwards these pass into, and are over-lain by ponded (turbiditic?, episodic) beds under the Porcupine Abyssal Plain, which lap round and flank the western Clare Terrace and abut sharply against the southern walls of the Clare Lineament ridges. Uphill directions suggest northerly and easterly sources. North of the Clare Lineament, much of the material adjacent to the Porcupine Bank is not obviously slumped (i.e., no glide planes or suggestive microrelief) and it may represent bottom current-controlled deposition, possibly re-working local slumps.

The Feni Ridge has been investigated in some detail by Roberts and Kidd (1979), Flood et al. (1979), and Lonsdale and Hollister (1979). They have shown it to be a large, low-crested feature with numerous small bed forms on its flanks. Lonsdale and Hollister (1979) have cited photographic evidence for southwest flowing, strong to moderate (ca. 12—15 cm/sec) bottom currents on its northern flank in the northern part of our area, though no evidence has been brought forward to prove that these ripples are actually migrating under the influence of the present-day currents.

Lonsdale and Hollister (1979) also postulate moderate to strong (ca. 12—15 cm/sec) ENE bottom currents along the base of the continental rise (at ca. 2850 m) on the northern edge of the Porcupine Bank (ca. 12.25°W) which are attributed to an "intense but thin (500 m) and narrow (10 km) current" of West European Basin water. The implication (Lonsdale and Hollister, 1979; Fig.7) is that this narrow core sweeps northwards along the western foot of the Porcupine Bank. However, they cite no evidence for this, nor present data on either direction or velocity of bottom currents over the Clare Terrace.

ACKNOWLEDGEMENTS

We gratefully acknowledge equipment loan and seatime funding by the Natural Environment Research Council to one of us (RAS), and thank the officers and crew of RRS "Shackleton" and RRS "Challenger" for their cooperation at sea. JBM gratefully acknowledges a NERC research studentship, and RVD thanks the South African CSIR and University of Cape Town for travel funds. N.Z. Cherkis of USNOO kindly made a copy of a seismic line available to us. D.A.V. Stow read the manuscript and made very helpful suggestions for its improvement.

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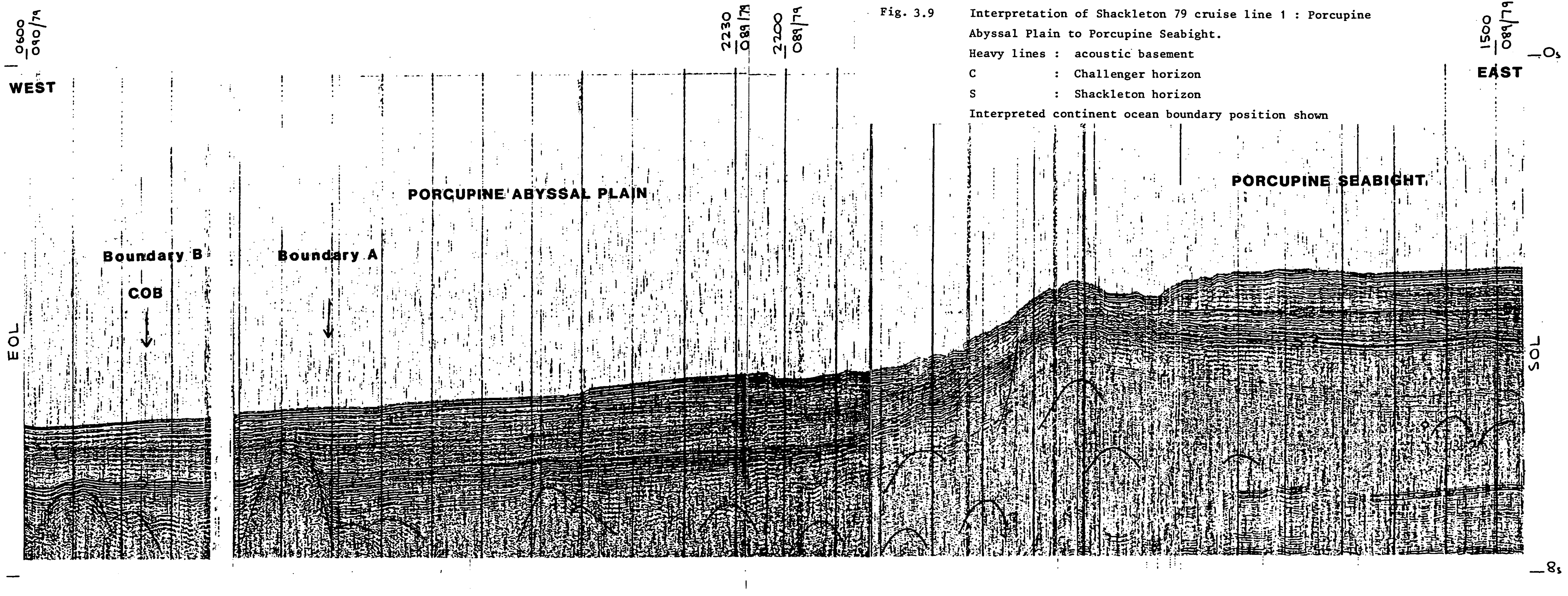

Fig. 3.9 Interpretation of Shackleton 79 cruise line 1 : Porcupine Abyssal Plain to Porcupine Seabight.

Heavy lines : acoustic basement

C : Challenger horizon

S : Shackleton horizon

Interpreted continent ocean boundary position shown

Fig. 4.11 Interpretation of Shackleton line 14 across Rockall Trough

C - Challenger horizon

Ch - Charcot horizon

S - Shackleton horizon

V's indicate position of probable sills

Fig. 4.5 Interpretation of Challenger line 1 across Rockall Trough

C – Challenger horizon

Ch– Charcot horizon

S – Shackleton horizon

J – possible base Cretaceous

V's indicate positions of sills

Chart 2 : FREE AIR ANOMALY CONTOUR CHART

Contour interval 10mGal
Heavy contour 0mGal

Data sources : A – University of Cambridge, B – University of Edinburgh
C – University College of North Wales
D – University of Birmingham, E – CNEXO

Chart 3 : DEPTH TO ACOUSTIC BASEMENT

Contour interval 1.0 sec. two-way travel time plus 6.5 sec contour.
Areas where acoustic basement is known to be deeper than 6.5 sec, shown stippled.
All data except that north of line A-A and south of line B-B (Roberts, 1981) from University of Edinburgh.